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The Structural and Metamorphic History of THE LANGSTRAND—FINFJORD AREA Sørøy, Northern Norway

By

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Abstract

The Langstrand—Finfjord area of N.E. Sørøy is characterised by a metasedimentary sequence of quartzites, pelitic and semi-pelitic schists and limestones, the stratigraphical order of which can be demonstrated by reference to abundant sedimentary structures preserved in massive psammites. Although the basal part of the succession is assigned to the Eocambrian it is probable that, on indirect fossil evidence, the bulk of the sequence is of Cambrian age.

On primary structural considerations it is shown that strata are inverted over a large part of the area, the inversion being attributed to the presence of a macroscopic recumbent fold belonging to the first episode of deformation. At least two major generations of folding have affected the metasediments but the second deformation episode is itself quite complex and divisible into two or possibly three phases. Major folds dated to the first phase of the second deformation episode dominate the tectonic picture and largely account for the present disposition of the strata. Later phases of this same episode include cross-fold type structures and local folding associated with a more brittle deformation and faulting. In this part of Sørøy, mesoscopic F_1 folds are relatively few in number and are consistently of tight or isoclinal style: fold axes are quite variable in trend. The regional schistosity is related to this early fold episode. F_2 folds are abundant, developed on all scales and clearly deform the earlier structures: they vary in style from open to tight or near-isoclinal structures. Over much of the area F_2 axes and lineations trend ca. NNW-SSE but in the south a marked swing round to an ENE-WSW trend is manifest. While the NNW-trending F_2 folds display monoclinic structural symmetry, those of ENE trend show no uni-directional sense of overturning and are characterised by orthorhombic or, quite often, triclinic symmetry. A transition is noted in fold character and symmetry between the two orthogonal strike directions such that this marked swing cannot be ascribed to a post-F_2 deformation. Possible causes of the differing development of F_2 folds are dis-

cussed and comparisons made with the other areas in the Caledonian mountain chain.

Textural studies of the metasediments in relation to the tectonic episodes allow a sub-division of the metamorphic history into a number of phases. The movements giving rise to the early isoclinal folding were probably accompanied by only a low-grade metamorphism. The highest grade of regional metamorphism (sillimanite-almandine-orthoclase subfacies of the almandine-amphibolite facies) was established towards the close of the static interval separating the first and second deformation episodes and extends into the earliest part of F_2. Crystallisation and grain growth continued into this latter episode and can also be shown to post-date the initial kinematic phase of F_2. Diaphthoretic phenomena characterise the later phase of brittle deformation.

Coeval with the peak of the metamorphism a local granitisation of the metasediments was effected which continued, in part, into the early stages of the second deformation. Two principal types of gneiss or granitic gneiss were developed, as well as pegmatites and aplites. Basic rocks, of infrequent occurrence over the area, are present as (1) metagabbroic sheets and (2) amphibolitic sills and dykes. While the latter were intruded after the first folding but prior to the highest grades of metamorphism, the derivation and age of the lenticular sheets of metagabbro are less certain. It is thought probable that they are sheared derivatives of the Storelv Gabbro described by Stumpfl and Sturt (1965) from a wide area S.W. of Langstrand.

Introduction

Sørøy, an island situated off the coast of West Finnmark (Fig. 1) west of the town of Hammerfest (lat. 70°30'N., long. 23°40'E), is composed essentially of a variable sequence of metasediments along with gabbroic, dioritic and alkaline igneous rocks which have been subjected to a complex structural and metamorphic history. The metasediments form an integral part of the West Finnmark Eocambrian succession, the island lying within the province designated as "rocks mainly of Eocambrian age" on the geological map of Norway (Holtedahl and Dons, 1960). This was also extended to include, in part, metamorphosed Cambro-Silurian.

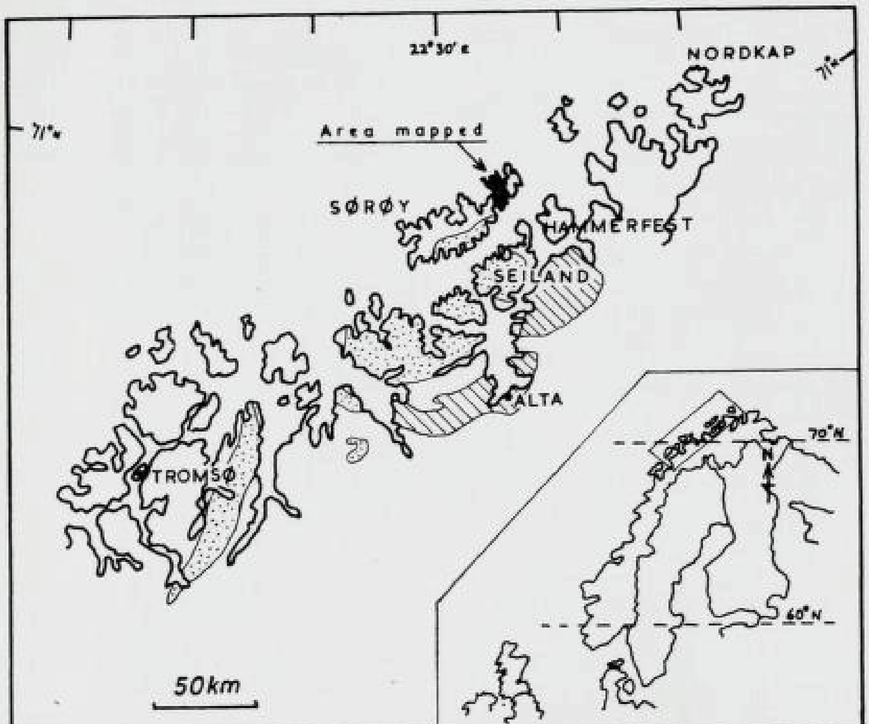

Fig. 1. Location map. Stippled areas – gabbro and ultrabasics; Lined areas – Precambrian windows.

A generalized description of the Caledonides of the Finnmark region has been given by Strand (pp. 270–278) in the "Geology of Norway" (Holtedahl 1960) so that only the salient features of the geology — in so far as they directly concern Sørøy — need be outlined here. The metamorphic complex of north-west Finnmark forms a tectonic unit overthrust upon either autochthonous Cambrian sediments (in the S. W.) or the unmetamorphosed Eocambrian (in the N. E.). Archean basement rocks are found farther to the east beyond the major Caledonian thrust front. This major thrust, while situated in the main some 70–100 km south-east of Sørøy, appears within the metamorphic complex bounding the Pre-cambrian tectonic windows of Komagfjord (Strand 1952b, Reitan 1960, 1963) and Alta–Kvænangen (Holtedahl 1918, Geukens and Moreau 1958, 1960).

A complex basic and ultrabasic petrographic province extends from the Lyngen peninsula in the south up to the islands of Seiland, Stjernøy and Sørøy. The rocks of this province — layered gabbros, diorites, peridotites, pyroxenites, nepheline syenites and carbonatites — have been described in a series of papers (Strand 1952a, Barth 1953, 1961, Krauskopf 1954, Oosterom 1956, 1963, Sturt 1961, Heier 1961, 1964, 1965, Heier and Taylor 1964, Stumpfl and Sturt 1965) while more recently the alkaline rocks of S. W. Sørøy have been dealt with in considerable detail by Sturt and Ramsay (1965).

The metasediments of West Finnmark are relatively poorly documented, detailed investigations being confined to the works of Holtedahl (1918) and Holmsen et al. (1957). A preliminary survey of the Loppen–Kvænangen area, south of Sørøy, has been described by Ball et al. (1963). Prior to the last decade the only literature published on the geology of Sørøy was a series of accounts written by Karl Petersen (1868, 1883). More recently a study of fold styles in the Sandøfjord area by Ramsay and Sturt (1963) provides the only detailed reference to the metasediments and tectonics of the island. A short account of the geology of the Dønnesfjord area has been given by Appleyard (1965).

The present paper is a condensation of the results of a structural and petrological examination of metasediments and gneisses occurring in the Langstrand–Finfjord area of north-east Sørøy, the mapping for which was carried out in the three field seasons 1961–1963. The work also formed the subject matter of a Ph. D. thesis presented at the University of London (Roberts 1965).

I. Stratigraphy

Throughout the area, a conformable metasedimentary succession of quartzites, pelitic and semi-pelitic schists, calc-silicate schists and re-crystallised limestones is clearly recognisable, the stratigraphical order of which can be demonstrated by reference to the excellently preserved primary structures within a thick quartzitic sequence. Although no fossils have been found in the Langstrand area, limestones containing archaeocyathids (Fig. 2) were discovered by Dr. B. A. Sturt during a reconnaissance traverse over part of south-west Sørøy. A preliminary study of these organisms by Professor C. H. Holland of Trinity College, Dublin, points to a Lower Cambrian (or younger) age for a large part of the sequence. The Langstrand–Finfjord succession, along with approximate thickness, is as follows:

Top	Hellefjord Schist		(?)
	Falkenes Limestones Group		60 m
	Storelv Schist		110 m
	Transitional Group		0–25 m
	Klubben Quartzite Group	Quartzite 3	(?) ca. 540 m
		Upper Semi-pelite	0–75 m
		Quartzite 2	280 m
		Lower Semi-pelite	0–280 m
Bottom		Quartzite 1	ca. 1120m

This succession and those established in other parts of Sørøy are broadly analogous, as can be seen from Table I. A composite sequence from the Loppa–Andsnes area (Ball et al., 1963) 50 km south-west of Sørøy is included along with the Sørøy successions in view of its striking similarity to the latter.

Lateral variations of lithology due to original sedimentary facies changes are not uncommon over the Langstrand area. Intercalative pelites within the Klubben Quartzite Group are quite variable in occurrence and gradational boundaries of quartzite and semi-pelite are ubiquitous. Similarly, facies variations can be traced throughout the Falkenes Limestone Group; these are distinguishable from apparent lithological variations caused by tectonic slides.

Fig. 2. Archaeocyathids. Limestone, 5 km E. of Breivikbotn. (Publ. by permission of Dr. B. A. Sturt and Professor C. H. Holland; photo, B. A. S.)

1. *The Klubben Quartzite Group.*

Three psammite units separated by two semi-pelitic horizons constitute the Klubben Quartzite Group, each of these units exhibiting variably transitional boundaries and changes of lithology along the strike. The possibility of these members representing tectonic repetition can be refuted by reference to fairly abundant, though localised, current bedding in the quartzites.

Preservation of sedimentary structures of such abundance and magnitude (Figs. 3, 4 and 5) in areas of high metamorphic grade is not a common occurrence in the Eocambrian rocks of Northern Norway. In the Skarvfjordhamn area, perfectly preserved current bedding is present in individual quartzite beds of 32–35 cm thickness. Although developed throughout the Klubben Quartzite Group, current bedding tends to abound locally within the psammites particularly where pelitic material is scarce and the quartzites fairly massive: invariably the current-bedded units are parallel or tabular (Shrock, 1948). Further evidence of depositional environment is seen at Hønseby where scour-and-fill structures are recognisable (Fig. 6) and in one case ripple-marks have been noted on a bedding surface. Inconstancy of current direction

Fig. 3. Current bedding in psammite. Quartzite 2, lake 135, N.E. of Langstrand.

Fig. 4. Inverted current bedding in psammite. Quartzite 1, S.W. Hønsebyfjord.

Fig. 5. Inverted current bedding in psammite. Quartzite 1, south coast of Hønsebyfjord.

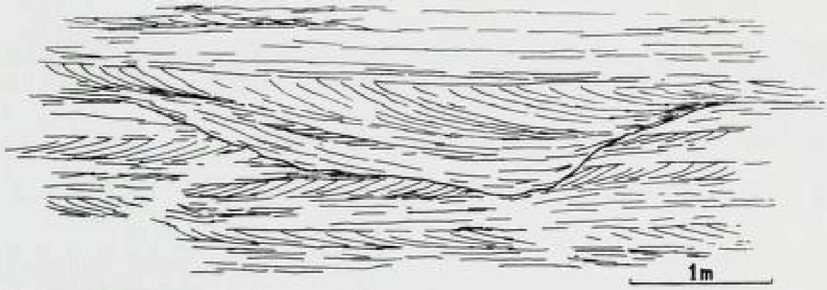

Fig. 6. Scour-and-fill structure, Klubben Quartzite: wave-washed shore section, N. Hønsebyfjord.

is indicated by the arrangement of foreset laminae in opposite directions, often in contiguous beds, producing a 'herring-bone cross lamination' (Fig. 7).

In terms of lithology the psammitic members of the Klubben Quartzite Group exhibit varying characters from massive, extremely well-bedded, white to buff granulitic quartzites to flaggy, dark grey, mica-

Table 1. *Comparative metasedimentary successions of Sørøy (with Loppa–Andsnes area — Ball et al. 1963).*

Langstrand-Finfjord	S. Sandøfjord[1]	S. W. Sørøy[2]	Loppa-Andsnes[3]
Hellefjord Schist		Hellefjord Schist	Pelitic schists
Falkenes Limestone Group	Falkenes Limestone Group	Breivik Group Falkenes Limestone Group	Calcareous series with meta limestone(s) Pelitic schists
Storelv Schist Transitional Group	Pelite 3 Psammite 2 Pelite 2 Psammite 1 Pelite 1 Storelv Formation	Storelv Group	Impure quartzite
Klubben Quartzite Group: Quartzite 3 Upper Semi-pelite Quartzite 2 Lower Semi-pelite Quartzite 1	Klubben Quartzite	Klubben Quartzite	Quartzite and amphibolite Flaggy quartzite Massive and/or folded quartzite

[1]) From Ramsay and Sturt 1963. [2]) From Sturt and Ramsay 1965. [3]) Composition and correlation the responsibility of the present writer.

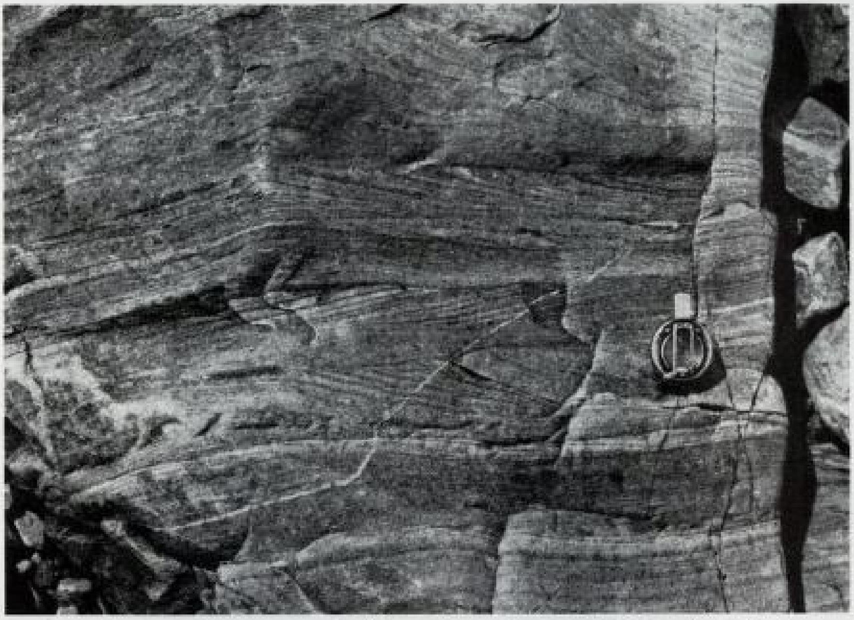

Fig. 7. 'Herring-bone' current bedding. Quartzite 1, South Hønsebyfjord shore.

ceous psammites. Closer examination of hand-specimens reveals a sugary texture in the more massive bands, with pink or buff weathering surfaces usually indicative of appreciable feldspar content. Feldspar is clearly distinguishable in many hand-specimens, and thin-section examination reveals that many horizons within the quartzites are distinctly arkosic in composition.

The darker, banded psammites contain biotite as an essential mineral, and evince a penetrative schistosity paralleling the banding. Specimens of second generation, F_2* fold closures, however, often display a well developed axial plane schistosity. Garnet is frequently developed in pelite ribs and quartzite alike, although it is occasionally absent in the less impure quartzites. It is usually conspicuous in the thin pelite intercalations. In the Hønseby area, in psammite specimens taken for petrofabric analysis, strings of skeletal garnets are clearly aligned parallel to the F_2 axial plane schistosity.

In the Skarvfjord–Hønsebynes area, extreme tectonic stretching has

* Early folds are designated F_1, second period folds F_2, third generation F_3. Schistosities S_2 and S_3 relate to the F_1 and F_2 episodes respectively, S_1 being the original bedding.

often accentuated the stratification, the quartzite now having the appearance of a very closely banded lithology individual units varying appreciably in thickness. Tectonic lenses or segregations of garnet-biotite rock are fairly common, garnets occurring up to 8 cm in diameter, while boudined pegmatite and quartz-kyanite segregations are also notable.

The two semi-pelitic members of the Klubben Quartzite Group are characterized by a fairly rapid alternation of pelitic and psammitic bands, often of microscopic dimensions, containing garnet as an essential mineral. The older, thicker semi-pelite attains a maximum thickness of ca. 280 m north of Veirbukten, although the wide outcrop in that area (see Plate I) is in part a consequence of shallower dips. Boundaries between the semi-pelite and psammite members of the group are often imperceptibly gradational moreso towards the bases of the semi-pelitic horizons. Within the upper semi-pelite, a fairly constant 3-4 m quartzite band contains sporadic current bedding.

Tracing the lower semi-pelite N.N.E. towards Hønsebydalen, quartzitic bands progressively interrupt the sequence until, on the north side of the Hønseby river, the lithology is predominantly a finely laminated psammite with thin, irregular pelite partings. Continuing north of Hønsebydalen, granite gneiss becomes prominent in the succession in the form of concordant partings and irregular lenticular bodies, some of which exceed 1,200 m in length.

Amphibolite sills and dykes cut the metasediments of the Klubben Quartzite Group (Figs. 8, 30 and 42). These vary from 15 cm to 3.5 m thickness and are notably schistose, the schistosity clearly axial planar to F_2 structures.

Within the quartzitic member of the Klubben Quartzite Group, the tendency is for a general increase in pelite content towards the top of the succession. Furthermore, sedimentary structures are much less common in Quartzite 3 as compared with their local profusion in the oldest of the quartzite members. Considering the group as a whole, their lithological development in the Langstrand area and in other parts of Sørøy is quite reminiscent of the uppermost part of the Scandinavian Sparagmite Group (Holtedahl 1960, 1961).

2. *The Transitional Group.*

From Langstrand to upper Hønsebydalen there occurs a particularly flaggy and distinctive lithology, essentially a transitional series between

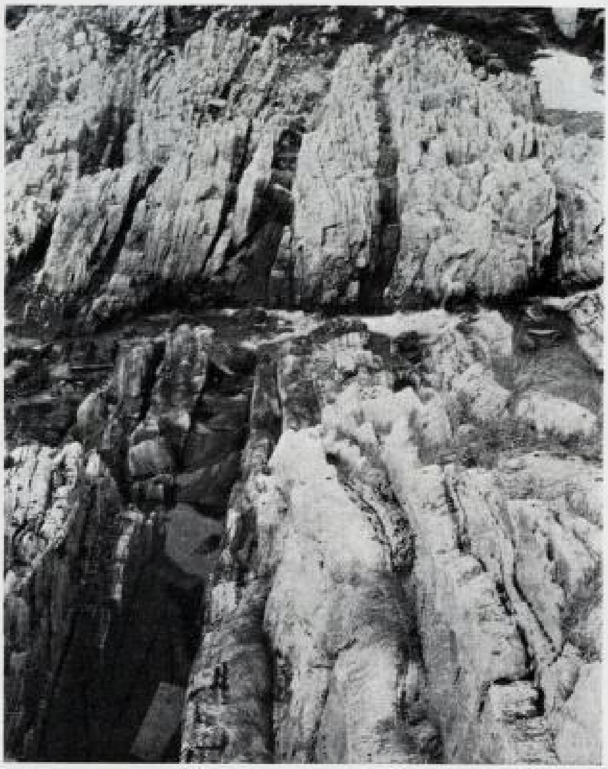

Fig. 8. Transgressive amphibolite sills in Klubben Quartzite; coast south of Hønsebynes.

the Klubben Quartzite Group and the richly garnetiferous Storelv Schist. It is composed of alternating thin, white quartzite bands and pelitic laminae, the quartzite bands averaging 2-3 cm thickness, with boudinage a common feature (Figs. 40 and 46). Apart from quartz, plagioclase and muscovite are the only significant minerals in this qaurtzite: in the pelitic bands, muscovite is in excess of biotite while garnet is only occasionally present.

North of Hønsebydalen this lithology is less readily identifiable, becoming more pelitic and schistose along the strike. Although it is present again in the extreme N.W. of the area, quartzitic ribs are there relatively subordinate. This affords further evidence of lateral facies variations during sedimentation.

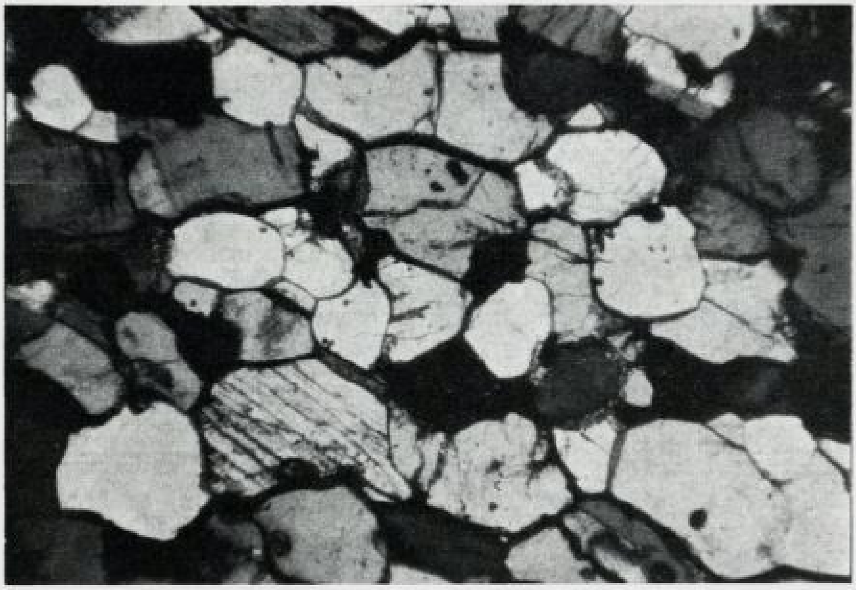

Fig. 9. Sub-rounded clastic quartz grains in friable ferruginous 'quartzite'. From Storelv Schist NNW of Rødbergodden. Crossed polarised light, x60.

3. *The Storelv Schist.*

This is a rusty-weathering pelitic schist which crops out continuously from Langstrand to Finfjord, and also in the north-west of the area west of the Skarvfjord Synform. Two lithologies are distinguishable, the dominant one a garnet-muscovite schist, the other a muscovite-biotite schist. The latter occurs stratigraphically above the garnetiferous schist, the boundary between them usually being gradational.

The garnet-muscovite schist develops a brownish rusty-weathering surface. Muscovite, garnet and quartz constitute the bulk of the rock, with biotite present in only small amounts. Garnets are generally profuse occurring as small deep red or purple-red porphyroblasts rarely exceeding 2 mm diameter. Kyanite is common in the schists in the north-west of the area and has been observed in the Skarvdal-col region. Staurolite has been noted in four thin-sections from widely scattered localities while sillimanite is present in one specimen (LR 2) from Langstrand.

Between lake 170 and Finfjord, conspicuous pods or lenticles of pegmatitic material are present in this schist. Pods rarely exceed 10 cm in length and have diffuse contacts, grain size often grading into that of the schist. Tourmaline needles (up to 1.5 cm in length) are common and frequently extend into the pelite. Along other horizons, quartzo-feldspathic material seems to have been more pervasive producing a gneissic texture in the schists.

The quartz-muscovite-biotite schist is a compact, grey-weathering lithology sometimes containing thin, impersistent garnetiferous horizons. As well as the previously mentioned gradation into the main garnetiferous schist, a lithological transition into overlying calc-silicate schists may be observed.

Locally, within the mica schist, the lithology takes on a more psammitic aspect and north of Rødbergodden (Skarvfjordhamn) a conspicuous 2–3 m band of yellow-brown weathering, friable psammite is present. This psammite, of gritty appearance, contains biotite and sporadic small garnets, and thin-section and binocular microscope examination reveals well-preserved sedimentary grains coated with a film of iron-oxides (Fig. 9).

Feldspathisation features are present in the mica schist near Lundvatn, the lithology being quite gneissose with many quartz and quartzo-feldspathic segregations.

4. *The Falkenes Limestone Group.*

This group embraces a variety of rock types ranging from graphitic kyanite phyllites to calc-siliceous schists and limestones. Many lithological variations occur although the two main limestones can be traced over most of the area (Fig. 10) forming a prominent minor scarp between Hønsebyvatn and the fault-controlled Skarvdal-col. A downthrow of an estimated 60–80 m on the Skarvdal fault displaces the group to near sea-level in W. Finfjord (Plate I). Outside the limits of the area mapped by the present writer a siliceous limestone occurring on the small island Lille Kamøy, 4 km N. of Saksnæringen headland, probably represents the continuation of the Falkenes Limestone Group (Dr. B. A. Sturt, personal communication).

Epidote is a common mineral in the pale, grey-green, basal calc-silicate schists which pass up into tremolite schists and phyllites with frequent diopsidic bands. Tremolite is often present in well-developed 'knotenschiefer'.

The two main limestones are separated by 20–25 m of variable schists and phyllites. The Lower Limestone (maximum thickness 11 metres)

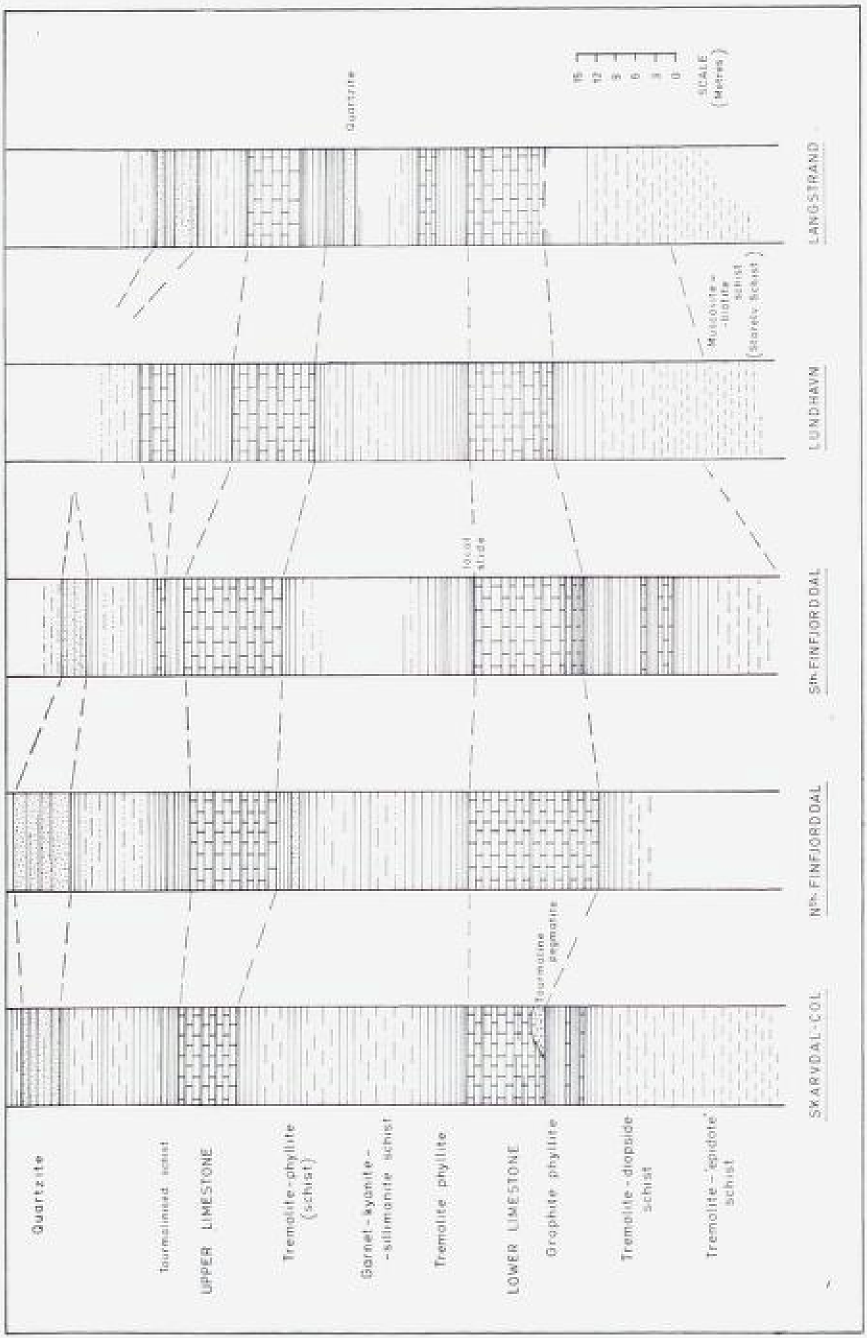

Fig. 10. Comparative sections through the Falkenes Limestone Group.

is generally pale grey or blue-grey with a banding or lamination readily discernible in the form of thin calc-silicate layers. Impersistent tremolite/actinolite and diopside-rich bands are commonly drawn out into boudins and larger isolated segregations of dark green actinolite occur within the limestone. The limestone has clearly been extensively recrystallized, grain size varying to a maximum of 5 mm in certain bands. Staining techniques (Evamy 1962) reveal that the carbonate is apparently 100 % calcite, the calcite being notably iron-free.

Towards the top of the Lower Limestone near Skarvdal-col, a 20–60 cm, indurated, dark blue crystalline limestone band is found to exhibit intensive cataclastic features, and represents a minor slide zone. Within the normal grey limestone, other localised slides are devoid of any notably mylonitic zones.

A tourmaline-bearing pegmatite occurs irregularly in the basal part of the Falkenes Limestone Group as lenticular or pod-like bodies. The contacts sometimes appear to be cross-cutting, but this relationship with the metasediment is principally a tectonic feature.

The Upper Limestone is decidedly more impure often with a brownish weathering surface, and contains frequent calc-silicate and pelitic intercalations which at times constitute the bulk of the rock. In the north-east of the area the degree of impurity is fairly high, whereas towards Langstrand pelitic bands are of infrequent occurrence.

A curious structure noted on a bedding surface of a limestone band near Langstrand school is almost certainly representative of infilled mud — or desiccation — cracks. The rectilinear pattern of 2–3 cm wide ridges (Fig. 11) — the ridges composed of calcareous metasiltstone — is quite similar to those depicted by Shrock (1948, Figs. 152, 156 and 161) and Mikkola (1960, Fig. 3).

The lithology separating the two main limestones is extremely variable comprising graphitic kyanite phyllites and schists, garnet-sillimanite schists and various calc-silicate schists (actinolite and/or diopside being the dominant minerals). Along the coast south of Langstrand a particularly massive, kyanite-rich, black, graphitic schist with yellow and brown weathering stains is quite prominent, the kyanite occurring in radiate or bladed form.

Above the Upper Limestone, a calc-silicate schist is the prevalent lithology although a thin, knobbly-weathering, garnet-rich band is found to contain staurolite, kyanite and tourmaline segregations. Some 15–20 m above the limestone, a conspicuous quartzite marker band is present

Fig. 11. Infilled mud- or desiccation-cracks on bedding surface in impure limestone. Upper Limestone, Langstrand school.

Fig. 12. Deformed 'slump' lithology; psammite in lower left. Hellefjord Schist Group, Gamnes.

over large parts of the area. This is a massive, brittle, sugary-textured quartzite with close, iron-stained jointing which, although relatively thin, is traceable over much of Sørøy. It has been mapped by Sturt and Ramsay in the Sandøfjord and Breivikbotn areas, where it has been termed the White Saccharoidal Quartzite, and by the present writer between Dønnesfjord and Bølefjord during a reconnaissance survey.

5. *The Hellefjord Schist.*

The Falkenes Limestone Group is succeeded by a series of monotonous flaggy metasediments, the thickness of which, although quite considerable, is indeterminable because of tight folds and tectonic stretching and thinning. The metasediment is typically a grey, closely banded, fine-grained phyllitic schist, often extremely flaggy and fissile and containing abundant small garnets and lineated needles or laths of dark green amphibole. Schistosity planes commonly have a satiny lustre. Paler grey or grey-green bands, which tend to be of slightly larger grain size (0.1–0.3 mm), are relatively rich in amphibole and deficient in biotite.

Garnets, though present in varying amount throughout the Hellefjord Schist, are especially profuse in the darker grey phyllitic bands. The small, porphyroblastic, purply-red or translucent garnets rarely exceed 2 mm diameter and exhibit perfect rhombdodecahedral form. Amphibole-rich bands usually contain only scattered garnets. The amphibole laths are generally < 2 cm in length and aligned parallel to F_2 fold axes.

A characteristic feature of the Hellefjord Schist is the presence of segregatory quartz commonly seen as boudins and rodded segregations. Recrystallised diopside and calcite are associated with these boudins in the south-east of the area. As a rule the segregatory quartz is sheared and drawn out parallel to the F_2 axial planar cleavage or schistosity where this is penetrative, and thickening is noted in fold closure zones. Similar phenomena are represented in pegmatitic veins.

The flagginess of the lithology has been accentuated locally by tectonic stretching to the extent that individual 'flags' in some localities are < 1 cm thick and the rock quite fissile. It is of interest to note that the word 'helle' of Hellefjord is translated as 'slate, slab or flag'.

Around the shore- and cliff-sections of Skippernesfjord and Helle-

Table 2. *Modal Compositions of Hellefjord Schists.*

	1 D88	2 D816A	3 D319	4 D607	5 D530A	6 D544
Quartz	23,2	26,6	31,4	32,2	41,5	40,2
Plagioclase	21,4	15,6	20,2	14,4	23,8	24,0
Biotite	48,5	47,9	24,3	23,5	8,4	1,8
Garnet	5,9	7,7	4,2	—	—	—
Amphibole	—	—	4,4	16,8	20,9	28,0
Diopside	—	—	9,8	x	—	—
Scapolite	—	—	3,6	11,7	—	—
Ores	0,1	1,8	0,3	0,4	0,2	—
Apatite	x	0,5	0,6	0,4	0,3	0,3
Sphene	—	—	1,4	0,6	1,8	1,7
Clinozoisite	0,2	—	—	—	0,2	1,7
Tourmaline	0,3	x	x	—	—	—
Chlorite	0,4	—	—	—	—	1,2

1 and 2. Garnet-biotite phyllite.
3 and 4. (Garnet-) calc-silicate schist.
5 and 6. Quartz-amphibole schist.
x Present in minor amounts only.

fjord, rhythmic layering is depicted in alternating paler and darker grey units of phyllite or schist. It is generally not possible to establish a younging direction utilizing graded bedding, since the protracted recrystallisation and tectonic history has usually rendered the grading ill-defined. Along the coast of west Skippernesfjord, thin limestone ribs and laminae appear in the sequence, these showing strong tectonic features and boudinage. S. W. of Lundhavn the lithology changes gradually to a variable sequence of semi-pelitic and schistose rocks, lacking the typical flaggy character of this group.

An unusual rock-type at Gamnes, of restricted occurrence and situated within a banded semi-pelitic or pelitic sequence, has the appearance of a deformed conglomerate or slump-breccia (Fig. 12). It is massive, mostly unbedded though with a poor lamination in its upper parts, and consists of elongate buff or cream-coloured fragments, commonly with sharp outline, set in a dark grey, fine-grained, schistose matrix. The fragmentary material is of variable size and abundance; viewed in end-section, fragments show a general lenticular outline though some are sub-rounded, sub-angular or irregular in shape. There is, in nearly all cases, a crude planar arrangement of the lenticles (as seen in end-section) parallel or sub-parallel to the banding outside this lithology. Fragment size varies from micro-particles to lenticles measuring 4 cm x 15 cm. In longitudinal section, fragments exhibit a pronounced elongation with pencil and cigar shapes dominant.

Fragments are found to consist mostly of fine-grained psammite, often feldspar-rich, the grain size normally smaller than that of the matrix, which is predominantly a biotite schist sometimes containing garnet. Small, similarly lineated fragments of shale, phyllite and semi-pelitic material are also present. A psammite, with traces of current bedding indicating that the beds are normally disposed, occurs immediately below the fragmentary lithology and many of the fragments are composed of this, or of similar, psammitic material.

Where visible — just above sea-level at the foot of a small cliff — the base of the fragmentary lithology is fairly irregular such that this breccia-like rock appears to rest on a channelled or eroded surface of the subjacent laminated psammite. The fragments die out quite rapidly up the sequence and along the strike and there is a perceptible overall gradation of particle size. At the most this lithology is here 8–9 metres thick; it is not possible to determine its lateral dimensions since it occurs at the very tip of Gamnes headland.

Derivation of a large proportion of the fragments from the subjacent psammite and the remainder from lithologies which occur locally is suggestive of the lithology being of non-diastrophic origin, feasibly a slump-breccia or slide-conglomerate attributed to sub-aqueous gliding or slumping. The restricted occurrence, largely unbedded appearance and irregular, unconformable base lend support to this view. It is not impossible, however, that this lithology could have developed in part at least, tectonically, particularly as it is quite strongly deformed. Thin laminae of psammitic material in a semi-pelitic sequence could be dissected, rolled and stretched, simulating the present confused lithology. Despite this possibility, the writer is of the opinion that the 'breccia' was formed subaqueously during the slumping or sliding of unconsolidated material on an unstable submarine slope.

A prominent hornblende-biotite gneiss interrupts the Hellefjord Schist metasedimentary sequence over a large part of the area (Plate I). Its description is reserved for a later chapter. Throughout the eastern part of the area, elongate tectonic lenses of metagabbro are present in these schists. They are regarded as sheared derivatives of the Storelv Gabbro (Stumpfl and Sturt, 1965).

A brief reconnaissance survey has shown that the Hellefjord Schist sequence extends over most of the extreme north-east part of Sørøy, though minor lithological changes have been observed. Hornblende-biotite gneisses are also present but metagabbroic bodies are relatively uncommon.

Précis of conditions of sedimentation.

A relating of the various lithologies to their probable depositional environments is quite possible despite the high grade of regional metamorphism. The massive quartzites of the Klubben Quartzite Group are indicative of a shallow water environment. Cross-bedded units show well developed foreset beds merging asymptotically into bottomset beds which are very occasionally rippled. While the current direction is statistically constant in any one locality, occasional herring-bone cross-lamination reveals conditions of frequently shifting current direction. Although current directions have been recorded in a number of localities, it has not been possible to establish a single original major current direction.

The semi-pelitic members of the Klubben Quartzite Group indicate departures from the littoral conditions to a more sub-littoral or neritic environment, crustal instability possibly having given rise to these changes. The Transitional Group with its rapid alternation of quartzitic and pelitic bands is also suggestive of tectonic instability.

Both the Storelv Schist and Falkenes Limestone Group appear to be representative of a fairly shallow water 'shelf-sea' environment (with probable temporary subaerial conditions as evidenced by desiccation-cracked surfaces), while a deepening of the depositional basin is indicated by the rhythmically sedimented Hellefjord Schist. The slump-breccia of Gamnes is suggestive of a local instability on the (possibly steepening) submarine slope.

In conclusion, it is apparent that the stratigraphical succession exhibits features of changing environments of sedimentation, from littoral and sub-littoral (and possibly in part estuarine) through neritic to a deepening trough (bathyal) analogous to a developing geosyncline.

II. Tectonics

Introduction

In recent years, several workers have been mapping in the West Finnmark Eocambrian province but at present relatively little published material is available on the structural geology of Finnmark. The papers of Sturt (1961), Ramsay and Sturt (1963), Appleyard (1965) and Sturt and Ramsay (1965) provide the only references to the structural geology of Sørøy, though it is expected that a comprehensive account of the broader aspects and implications of the geological structures on Sørøy will be published as soon as the island has been completely mapped. Neighbouring areas of Loppen and Øksfjord have been preliminarily surveyed and a generalized account published (Ball et al. 1963) on the stratigraphy and structure.

The metamorphic rocks of the Langstrand area have been subjected to polyphase deformation. Two distinct episodes of folding are present designated F_1 and F_2, together with a possible later fold phase, F_3, of lesser significance. Furthermore, gentle warp folds, of cross-fold type, were developed during the protracted F_2 folding. The primary schistosity parallels the banding and is related to F_1. A schistosity axial planar to F_2 and often penetrative is found to be the dominant planar structure in many areas, notably in the Hellefjord Schist lithology. Elsewhere, micas have recrystallised mimetically after the earlier fabric, so that little evidence of S_2 is preserved.

Three major structures are recognised, these being the Hønseby Fold (F_1), Langstrand Antiform (F_2) and Hellefjord Synform (F_2). Since the minor structures are dealt with in the first part of the tectonic section of this paper, it will be necessary to consult the many geological profiles drawn across the area (Plate III), together with the structural map (Plate II), in order to relate these structures to the major folds.

Tectonic structures have been divided into three classes by Weiss (1955), namely macroscopic, mesoscopic and microscopic. Major and minor structures are here used in an equivalent sense to the macroscopic and mesoscopic of Weiss. The term microfold is used to describe folds of microscopic dimensions, including minor drag folds and corrugations.

1. Folds and linear structures

Geological mapping of the area has revealed a striking homogeneity and uniformity of succession, minor facies changes apart, so that subdivision on the basis of major structural units has not been possible. The ensuing description and discussion of folds and related structures present over the area is divided into tectonic episodes which are part of the protracted Caledonian orogeny.

One of the most notable features of the geology and particularly of the structural geology is the marked strike swing of some 90°–120° from Saksfjord in the north to Langstrand in the south. The conformity of succession together with the relatively gentle swing of strike gives a deceptively simple picture of the structures involved. As can be seen from the structural map and cross-sections (Plates II and III), the larger part of the area shows the banding dipping to the south-west or west, whereas the Klubben Quartzite Group to the south of the Hønsebyfjorden is characterised by S-SE dipping strata. The detailed significance and interpretations of these observations are discussed later in the structural chapter. It should be noted, however, that near-recumbent folds of major proportions dominate the tectonic picture.

Of major importance from both stratigraphical and tectonic points of view are the well-preserved sedimentary structures in the massive quartzites. These criteria indicate that strata in well over half the area mapped are inverted. Moreover, the presence of such structures has proved invaluable in locating the axial traces of tight folds in strata with near-constant dip and strike values, which would otherwise have been virtually impossible.

Despite the evidence afforded by current bedding, the magnitude of the major folds and their problematical relationships with other structures deems it necessary to consider the structural geology of other parts of Sørøy in formulating a large-scale picture of the tectonic history. Unfortunately the areas in proximity to the Langstrand region, which would be expected to provide the most valuable information for structural interpretation, have yet to be surveyed in detail. The writer has summarily traversed the extreme north-east part of Sørøy which appears to be characterized by a monotonous sequence of flaggy Hellefjord Schist, while Dr. D. M. Ramsay has recently begun a survey west of Langstrandfjord.

A. First generation minor structures – F_1.

Minor folds.

The minor folds of known F_1 age fold a banding which, in most cases, is clearly of sedimentary origin. Over certain parts of the area however, appreciable tectonic stretching has occurred; slip between lithological units and modification of the primary features are of considerable importance in this context.

Early minor folds are noticeably fewer in number than F_2 minor folds. Evidence with regard to the precedence of folds is the refolding of one structure by another; in this instance, F_2 refolding F_1, it has been observed in only six localites, although other larger scale examples of refolding are demonstrable.

No great variation is seen in the style of the F_1 minor folds. Tight or isoclinal, symmetrically disposed structures prevail, of closed U-shape, as illustrated in figures 13 and 14. Amplitudes of the folds are usually indeterminate and immeasureable, whilst wavelengths are as small as 6–7 cm. Folds are neither perfectly similar nor concentric, with a variable degree of thickening at closures. The sense of overturning of F_1 minor folds is unknown since it is impossible to differentiate between long and short limbs in these isoclinal folds. F_1 fold closures display an axial plane schistosity at times visible only on close examination. Along the limbs of these early folds schistosity essentially parallels the banding, though this has often been modified by later recrystallisation and by stretching and slippage during the F_2 deformation. In areas where second generation folds abound, particularly in the Hellefjord Schist, the dominant schistosity though apparently parallel to the banding can be seen on detailed examination to be slightly discordant to it. The earlier schistosity is, thus, almost entirely erased by the later recrystallisation and mineral growth, and indeed the only evidence of an earlier fabric may be confined to that contained within porphyroblasts.

While many outcrops of the closely banded psammites and of the Hellefjord Schist do not exhibit any noticable early folding, certain areas provide several examples of F_1 fold closures .In the North Skarvfjord cliff- and coast-section composed principally of quartzites, minor early folds are relatively abundant. In Fig. 13 one prominent example of an early isocline can be seen but other less readily discernible fold closures can be located to the right of this fold. Because of the extensive stretching phenomena associated with the F_2 folding in this area the

Fig. 13. F_1 isoclinal folds in psammite. Klubben Quartzite Group, north shore of Skarvfjord.

present styles and amplitudes of early folds cannot be regarded as representative of typical F_1 folds in these psammites. They are clearly modified by the later tectonic episode, and much of the attenuation and thinning-out of bands within the limbs of such folds is here attributed to this later, F_2 deformation. On account of these later movements, it is not possible to determine the original style of the early folds prior to the reorganisation of the structures.

Axial directions are rarely constant, so much so that it is impossible to speak of a dominant F_1 axial trend. Curved axes are observed; whether such curvature is a contemporaneous product of the initial

Fig. 14. F_1 fold in Hellefjord Schist, 1 km north of Hellefjord.

Fig. 15. Eyed fold and other F_1 minor folds deformed by F_2 structures. Limestone of Falkenes Limestone Group, Langstrand.

folding or whether it results from fold superimposition is in many cases indeterminate. In certain areas of Sørøy visited by the writer, F_1 folds with curved axes occur in some limestones, the curvature being unrelated to later deformations. Fig. 15 depicts refolded folds together with an 'eyed' fold (on a flat surface) in impure limestone from Langstrand. The closed eyed fold is consequent upon axial plunges in directions oblique or near-orthogonal to one another. The early folds are essentially flow folds, and on account of this and the relative plasticity of the limestone during deformation the inference is that axes could not be regarded as having any constancy of direction ensuing the folding. In view of this, it is conjectural to determine the extent to which refolding has contributed to the development of this closed fold pattern. Voll (1960) has noted that axes of folds begin to curve during advanced stages of deformation, though in certain instances curvature may arise from the beginning of folding. This same writer has, in some examples of varying (curving) fold axes, assumed the direction of flow — while generally normal to the axes — to vary within the s-plane associated with the folding.

Nicholson (1963), seeking an explanation for certain closed eyed folds and other similar non-cylindroidal structures in marbles from Nordland, Central Norway, partially rejects fold superposition and regards their formation as a possible result of uniaxial compression. Such a mechanism was also favoured by Ramberg (1959) in explaining similar closed forms occurring in ptygmatic veins: he noted that these closed structures are not uncommon in veined gneisses and migmatites. These rocks, like most limestones, were probably quite susceptible to plastic deformation. This explanation — production of the closed folds in one phase of movement, by uniaxial compression — is favoured by the writer for the Langstrand eyed folds, though some degree of fold interference cannot be entirely discounted.

Examples of refolding involving early minor folds are found in the psammites of the Skarvfjordhamn area, and sometimes in those metasediments which have been partially granitised. In the coarse Storelv Schist and in the calcareous schists of the Falkenes Limestone Group early folds are rarely visible, but it is clear that the schistosity which is now deformed by younger structures is that which elsewhere is parallel to the axial planes of F_1 folds. Apparent absence of early folds in large tracts of Storelv Schist is considered not to imply the non-existence of such folds but to infer that the development of schistosity, along with later recrystallisation, has dominated the rock fabric. Hence the accompanying folds are rarely discernible, except where the homogeneity of the lithology is disturbed by psammitic intercalations.

Linear structures.

Despite the presence of F_1 folds in several members of the stratigraphical sequence, early linear structures are generally lacking, this being partly a consequence of the main metamorphism and second episode deformation (F_2) which have tended to obliterate the earlier structures.

Intersection of bedding and schistosity, here essentially the linear elements produced by the intersection of S_1 and S_2, is poorly represented in this area on account of the pervasive later recrystallisation and more intensive deformation. Other tectonic linear structures such as boudinage, mullions, quartz rods and mineral lineations cannot be unequivocally recognised as being associated with the F_1 deformation except in isolated cases.

B. Second generation minor structures – F_2.

Minor folds.

Over the whole area of north-east Sørøy minor structures associated with the second episode of deformation are extremely abundant and constitute the dominant linear element in the region. Indubitable structures of F_2 age are seen principally in:

(a) the folding of the fundamental early schistosity (S_2) with an associated cleavage or schistosity.

(b) the presence of a lineation produced by the intersection of S_1 and S_3, or S_2 and S_3.

(c) the folding of earlier minor folds.

In addition, linear elements such as mullions, boudinage, rodded quartz and lineation of amphibole are demonstrably associated with the second generation of folds.

The style of the F_2 folds is quite variable, this variation dependent mainly upon (i) local tectonic environment and (ii) differences of lithology. A comparative study of the Skarvfjordhamn–Hønseby and Hellefjord–Skippernesfjord areas, for example, reveals appreciable differences in fold style due partly to a variable intensity of deformation but largely to differences of lithology. Within any one major lithological unit however, F_2 fold style may vary considerably solely on account of differing intensities of deformation; such style variations are displayed to good effect in Quartzite 3 of the Klubben Quartzite Group when this particular psammitic horizon is traced across the mapped area.

For reasons of fold style variation and abundance and also of lithological differences, descriptions of F_2 minor folds are divided conveniently and geographically into five sub-areas (Fig. 16). Other minor structures will be discussed later in the chapter.

A generalisation regarding F_2 minor folds is that they are almost wholly asymmetrical and are, with few exceptions, overturned to varying degrees. The degree of recumbency is generally conditioned by the local intensity of deformation, but again lithological character is not unimportant in this context. Where interbedded psammitic and pelitic bands are folded, corresponding variations in fold style are noted according to lithology and position relative to the main fold closures.

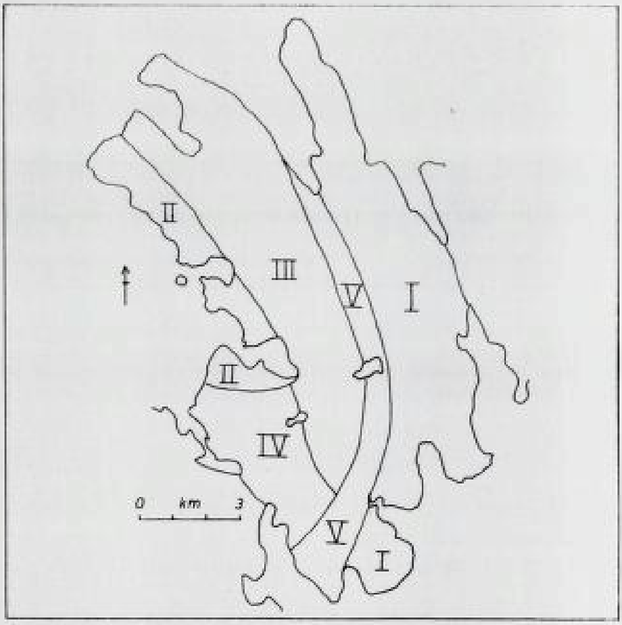

Fig. 16. Sub-areas used in F_2 fold descriptions. I – Hellefjord–Finfjordnæringen area: II – Skarvfjordhamn–Hønsebyfjord area: III – North and central area: IV – Veststrømmen–Veirbukten area, V–The schist – limestone belt.

(a) *The Hellefjord–Finfjordnæringen area.*

This is a 2–3 km wide area extending from the Lundhavn–Hellefjord coast in the south to the Finfjord peninsula of Finfjordnæringen in the north, and consisting essentially of flaggy, rhythmically banded Hellefjord Schist. As a rule, dips are shallow to the south-west, though in the south the major Hellefjord Synform partly disrupts this general picture with strikes swinging round to N-S (Plate II).

Most of this sub-area constitutes the lower limb of the Hellefjord Synform and here the sense of overturning of the F_2 minor folds is towards the N.E. or E.N.E.: consequently the related axial plane schistosity is notably steeper than the banding in the metasediments. Both on the inverted western limb of the Hellefjord Synform and on corresponding limbs of the larger folds along Finfjordnæringen, minor folds exhibit overturning towards the west and S.W. respectively. The axial plane schistosity is here less steep than the banding. It is clear therefore, that all F_2 minor folds are congruous with respect to the larger F_2 folds, including the major Hellefjord Synform.

Asymmetry is ubiquitous in these minor folds. While the majority are overturned, many approaching recumbency, in some instances the shorter limbs do not exceed the vertical. In several cases, minor folds can be traced from unflexured bands through to overturned folds of usually small wavelength and amplitude.

Invariably there is a thinning of the banding along the long limbs of these F_2 minor folds and a relative thickening at fold closures and along short limbs. In this respect the folds can be likened to similar folds

Fig. 17. Typical F_2 folds in Hellefjord Schist, Skipsvik, Finfjord.

Fig. 18. F_2 monoclinal fold in Hellefjord Schist. West Bastafjord.

though many variations are observed and parallel folds are also demonstrable (Fig 18). It is impossible therefore, to postulate any one deformational mechanism as being solely responsible for F_2 fold development in this sub-area. Clearly, various combinations of flexural and shearing mechanisms have to be considered. Where folds approach a similar style, shearing is often visibly more prominent and the axial plane schistosity, S_3, is penetrative. Throughout all this sub-area S_3 is, indeed, pervasive though not always conspicuous, becoming less conspicuous where flexural folding is present and where the lithology is less pelitic. Fold style may vary within one outcrop: complementary folds in Fig. 19 are both flexural yet one is V-shaped whilst the other displays a distinct U-form.

The F_2 axial planar schistosity is best seen in the south of the sub-area, particularly where the lithology is of a more pelitic nature. Fig. 20 depicts S_3, here more of a pronounced cleavage, transgressing the banding and early schistosity at 15°–18°; S_3 is conspicuous in the darker pelite but more or less absent in the pale psammitic bands. Along the Nordnes coast, weathering along both S_2 and S_3 planes has produced a rather unusual surface with en échelon lozenge-shaped protuberances each ca. 1 cm in breadth. These are essentially 'micro-mullions'. Nearer

Fig. 19. Complementary F_2 folds of differing style. Hellefjord Schist, West Bastafjord.

Fig. 20. Slaty Hellefjord Schist with S_3 oblique to banding.

Fig. 21. F_2 fracture cleavage (S_3) oblique to banding (S_1): hammer-shaft parallel to the banding. Hellefjord Schist, 600 m N.W. of Hellefjord.

the closures of the Hellefjord Synform the lithology is less flaggy, S_3 often simulating a fracture cleavage which in places almost obscures the banding (Fig. 21).

Apart from fold axes, the alignment of amphibole needles or prisms is the most notable linear element over this sub-area (Fig. 22): this mineral lineation, in common with most other linear structures (described later), is always parallel to the fold and tectonic 'b' axes. It would thus appear that monoclinic symmetry of movement has prevailed, wherein the 'ac' plane and deformation plane are coincident and the translation in this plane is normal to the 'b' axis which is an axis of binary symmetry. Fold styles, in general, point to shearing stress as the likely major deformative mechanism responsible for their development but the presence of a number of folds of concentric style suggests that flexural-slip was operative locally.

Brief mention can be made here that the massive biotite-hornblende granitic gneiss occurring extensively over this sub-area (Plate 1) has been subjected to F_2 deformation. Minor structures of this generation are abundant in the gneiss, particularly noteworthy being the marked lineation of rodded quartz-feldspar clots (Fig. 53). The lenticular

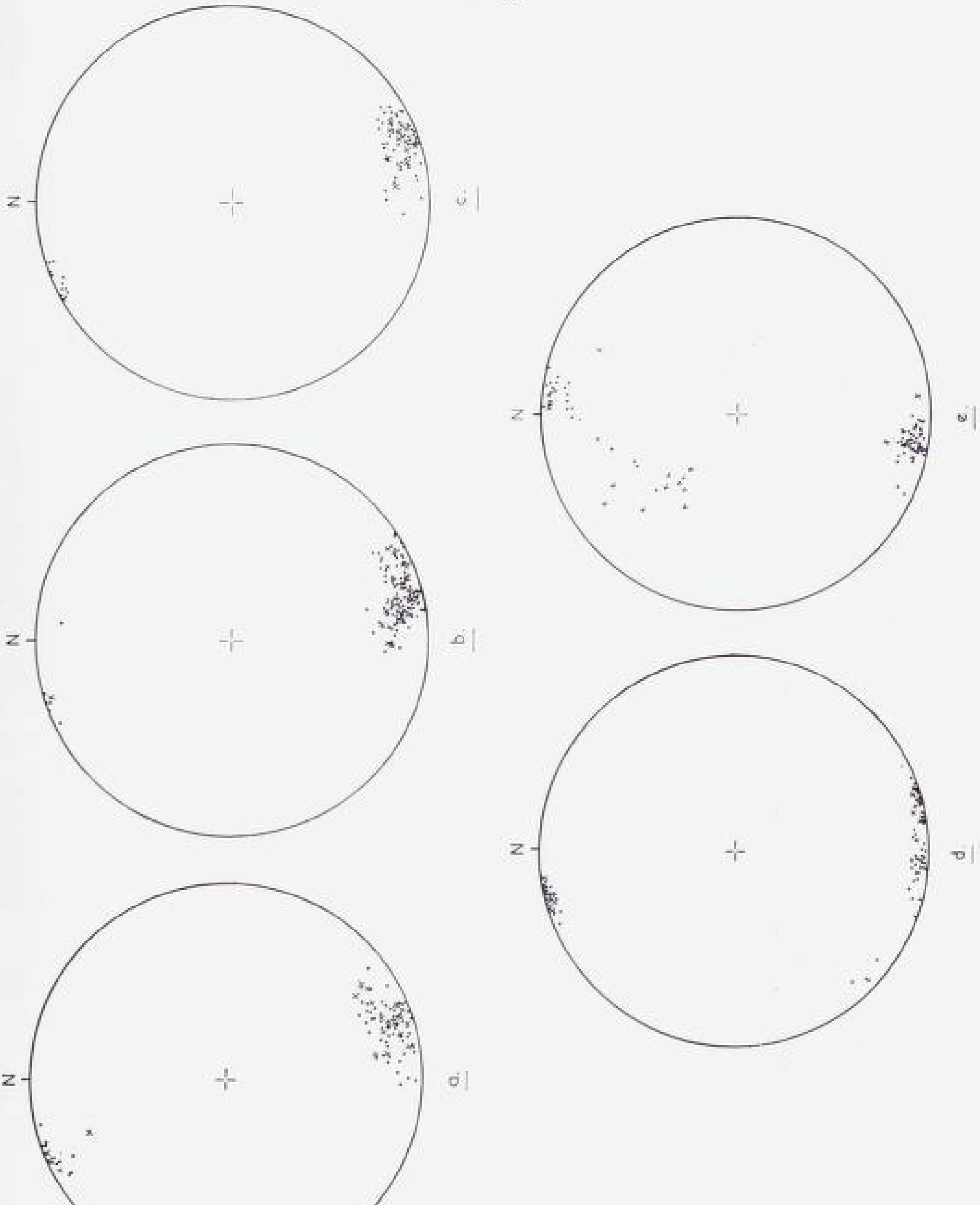

Fig. 22. Structural data (F_2) from the Hellefjord–Finfjordnæringen area; (a) Finfjordnæringen, (b) Veiviken–Finfjordfjell (c) Basteidet–Otervikfjell (d) Hellefjord–Skippernesfjord (e) Lundhavn–W. Hellefjord. Dots – mineral lineation; crosses – F_2 minor fold axes.

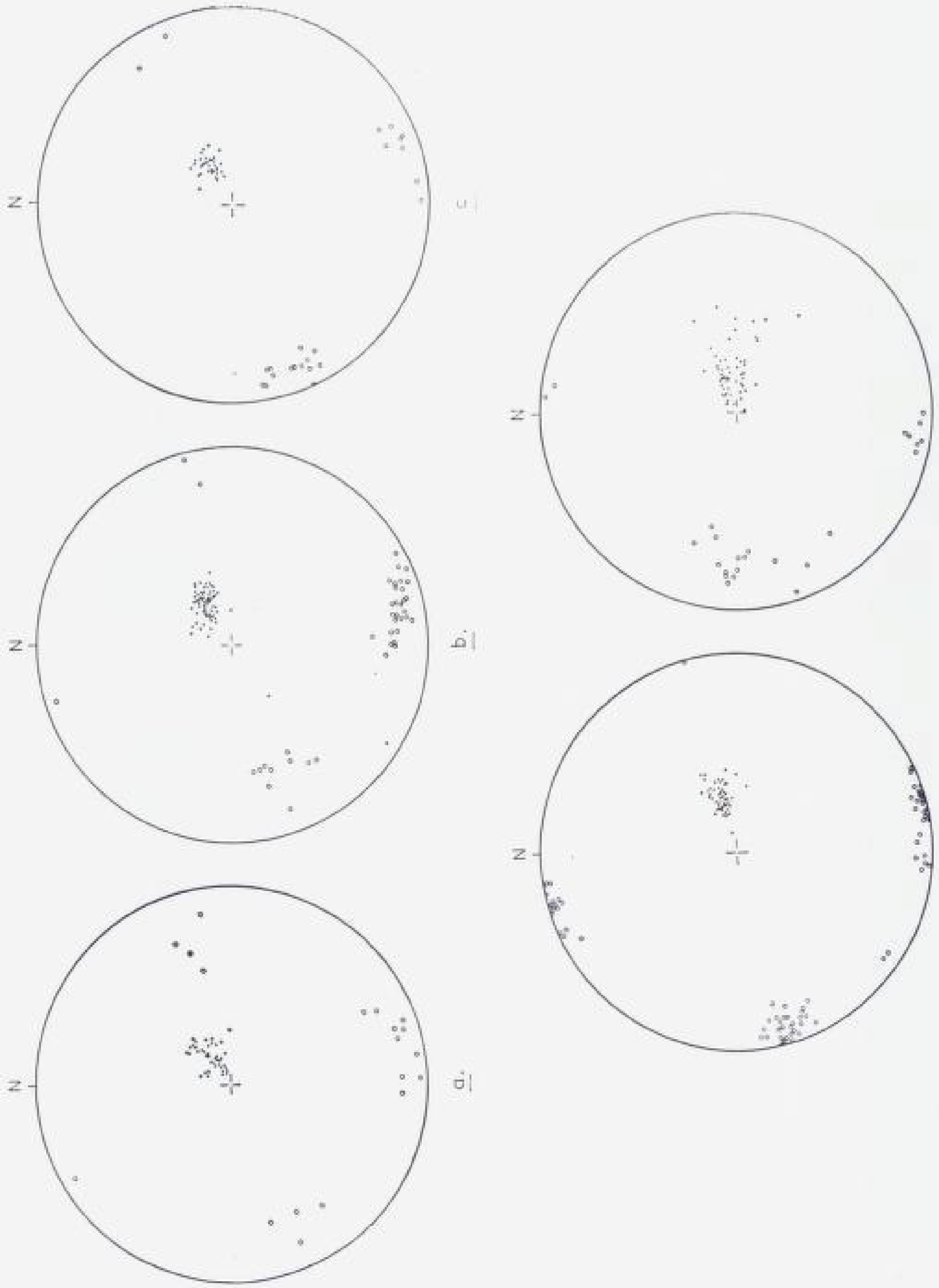

Fig. 23. Structural data (F_2) from the Hellefjord–Finfjordnæringen area; sub-areas as in Fig. 22. Dots – poles to F_2 axial planes; circles – boudinage and quartz rods.

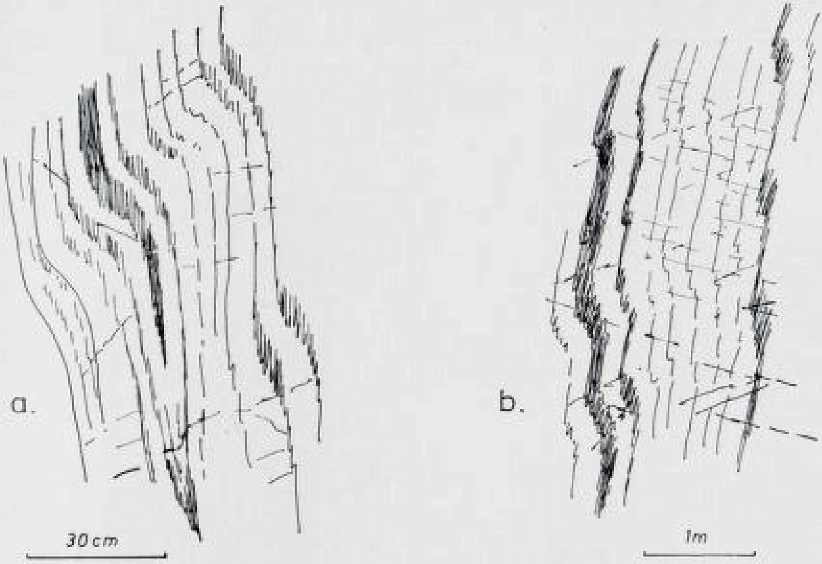

Fig. 24. Incipient F_2 minor folds, psammite, North Skarvfjord: refolded early (F_1) fold present in sketch (a).

amphibolite or metagabbro sheets also exhibit clear evidence of having been deformed by the F_2 movements; a lineation of hornblende is notable, while F_2 minor folds also deform these amphibolites.

(b) *The Skarvfjordhamn – Hønsebyfjord area.*

Along the eastern coastline of Sandøfjord the Klubben Quartzite Group, and in the north-west the Storelv Schist, is involved in F_2 folding of a higher intensity than that existing elsewhere. Consequently, second generation minor structures attain a stage of development not present in other parts of N.E. Sørøy, this being particularly so on the near-vertical western limb of the Skarvfjord Synform.

Over the whole of this sub-area F_2 folds, whatever their magnitude, are deforming an inverted sequence, as evidenced by sedimentary structures. These folds are usually tight and near-recumbent, except on the steep western limb of the Skarvfjord Synform; there tectonic thinning of quartzite bands is sometimes extreme to the extent of causing the disappearance of units. Minor F_2 folds occur infrequently in this near-vertical zone and may display only incipient development (Figs. 24 and

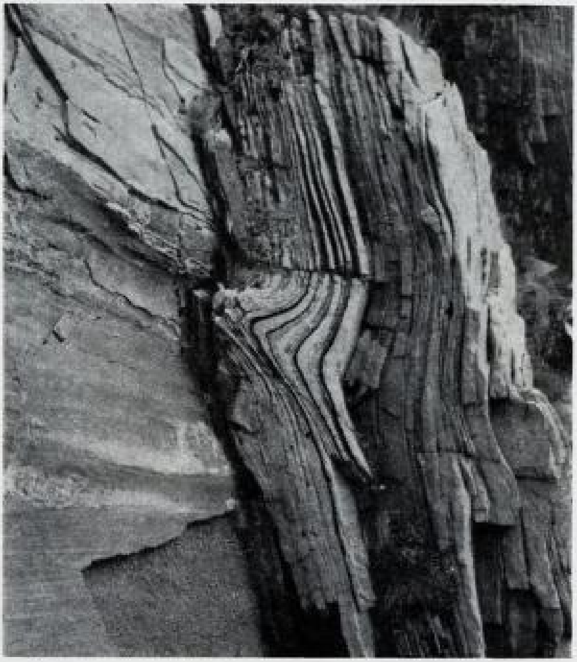

Fig. 25. F_2 minor fold on west limb of Skarvfjord Synform. Psammites, 200 m N.E. of Rødbergodden.

25). In the latter figure the growth of the fold is well illustrated, as is the thickening of bands at the closure. Such folds, traceable from nothing to a local maximum development then diminishing again beyond the zenith both in the 'ac' plane and in the 'b' direction, have been called 'pod folds' (Mendelsohn 1959). They may be arranged en échelon.

On this steep western limb of the Skarvfjord Synform, S_3 is prominent in the thin pelitic laminae, often displaying a fanning around the minor fold closures. Near Hønsebynes, on the equivalent steep limb of the synform extension, F_2 axial planar shearing is demonstrable with aplitic veins emplaced along these shear zones. Thinning and pinching-out of bands occurs extensively. Thin amphibolite sills appear to have been the most vulnerable in this respect and may be drawn out into isolated 'tectonic inclusions'.

The flatter eastern limb of the Skarvfjord Synform is characterized by an abundance of minor F_2 folds and related structures not developed

Fig. 26. F_2 folds in mixed psammite-pelite sequence, Klubben Quartzite Group. N.W. Hønsebyfjord.

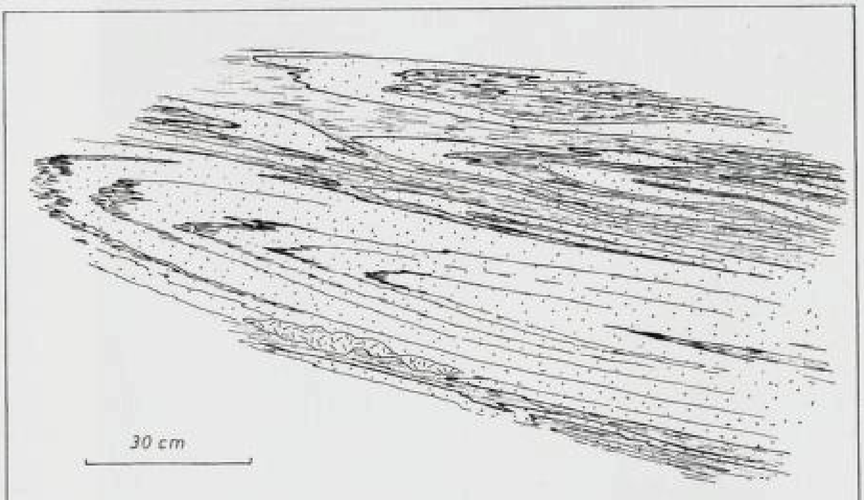

Fig. 27. Shearing of tight F_2 folds, mixed psammite-pelite, Skarvfjordhamn. Sketched from joint face 'ac' to F_2 axes.

so intensively elsewhere. Dips generally vary between 15° and 40° to the W.S.W., fold overturning being constantly towards the E.N.E. A marked axial plane schistosity, S_3, is common throughout the area, becoming particularly penetrative where minor folds occur in pro-

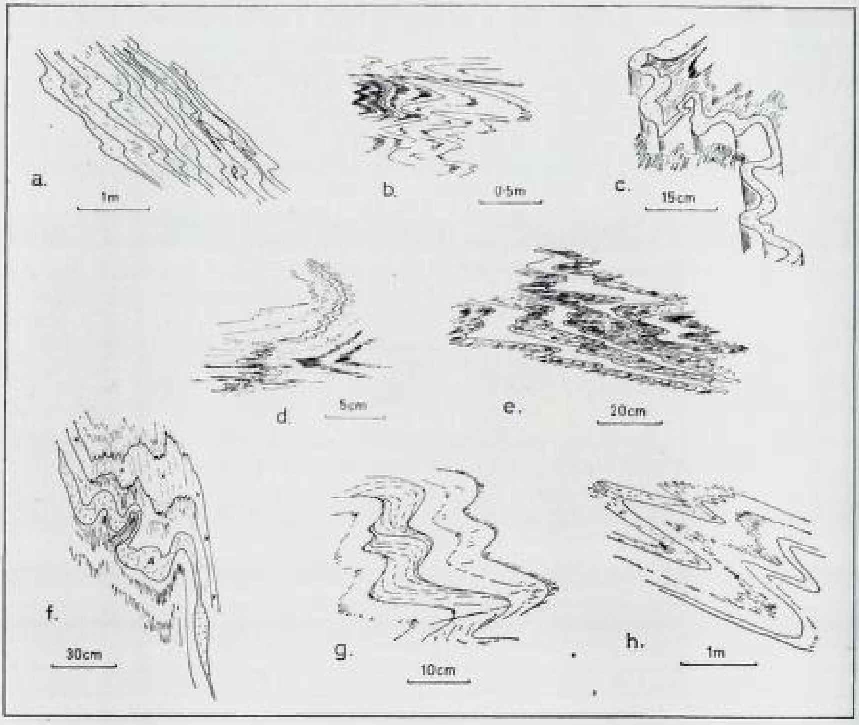

Fig. 28. F_2 structures, Skarvfjord–Hønseby area. (a) Fold style on reverse limb of larger fold, psammite-pelite. (b) Changing fold style, with shear plane developed. (c) Disharmonic 'oak-leaf' fold, quartzite band. (d) Shearing of aplite vein. (s) Well-developed S_3 in pelite (with psammite bands). (f) Disharmonically folded aplite (A); amphibolite sill (X) shows local development of garnets (dots). (g) Current bedding deformed by F_2 folds, massive psammite. (h) Typical F_2 folds, psammite.

fusion. By noting the attitude and relationship of S_3 to the banding it has been possible to map the axial traces of F_2 folds of intermediate size.

Minor folds are generally similar, in both the descriptive and generic sense, invariably showing thickening at fold closures and also where short limbs are poorly developed (Fig. 26). Axial planes rarely exhibit perfect parallelism due to factors such as shearing, inconstant laminar slip and variations in lithology. Uncommon examples of concentric folds are present, as are several transitional fold styles which conform neither to the type pattern of similar folds nor to that specified for parallel folds. Many minor F_2 folds in the vicinity of Skarvfjordhamn and Grasdalsnes have been involved in pronounced shearing and

Fig. 29. F_2 fold with fracture cleavage, psammite, N.W. Hønsebyfjord.

stretching causing the thinning and severing of limbs (Fig. 27). Folds of a disharmonic style are occasionally observed (Fig. 28 c and f).

Lithological control has a decided effect on the resultant style of F_2 folds, as does their position in relation to larger fold closures. With pelitic horizons within psammites, puckers and crumples of small amplitude characterize the closure regions of folds of all magnitudes. Quite different is the fold depicted in Fig. 29 which shows a coarse fracture cleavage in fairly massive quartzite. Amphibolite sills and sheets are common in this sub-area and are clearly pre-F_2 in age (Fig. 30). Such amphibolites normally display a perfect S_3 schistosity and may be drawn out into disconnected portions by shearing.

This sub-area also includes the schists which crop out to the north-west of Skarvfjordhamn, above the Klubben Quartzite Group. Considerable shearing and sliding has taken place in this lithology during the F_2 period and minor F_2 folds are mostly restricted to micro-corrugations and puckers. In the extreme north-west in a more semi-pelitic schist minor folds are again abundant, these being congruous with respect to the larger structures. An axial plane schistosity is generally quite pronounced (Fig. 31) though often poorly developed and refracted

Fig. 30. Garnet amphibolite sill deformed by F_2 folds. Metasediment (psammite with pelite ribs) partially granitised. Grasdalsnes.

in the psammite intercalations. In one horizon of mica schist ptygmatic folding is prominent (Fig. 32); the ptygmas are composed of aplitic or pegmatitic material and are non-schistose. With regard to their generation, there have seemingly been differential responses to the imposed stresses, the metasediments apparently less rigid than the vein during deformation and injection.

Cross folds assigned to a late F_2 stage are also discernible in the N.W. schist area, and are to a large extent responsible for the variation in linear trends (see Fig. 33) noted north of Skarvfjord. Plunges are southerly near Skarvfjord but north-westerly further north. This marked divergence in tectonic 'b' axes seems to be characteristic of the schist lithologies — it occurs elsewhere in the Storelv Schist as well as in the semi-pelitic lithology east of Saksfjord — while the quartzites within this sub-area do not show any such notable divergent linear

Fig. 31. F_2 folds with well-developed axial plane schistosity (S_3) in pelite bands. Upper part of Storelv Schist, 2 km NNW of Rødbergodden.

trends. It would thus appear that lithology is, in part, a controlling factor in the variability, or constancy, of 'b' axes. This has been noted by Ramsay and Sturt (1963) in the west Sandøfjord area.

Summarizing the features representative of this Skarvfjordhamn – Hønsebyfjord area, it is abundantly clear that F_2 minor folds on the steep limb of the major Skarvfjord Synform have been confined in their growth, in contrast to minor folds characteristic of the shallower eastern limb. The causative factors for this retardation of fold development are considered to be the protracted shearing, stretching, laminar slip and thinning of lithological units attending the F_2 deformation. Monoclinic symmetry of movement characterises this deformation, the pattern being disrupted by cross-fold warps related to the development of conjugate shear planes and boudinage. Petrofabric studies of psammites from near Hønseby corroborate the evidence of monoclinic symmetry since in all of 29 prepared analyses of quartzite thin-sections cut normal to the megascopic F_2 'b' lineation, the dominant type of quartz(0001) orientation is a peripheral 'ac' girdle. Asymmetrical maxima in the peripheral girdles (Fig. 34) point to a homotactic fabric and

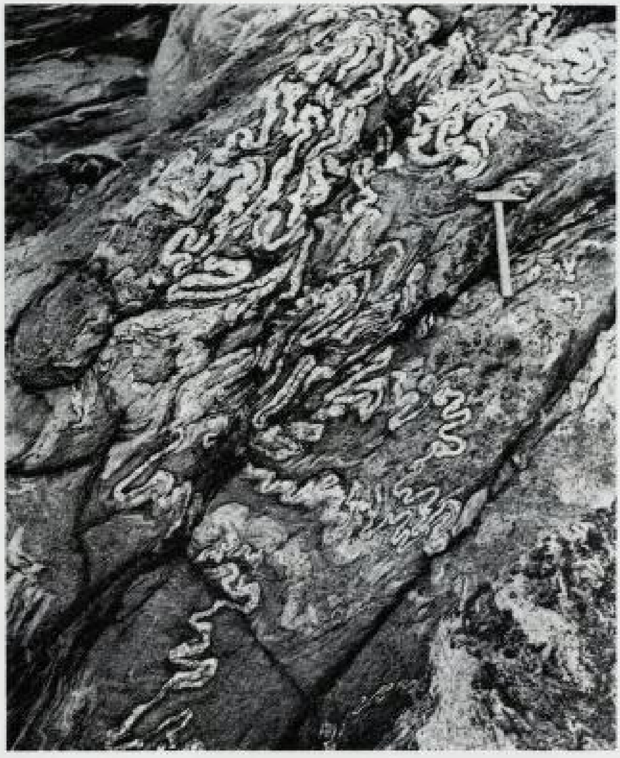

Fig. 32. Ptygmatic folding of aplite veins in Storelv Schist, 900 m NNW of Rødbergodden.

monoclinic symmetry. In one diagram (Fig. 34a) from the long limb of an F_2 fold, a cross-girdle is present; this is interpreted as a relict F_1 fabric. A cross-girdle is not represented at the closure of the same F_2 fold (Fig. 34a), nor is one present in any other diagram of fabrics from this strongly F_2-oriented environment.

(c) *North and Central area.*

Essentially this is the broad zone of psammites between the Hellefjord and Skarvfjord sub-areas already described, but excluding the narrow limestone/schist belt. In the northern sector dips are generally to the S.W.; inland from Hønseby (to the south) the westerly dipping banding turns over through the vertical, then striking ca. N.E.-S.W.

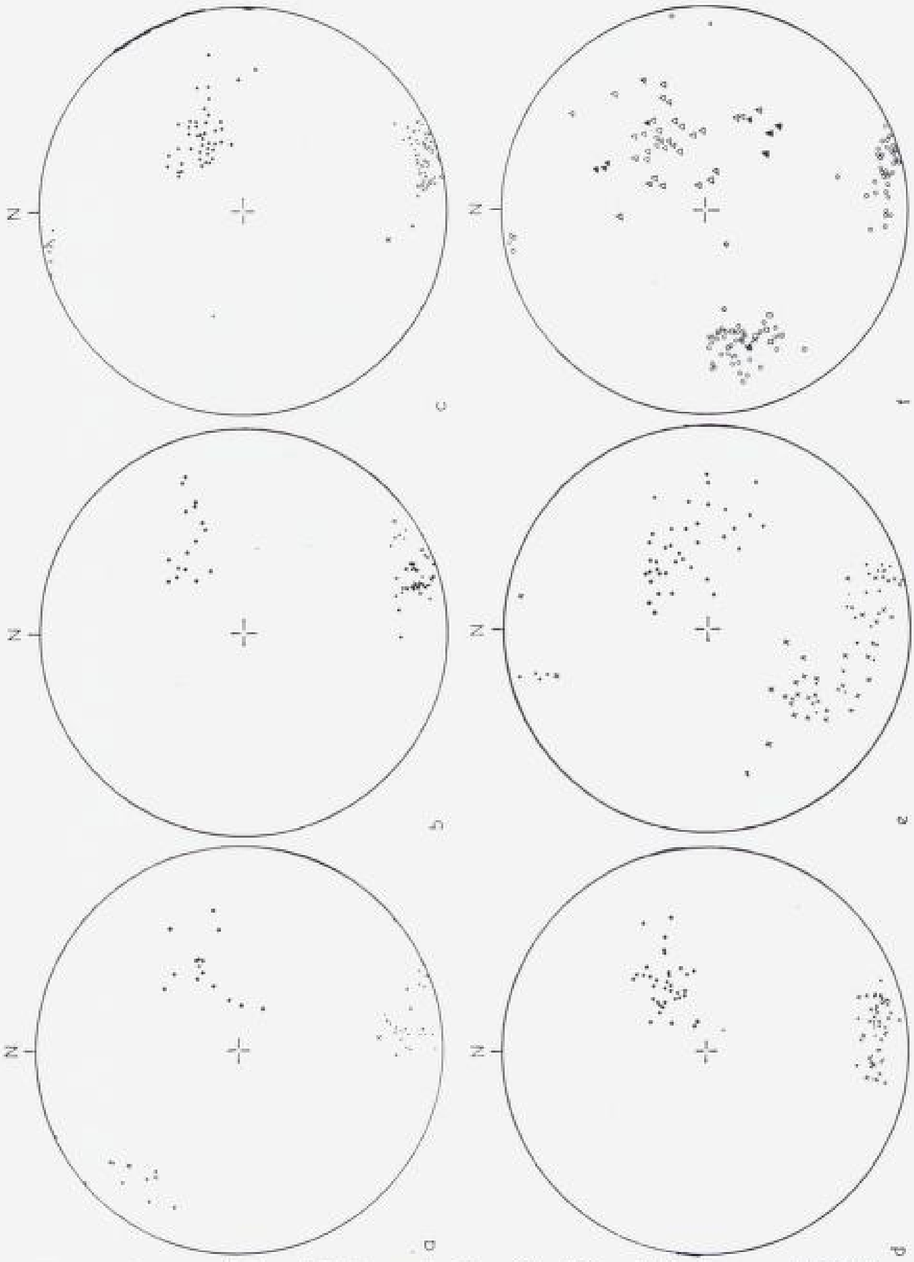

Fig. 33. Structural data (F_2) from the Skarvfjord–Hønsebyfjord area. (a) N.W. schist area. (b) North Skarvfjord. (c) Skarvfjordhamn–Grasdalsnes. (d) North Hønseby. (e) Hønsebynes–S. Hønsebyfjord. (f) Skarvfjord–Hønseby. Crosses represent fold axes; large dots – poles to F_2 axial planes; small dots – S_1/S_3 lineation; circles – boudinage axes; squares – cross fold axes; open triangles – poles to shear planes; full triangles – poles to conjugate shear planes; S – plunge of intersection of conjugate shear planes.

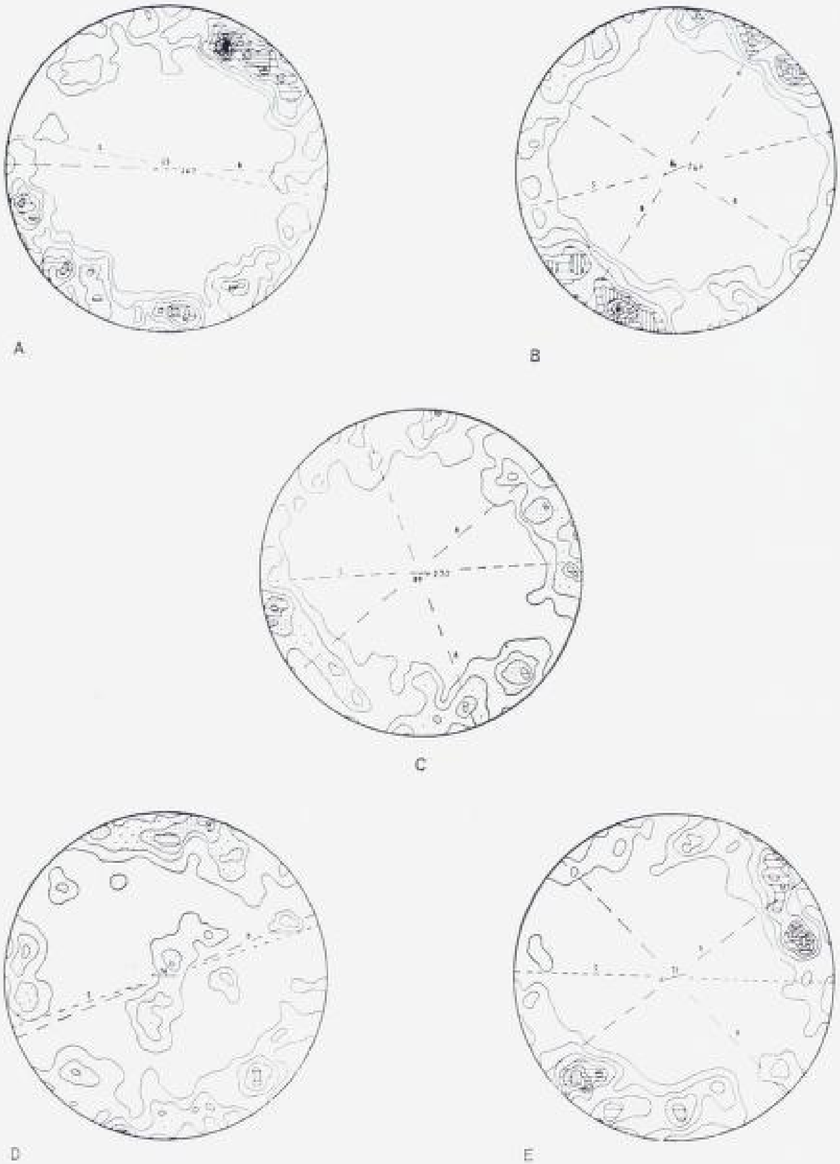

Fig. 34. Representative petrofabric diagrams – quartz optic axes. Klubben Quartzite, N.W. Hønsebyfjord. The two lowermost diagrams are from the long limb (d) and closure zone (e) of one particular minor F_2 fold (see text). Contours 1, 2, 3, 4, 5, 6, 7, 8 %. B – banding, S – S_3.

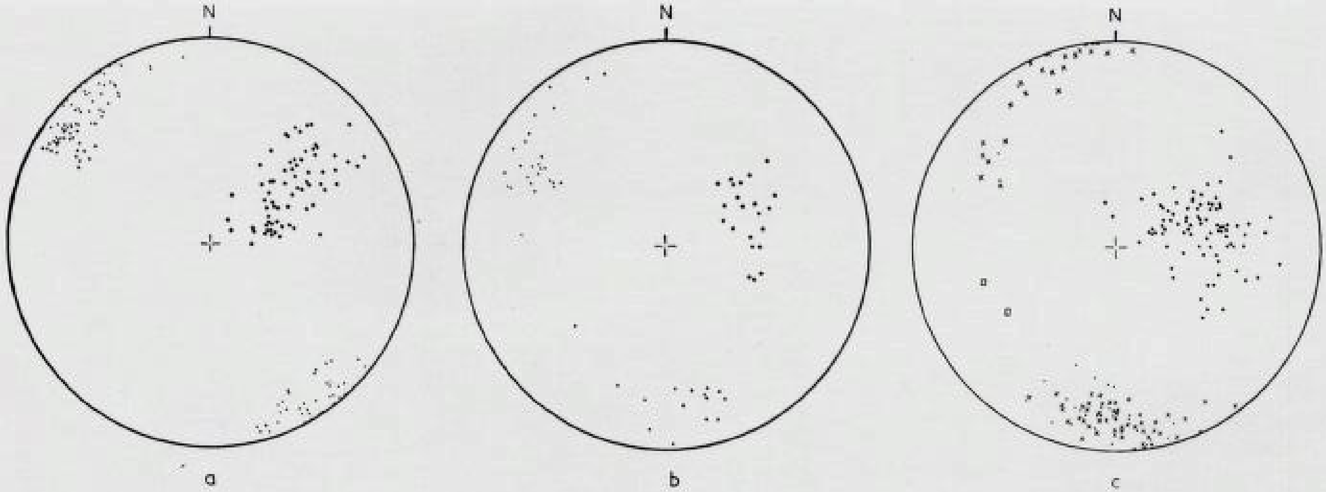

Fig. 35. Structural data (F_2) from the North and Central area. (a) Saksfjord. (b) Skarvfjordfjell. (c) Central fjell. Symbols in (a) and (b); small dots – fold axes and lineations; larger dots – poles to F_2 axial planes. Symbols in (c); crosses – fold axes; small dots – lineation (S_1/S_2); larger dots – poles to F_2 axial planes; squares – cross fold axes.

with a south-easterly dip. The zone of vertical dips marks the approximate axial plane trace of the Langstrand Antiform (Plate II).

A certain homogeneity of fold style exists in the region north of the Skarvdal Fault, broken only by differences attributed to lithological change and by position relative to mappable folds (i. e. whether on normal or inverted limbs of larger F_2 folds). F_2 axial planes dip to the S.W., fold overturning being north-eastward on normal limbs of larger structures (for local details see Roberts 1965). F_2 structural data are shown on the stereograms (Fig. 35).

An important exception to the general rule of fold overturning is found in a small area east of Skarvfjord. Incongruous asymmetrical folds here occur together with normal congruous structures and it is observed that the development of both set of folds, though more especially the incongruous type, is sometimes associated with complementary shear planes. This is in accordance with the observations of Ramsay and Sturt (1963) in the Storelv area of Sørøy. Fig. 36 depicts an example of these conjugate shears and folds.

Only three true fold pairs have been observed in this small area, although non-conjugate abnormal folds (with easterly dipping axial planes) are also present. Axial plane schistosity is rarely well developed but tends to be more advanced in the normal congruous folds. The axes of the two fold sets do not show any outstanding divergence of trend, the variation in the order of only 0°–6°. Because of the scarity of conjugate fold observations east of Skarvfjord, it is not possible to compare these occurrences with any accuracy or certainty with the pattern of fold pairs at Storelv. In consideration of the movements involved in producing these conjugate folds, an orthorhombic symmetry prevails. This is interesting in that it is a localized variant in an environment of monoclinic symmetry.

The development of F_2 minor folds in the area south of Hønsebydalen has been controlled largely by lithology but also by position in relation to minor and major fold closures. Where psammites predominate minor folds are sometimes asymmetrical but not overturned to any great extent, if at all (Fig. 37b and d). A more acute style is seen in the semi-pelitic horizons with an axial plane schistosity or cleavage frequently well developed. At some localities, where erosion has fortuitously permitted their inspection, en échelon 'pod folds' of variable size and style are demonstrable.

Folds of a remarkably open style with wavelengths up to 300 m

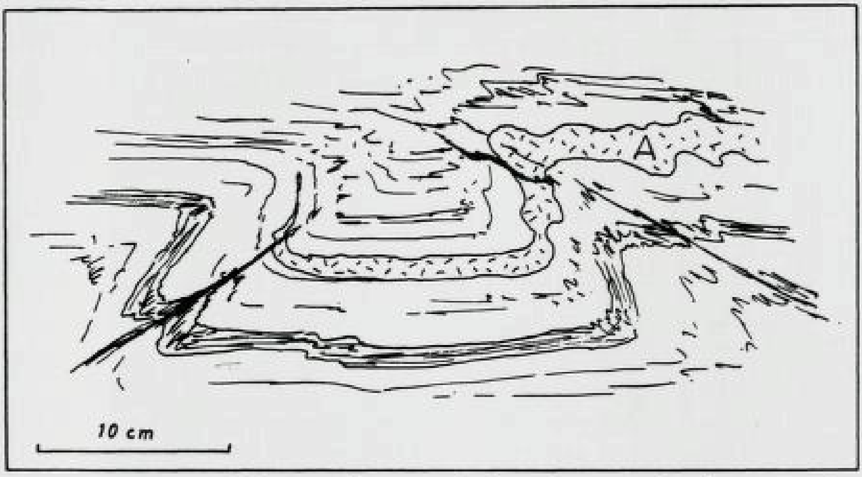

Fig. 36. Conjugate F_2 fold pair with associated shear planes. Mixed pelite – psammite sequence with aplite vein (A). 150 m E. of Skarvfjord.

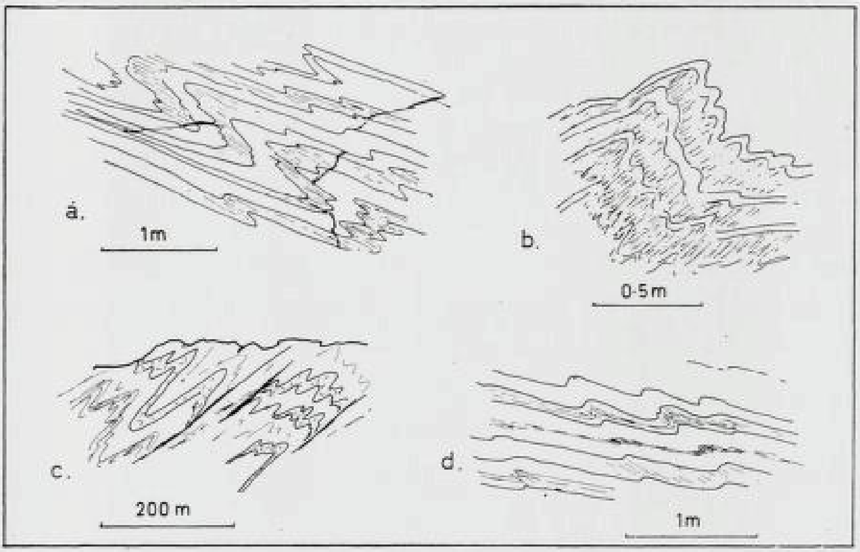

Fig. 37. F_2 structures, North and Central area. (a) Minor folds in mixed pelite-psammite, 800 m E. of Skarvfjord. (b) Minor folds, N.W. of lake 145. (c) Mesoscopic F_2 folds on cliff-face of W. Finfjord, showing granite gneiss sheets and major shear planes. (d) Minor folds in psammite, 300 m N. of lake 145.

are represented in the massive Quartzite 2 east of Lake 250. These can be traced northwards into folds with one inverted limb nearer the closure zone of the Langstrand Antiform. To the south these folds die out.

(d) *The Vest-strømmen – Veirbukten area.*

This is the region in the south-west, comprising the wide outcrop of Lower Semi-pelite together with quartzites of a prevalent easterly strike. Structurally it is the least consistent and conformable sub-area, despite sedimentological evidence indicating a regular younging to the south, without repetition of members.

Variations in F_2 fold attitude inherent over this sub-area are depicted on the stereographic plots (Fig. 38). Of particular significance is the spread of axial trends, with a maximum concentration between 060° and 090°, a lesser concentration around 180°–190° and a scattering of points between 190° and 250°. This is partly a reflection of the strike swing and partly, and perhaps more important, due to axial trend varying in relation to the direction of fold overturning.

It can be seen that, whilst folds overturned to the S.S.E. (with N.N.W.-inclined axial planes) in general plunge towards 060–065°, a larger number of F_2 minor folds in this area are overturned towards the N.N.W. with axes plunging in the 085°–090° direction (Fig. 38a). Thus, two systems of F_2 folds are present with opposed senses of overturning and diverging axial trends. The two systems are complementary and conjugate (Fig. 39). A third system of F_2 minor folds is represented by those which have N-S trending axes and W or W.N.W. dipping axial planes. This system of folds is largely restricted to the north and east parts of the sub-area where strikes are inconstant and roughly northerly and where the complementary set is absent. Moving towards Vest-strøm col and Veirbukten, fold axial direction tends to swing from N-S through S.W. to the 240° (or 060°) trend. At the same time, the N.N.W.-overturned folds appear reaching a maximum development in the col region in the semi-pelitic lithology. It is not strictly correct therefore to refer to the N-S trending folds as a separate system, insofar as they are correlative with the 060° system.

Minor fold styles are much the same in the complementary systems, the degree of overturning dependent largely on lithology. In the massive psammites folds, where developed, are open or monoclinal. Axial

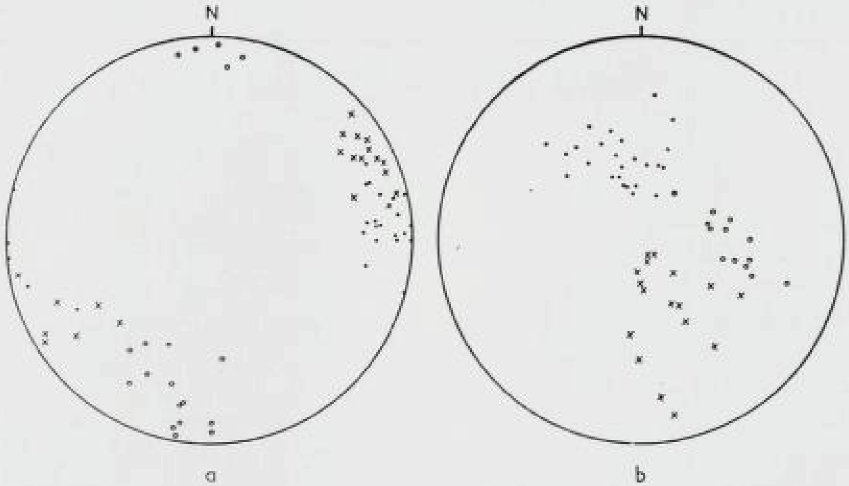

Fig. 38. Structural data (F_2) from the Vest-strømmen–Veirbukten area. (a) Fold axes; crosses represent axes of F_2 folds with N-dipping axial planes; dots – axes of folds with S-dipping axial planes; circles – axes of folds with W-dipping axial planes. (b) Poles to axial planes of the folds in diagram (a).

Fig. 39. Conjugate F_2 minor folds. Lower Semi-pelite, col between Vest-strømmen and Veirbukten.

plane schistosity is nowhere penetrative but may be present in both N.N.W.- and S.S.E.-overturned folds, moreso where pelitic material predominates. Shearing is often manifest parallel to F_2 axial planes.

It is evident that, on account of the presence of conjugate F_2 folds, the movement picture reflects orthorhombic symmetry, except for the north-easternmost sector wherein monoclinic symmetry is apparent. On studying the geometry of these conjugate structures however, the divergence of fold axial trend implies that the kinematic b-axis is not coincident with the layering. It follows that, geometrically, the conjugate folds have a lower order of symmetry, essentially triclinic and not orthorhombic, even though the movement picture favours the latter.

Recognition of complementary shears and fold pairs in this sub-area is compatible with the work of Ramsay and Sturt (1963) in a similar E-W striking region around Storelv, some 10 km S.W. of Vest-strømmen. In both these areas fold axial trends diverge by some 20°, the differing trends related to opposed axial planar inclinations – basically congruous and incongruous conjugate folds. Appleyard (1965) has also found congruous and incongruous F_2 folds further to the west in E-W-striking rocks at Dønnesfjord.

That conjugate structures have resulted from a shear mechanism is generally accepted. Both fold sets are regarded as having originated during the same F_2 system of stresses, though it is possible that one may have developed slightly later than the other in the protracted F_2 deformation episode. The presence of an orthorhombic shear pattern in the 'E-W strike belt' of Vest-strømmen and of central Sørøy generally, and the predominance of monoclinic symmetry in the 'N-S strike belt' is extremely interesting and will be discussed later along with the major structures.

(e) *The Schist – Limestone belt.*

This narrow tract of limestones, Storelv Schist and Transitional Group is discussed separately on account of its relative lithological incompetency and distinctiveness of fold styles. Dips are generally to the W.S.W. becoming S.S.E. or near-vertical in the south with a marked swing of strike.

F_2 minor folds are most readily observed in the Transitional Group, particularly in the region of steeper dips approximating to the hinge zone of the Langstrand Antiform. In this zone chevron folds and zig-

Fig. 40. F_2 folds in Transitional Group showing development of boudinage and radiate joint pattern. Road-cutting, W. Langstrandfjord.

zag folds (terminology dependent upon whether the limbs are of equal or unequal length respectively — see Fleuty (1964)) are profuse, acuteness of style proportional to the pelite content. Such folds have geometrical properties of both parallel and similar folds. In Fig. 40 several of the quartzite bands are attenuated on the limbs and boudinage-mullions are developed, so that both shear and flexural mechanisms can be considered to have operated in the fold development.

Away from the hinge zone of the Langstrand Antiform the sense of overturning of F_2 minor folds is congruous with respect to the major structure and amplitude-wavelength ratios increase. Northwards from Hønsebyvatn, the broadening of the Storelv Schist outcrop is attributed to an increasing presence of F_2 folds, these being of similar type with appreciable thickening at fold apices and a marked development of S_3. The local deflection of fold axial trend through some 40° is noteworthy (Fig. 41a), partly due to cross-folding though principally a consequence of the regional tendency for diverging 'b' axes in pelite lithologies.

The limestones display F_2 folds in the south of the sub-area but towards the north folds are rarely seen. These folds vary in style, this

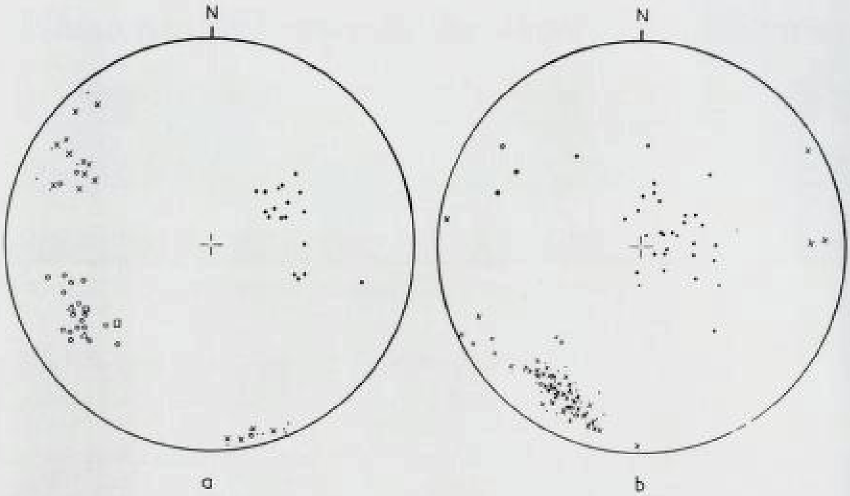

Fig. 41. Structural data (F_2) from the schist-limestone belt. (a) Northern area (b) Southern area. Crosses represent F_2 fold axes; small dots – S_1/S_2 intersection and mineral lineation; circles – boudinage axes; squares – cross folds; large dots – poles to F_2 axial planes; triangles – lineation of striae in slide zones.

dependent largely on the nature and degree of impurity of the limestone, though they rarely exhibit the acuteness of style as seen in the schists and banded psammite-pelite sequence. F_2 minor folds are present in Fig. 15. In the northern part of the sub-area, boudinage and minor sliding of F_2 age is present to the near exclusion of actual folds, these latter being represented only by microfolds in calc-silicate schists.

Cross folds.

Tectonic 'b' axes and lineations nowhere exhibit perfect constancy of direction and plunge. While en échelon folding represents one cause of this variation, in several localities broad cross-fold warps constitute an important element in the structural picture. Cross folds are demonstrable to maximum effect in the east Sandøfjord region: fold wavelengths vary appreciably, from several metres to 100–200 metres, the larger structures distinguishable more as general culminations and depressions in the plunge of F_2 'b' axes and lineations.

The presence of cross folds results in local strike swings of up to 80° and variations of 'b' linear trends of as much as 24° (near Skarvfjord). Such warps are often observed in association with conjugate shear planes which may sometimes be unequally developed. When the conjugate shear intersections are plotted on a stereographic net, they are found to correspond almost perfectly with the measured cross fold axes and also with a prevalent boudin orientation (see e. g., Fig 35 f). A close relationship would therefore appear to exist between shearing, boudinage in 'a' and cross folding during the F_2 tectonic episode. Moreover, since these various minor structures are essentially a consequence of the main folding, the inference is that cross folds have developed during a later phase of the F_2 deformation.

In the south of the area where an E.N.E.–W.S.W. strike prevails cross folds are rarely observed: they tend to increase in development towards the area of intense shearing and stretching in the north-west, so that a relationship would appear to exist between sinuosity of F_2 B axes – viz, cross fold warping – and intensity of deformation. This may have been effected by uneven and irregularly distributed tectonic transport, here towards the E.N.E. Working partly in conjunction with this hypothesis, a variability of tension orthogonal to the major compressive stresses would necessitate the localization of complementary stresses parallel to 'b', so producing warps in the 'a' direction. Since a shearing mechanism has clearly been associated with the production of the cross folds in the Skarvfjord area, a movement picture involving both shear and tensional features must be envisaged as having been operative.

Shear planes and Slides.

The complementary shear planes of the Skarvfjordhamn area and those associated with the conjugate folding are the most outstanding examples of this minor structure. Shearing is also developed, often considerably, parallel or sub-parallel to the axial planes of F_2 folds of varying magnitude, and amphibolite sills are quite vulnerable to disruption by such shears (Fig. 42). The inverted limbs of folds may sometimes be completely destroyed; in other cases incipient shears may be developed paralleling F_2 axial planes (Fig. 43).

Difficulty occasionally arises in distinguishing between shears and slides where one seemingly grades into the other, and some minor slides are probably the result of continued and lengthy shearing. Minor sliding is not uncommon in the limestones, moreso in the Lower Limestone.

Fig. 42. F_2 fold and axial planar shear with disrupted amphibolite sill. Klubben Quartzite, small cove 600 m N.W. of Hønsebyfjord.

Fig. 43. Garnetiferous amphibolite squeezed along F_2 axial planar fractures and shears. Pale clots within the dark amphibolite are garnet crystals. Grasdalsnes.

Fig. 44. Minor slide and associated pegmatite. Note changing joint pattern around F_2 fold (above the slide). Psammite, ca. 1 km S. of Hønsebynes.

A 20–60 cm thick cryptocrystalline slide zone, noted earlier as occurring in this limestone, displays conspicuous slickenside-like striae which trend parallel or sub-parallel to the main direction of F_2 tectonic transport. Locally a 3–5 cm thick brecciated band is present along the margin of this slide zone. Small-scale sliding is much in evidence along the south-west Finfjord shore and it is not improbable that a slide of larger dimensions occurs in the Falkenes Limestone Group (largely hidden by scree, or submerged) of the Finfjord district.

Although, as far as can be detected, most of the observed slides appear to have been generated during the F_2 tectonism, it is probable that some sliding may have accompanied the F_1 folding, particularly in the limestones, only to be masked by the later pervasive recrystallisation.

Linear structures.

Several pervasive lineations form an integral part of the F_2 structural picture. Fold axes, phyllitic lineations and intersecting S-surfaces have been discussed earlier; under this present heading a résumé is given of mineral lineation, boudinage, mullions, quartz rodding, elongate grain aggregates and slickensides occurring in this area of Sørøy.

(a) *Mineral lineation.*

A linear preferred orientation of minerals parallel to the axes of F_2 folds is common throughout this area, though not developed in all lithologies. The minerals generally exhibiting preferred orientation are amphiboles, tourmaline and sillimanite of which only the amphiboles are clearly visible in the field.

Acicular or long prismatic crystals of amphibole (hornblende or actinolitic hornblende) parallel the axes of F_2 folds in the Hellefjord Schist, the crystallographic 'c' axes being regularly orientated in the tectonic 'b' direction. A similar hornblende lineation is well developed in the amphibolites and metagabbroic sheets.

Sillimanite occurs only sporadically in schists of the Falkenes Limestone Group and then as bundles, knots or trains of fibres aligned distinctly parallel to the tectonic 'b' direction. On a microscopic scale tourmaline shows a marked preference for alignment in the F_2 axial direction. Elongate garnets occur infrequently .These stretched garnets are similar to those described by Wilson (1961) from Scottland, though lacking the quartz-filled transverse fractures displayed by the Scottish examples, but should not be confused with trains of fragmentary garnet, also evincing a linear arrangement, developed by more extreme F_2 deformation.

(b) *Boudinage.*

Boudins are developed by the tectonic disruption or lateral extension of competent layers in a yielding incompetent host-rock. They are widespread in the banded psammites and limestones exhibiting a variety of shapes dependent on lithology, environment and genesis. Typically, the boudins are sausage-, lozenge- or barrel-shaped in cross-section (Fig. 45 and 46), though often pillow-like when viewed in three dimensions. While the elongation of boudins is generally parallel to the local F_2 axial trend, this orientation in 'b' is sometimes complemented by boudinage axes approximately normal to this direction. Thus, boudinage is present showing two orthogonal linear elements developed during the same deformation episode (see the various stereographic plots and the map, Plate II): the mechanism of production of simultaneous boudinage axes has been discussed by Flinn (1962) and another example from Sørøy mentioned by Sturt (1962).

Quartzite bands in a semi-pelitic lithology frequently show the initial stages of boudinage, producing a thinning or 'necking' (Rast 1956).

Fig. 45. Boudins of calc-silicate bands in limestone. West Finfjorddalen scarp.

Fig. 46. Boudinage in Transitional Group. West Lunddalen.

Fig. 47. Rotated boudins of phyllite within limestone. Langstrand.

Where boudinage is fully developed, recrystallised quartz or calcite is invariably present at the nodes.

Movements along rotational shear joints resulting from the imposition of a shearing couple are envisaged as responsible for the pattern illustrated in Fig. 47. On a small scale however, purely tensional phenomena are manifest within the rotated phyllite segments. In the Skarvfjord region boudinage in granite gneiss is observed in both F_2 'a' and 'b' directions, the 'a'-orientated boudins being closely associated with conjugate shear planes or one strongly developed shear plane (Fig. 33f). Minor boudinage is sometimes visible within larger boudins of granite gneiss. In the case of this particular shear-controlled boudinage, although the two boudin axes are broadly coeval it would appear that those in 'a', controlled by the conjugate shears, are slighter later in development than the 'b' directed boudins.

Within the Falkenes Limestone Group, boudinage similar to Weg-

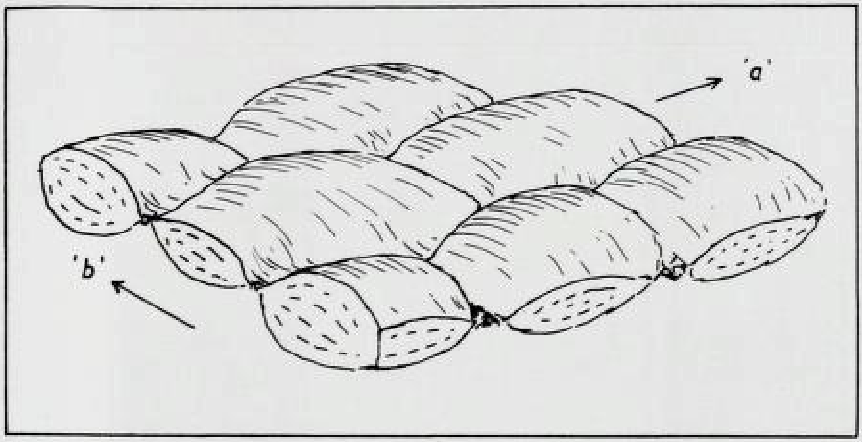

Fig. 48. En échelon arrangement of boudins; mineral orientation in F_2 'b'. Tremolite – diopside schist, Falkenes Limestone Group, 600 m N. of Hønsebyvatn.

mann's (1932) "chocolate tablet structure" may be observed. A difference is apparent, however, in that here the pillow-shaped segments are clearly arranged en échelon (Fig. 48). In such cases only one true linear element is present, here in 'a'; the observed mineral lineation in these calc-silicate schists is parallel to 'b'. It is noteworthy that the 'a' boudinage in this lithology is largely confined to zones wherein sliding has been widespread – the striae noted along such slide surfaces also approximate to the tectonic 'a' direction.

Isolated bodies derived primarily by a boudinage mechanism but involving extreme separation of the segments have been termed tectonic inclusions. The amphibolite lenses within the Hellefjord Schist are large-scale examples of this phenomenon – they are thought to represent tectonically transported severed segments of the Storelv Metagabbro. An elongation paralleling the local F_2 'b' direction is always noticeable. Other examples of tectonic inclusions include some of the lenticular pegmatitic bodies and calc-silicate schist pods within the Storelv Schist. In areas of advanced deformation where interbanded psammite and pelite is present, the quartzitic layers on account of the extreme shearing often lose their continuity, ultimately being represented by isolated tectonic inclusions which may be recognisable fold closures (Fig. 49).

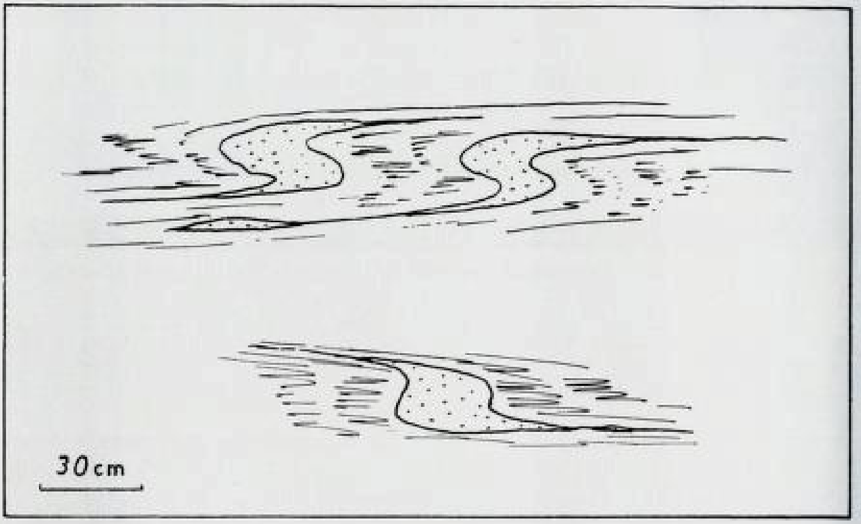

Fig. 49. F_2 fold closures (in quartzite bands) preserved in a strongly sheared semi-pelite. South Skarvfjordhamn.

(c) *Mullions.*

Mullion structures are an uncommon linear element in north-east Sørøy but where present afford additional evidence of the movement picture during the F_2 deformation. Though generally associated with folding, mullions have been classified by Wilson (1953) on the basis of morphogenetic character.

The Hellefjord Synform closure region of near-vertical dips provides excellent examples of cleavage-mullions (Fig. 50 and 51), these having been formed by the prismatic parting of the rock along intersecting cleavage and bedding surfaces so producing the grooved pattern. Mullion lineation coincides with the F_2 'b' direction; moreover the prominence of transverse joints is noteworthy. In the Hønseby–Skarvfjord area both fold- and ruler-mullions occur in psammites.

(d) *Quartz rodding.*

These are elongate, sheared out and partially rolled quartz segregations and veinlets, the axes of the rods again coincident with local F_2 fold axes. All manner of gradations from boudined quartz veins to cylindrical rods may be observed, mainly within the Hellefjord Schist lithology.

Fig. 50. Cleavage-mullions in Hellefjord Schist. Hellefjord Synform closure zone, N.W. of Hellefjord.

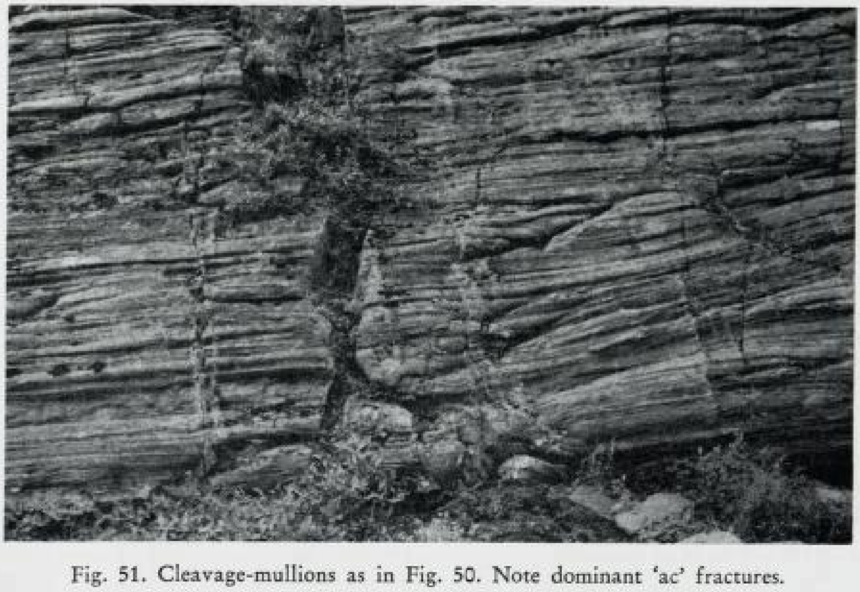

Fig. 51. Cleavage-mullions as in Fig. 50. Note dominant 'ac' fractures.

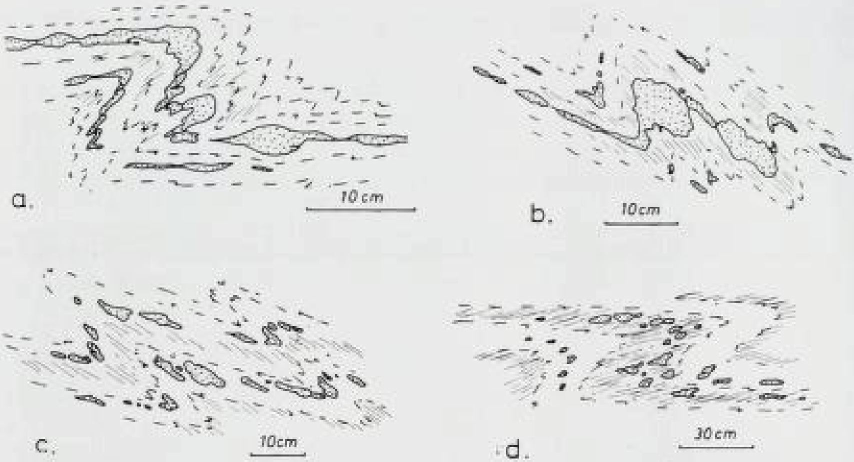

Fig. 52. Field sketches showing the development of quartz rods in Hellefjord Schist. In (a), quartz segregations are boudined; with more noticeable F_2 axial planar shearing the quartz is broken up into detached fragments with an ultimate cross-sectional form as seen in (d).

Fig. 53. Lineation of quartz-feldspar on surface of biotite-hornblende granitic gneiss. 800 m N. of Hellefjord.

Quartz veinlets lying in the banding are deformed to varying degrees by F_2 folds, depending on the extent of shear folding. The ultimate product, long separate rods of quartz roughly oval, discoidal or circular in cross-section present a pronounced lineation, although this perfected state is not so common as the intermediate structures. Stages of quartz rod development as typical of the Hellefjord Schist are depicted in Fig. 52.

A regular structural feature of the rods is the close fracturing transverse to their elongation, essentially 'ac' joints. Where rodding is better developed, fracturing is likewise more conspicuous and oblique shear joints may occur. Rodded quartz is sometimes accompanied by recrystallised calcite and occasionally by diopside. Along West Skippernesfjord elongate rods or lenticles of calcite (with subordinate quartz) are not uncommon, axes again trending parallel to the tectonic 'b' direction.

Elongate grain aggregates are quite prominent in the biotite-hornblende granitic gneiss occurring within the Hellefjord Schist sequence. These are seen as elongate clots of quartz-plagioclase or quartz-microcline-plagioclase (Fig. 53). The Gamnes slump lithology on account of its tectonic deformation now displays thin rod- or spindle-shaped aggregates of equant quartz and microcline grains.

(e) *Slickensides.*

Slickenside striae, unlike the preceding linear elements, are not pervasive; they occur on discrete surfaces of discontinuous movement associated with faults, slides or bedding-plane slip. Those observed in the Langstrand–Finfjord area are related to the F_2 tectonism. Along the Skarvdal Fault two directions of slickenside striae are demonstrable. Striae along slide surfaces are quite significant features – they differ from the previously described lineations in that their trend is parallel to the direction of tectonic transport and not normal to it.

C. Late F_2 structures.

While there is no emphatic evidence of a distinctive third fold episode, late microfolds noted in four thin-sections deform S_3 and are considered as representative of a later brittle phase of the F_2 generation of folding. Associated features include the flexing of biotites through 90°, fractured feldspars and the granulation of minerals along micro-shear planes. Fault zones often exhibit minor and microfolds with appreciable shearing and cataclastic features. Diaphthoretic phenomena are abundantly associated with this late brittle phase of deformation, the various facets of which are demonstrably related to the F_2 movement picture and stress system.

2. Faults

Of no mean importance in a tectonic synthesis of this part of Sørøy are the faults and pattern of faulting, clearly defined trends of which are immediately recognisable (Plate II, Fig. 54). As full details of the faulting are given elsewhere (Roberts 1965), only the salient features will be discussed here.

Although the great majority of faults are dip faults or faults showing small angles of obliquity to dip, strike faults are demonstrable around Hellefjord. Faulting is, almost without exception, normal, the hanging wall having been displaced downwards relative to the foot wall, and frictional drag features simulating monoclinal folds are not incommon. Evidence of near-horizontal movement along a fault plane has been observed on the Skarvdal Fault in the Veiviken area, in the form of slickensides plunging 10° to the S.W. A dextral movement is denoted by stepped features on the slickensided surface. Vertical slickenside striae are also prominent along this same fault plane and considering the various evidence it would appear that vertical displacement has been of greater magnitude and significance than horizontal movement. Two separate movements along this particular fault are postulated – differing responses to changing stress conditions.

Zones of brecciation are well developed along several fault planes, more especially the Skarval Fault and faults in the vicinity of Hellefjord. A breccia zone up to 1 metre wide associated with the Skarvdal Fault is best exposed along the north Skarvfjord shore; an adjacent 'shatter-zone' of closely spaced fractures parallels the fault plane. Further to the north-east, above Veiviken, this same fault affects flaggy Hellefjord Schist, the breccia there being more crumbly and clayey often resembling cemented rubble and rock flour.

The Hellefjord Fault and associated parallel strike faults display brecciated, and partially mylonitic, zones up to 1.5 m in width, as well as rather prominent frictional drag features. All these faults are normal, hading from near-vertical to 25° to either S.W. or S.E. Down-

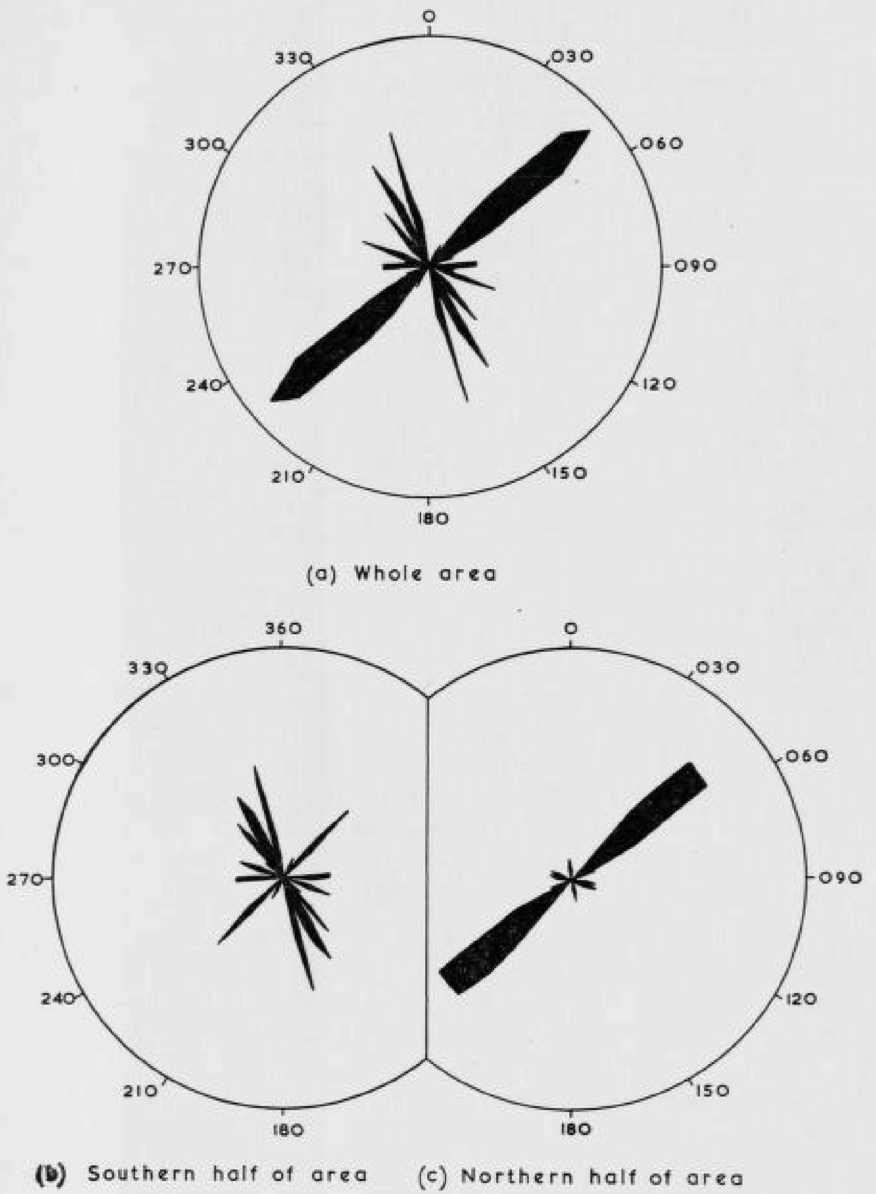

Fig. 54. Rose diagrams of faults – to nearest 5°.

throw values can rarely be measured accurately. In this area the Hellefjord Fault throws down some 45–50 m to the W.S.W. The Skarvdal Fault has an estimated throw of ca. 80 m (to the north) near Finfjord.

An examination of the pattern of faulting reveals notable trends of fault plane strike (Fig. 54). In the northern half of the area (Fig. 54c) a pronounced fault strike maximum of ca. 050° is evident with minor trends of 108° and 170°. Considering the F_2 'b' lineation maximum of ca. 170°, an explanation of the fault pattern can be formulated on the basis of the direction of tectonic transport having been towards ca. 080°. This direction bisects the angle between the 050° and 108° faults, so that it is possible to regard these as a conjugate set of first-order shears one of which (the 050° trend) has been greatly developed at the expense of the other.

The southern half of the area shows a more complicated fault pattern (Fig. 54b), the complexity of which is largely due to the swing of strike. Thus dip faults vary appreciably in trend. Three main maxima (165°, 153° and 045°) with other lesser trends are manifest, but it is difficult to resolve these into any definite stress field – not unexpected considering the variable strike and F_2 lineation directions. Over the whole area (Fig. 54a), major trends of 045°–055° and 140°–165° are oustanding with the N.E. frequency particularly strong; the absence of faults in the 170°–035° sector is noteworthy. It is of further interest to note that in the west Sandøfjord area in the 'E-W strike belt', normal faults trending 160°–170° are common (B. A. Sturt, personal communication). This accords well with observations of faults in the E-W strike sector of the Langstrand area.

On relating the pattern of faulting to a 'movement picture' it seems probable that the majority of faults can be resolved into a stress field concomitant with a waning of the F_2 tectonism. While the faulting clearly postdates the second generation folds it has developed largely in response to the compressional stresses which produced the F_2 folds: it is a brittle phenomenon which occurred after the rocks had attained a certain rigidity ensuing the relatively plastic folding episode.

3. Joints

As with faulting only the more significant features of jointing, insofar as they are of relevance to the main structural picture, are considered here. Rather than try to assess the joint pattern of the whole

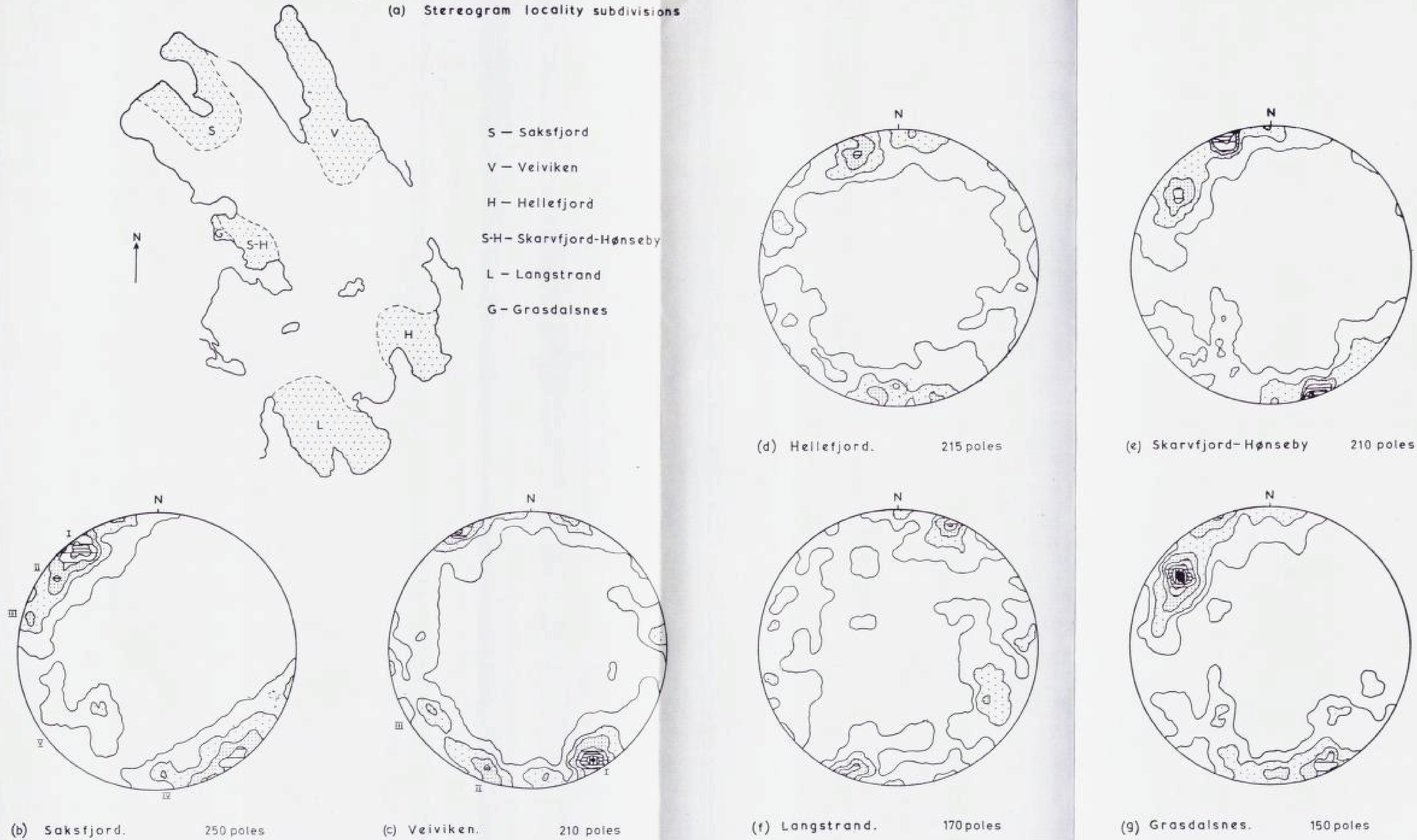

Fig. 55. Stereographic projections of jointing data from selected sub-areas. Contour intervals, 1, 3, 5, 7, 9, 11, 13 %. Schmidt net, lower hemisphere.

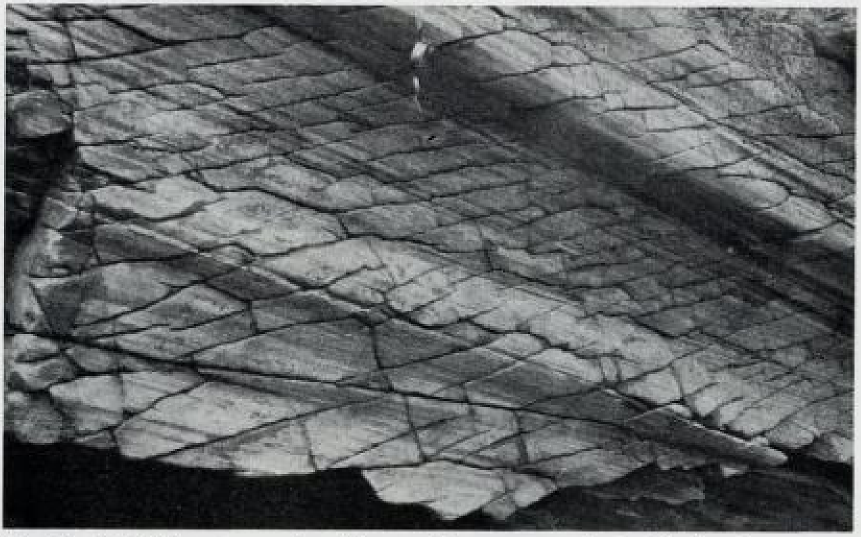

Fig. 56. F_2 'b' lineation and mullions with transverse, longitudinal and conjugate shear joints. Psammite, N.W. Hønsebyfjord.

being 'ac' or cross joints most probably of tensional origin. Maxima III and IV on this diagram represent trends of 015° and 085°, the acute bisectrix of which approximates to the major 052° cross joint (maximum I), so that one can regard these oblique sets as shear joints related to a N.E.-directed compressive stress. Maximum V, a 134° trend, is essentially longitudinal in relation to the F_2 tectonic pattern. Since this joint set hades at 40°–44° to the N.E., more or less normal to the average dip of banding in this Saksfjord area, the assumption is that this NW-SE joint is of tensional origin due to an elastic release of the main deformative stress.

Similar joint patterns prevail in the other sub-areas, all bearing some relationship to the F_2 structures. From the various evidence it would seem that both shear and tension joints have developed in response to the stresses imposed during this orogeny, such that the jointing can be conceived as occurring later in the F_2 tectonic episode. Tensile and shearing stresses, and not the initial compressive stress, are thus directly responsible for this jointing.

Around Hellefjord, late joints cut the partially silicified fault-breccias. Elsewhere, jointing within breccia fragments appears to be totally unrelated to – essentially earlier than – the faulting, but it is area, selected sub-areas have been chosen (Fig. 55) and an endeavour made to relate the patterns to the localised stress fields and general movement picture.

Both shear and tension joints are recognisable although the precise origin of many joints is difficult to ascertain. Conjugate joint sets are commonly present in the psammitic lithologies and in the flaggy Hellefjord Schist. Fig. 56 shows such a pattern in which one of the oblique joint sets is strongly developed in comparison with the other; the pronounced lineation and mullion-like structure – the F_2 'b' lineation – approximately bisects the angle between the two sets of joints. In addition, a longitudinal joint set is present paralleling the 'b' direction and (to the left of the photograph) a prominent 'ac' or cross joint is also visible. Thus, two systems of joints are apparent, the first comprising cross and longitudinal joints and regarded as being tensional fractures and the second embracing the conjugate oblique joint sets which are considered to be of shear origin. In the Langstrand area such systems of jointing have produced an hexagonal (almost columnar) pattern in massive quartzites.

Joint surfaces are rarely smooth, commonly revealing wrinkles and rugosities. In the Hellefjord Schist several examples of joint surfaces having plumose or feathery markings have been observed; these are termed 'feather-fractures' – not to be confused with the feather-joint of Cloos (1932) – and are apparently restricted to shear joints (Roberts 1961). The fine grain, homogeneity and compact nature of certain bands in the Hellefjord Schist are factors favourable for the development of this plumose joint structure.

Quartz infilling of joints is not uncommon in the Klubben Quartzite Group. In some instances tensional gash veinlets of quartz are prominent features. Within the Hellefjord Schist lithology joint surfaces are often coated with zeolite which may sometimes display radial acicular form, while in other cases scapolite- and quartz-infilled joints are observed .Locally calcite has penetrated joint-planes, generally as thin films, but in rare occurrences veins up to 8–10 cm thick have utilised these fractures.

Detailed accounts of the sub-area joint patterns are not possible here but it is useful to consider just one of the contoured diagrams in relation to the local stress field and corresponding F_2 linear elements. Taking the Saksfjord sub-area (Fig. 55b), maxima I and II can be related near-perfectly to the F_2 'b' maxima (Fig. 35 a), the joints clearly speculative to suggest a relationship with the F_1 deformation. It seems likely and logical, however, that some jointing developed prior to the F_2 faulting since, at the incipient stage, faults are joint planes. The probability is that the dominant joints sets have developed over a period of time, rather than at one definite stage, during the protracted F_2 deformation episode.

4. The major structures

Both major periods of folding discussed in the foregoing account of the minor structures are represented on a larger scale by overturned and recumbent folds of considerable magnitude. Dominating the structural picture in the southern half of the area is the Langstrand Antiform. This is a major second generation fold which has an inverted eastern limb. It can be demonstrated, however, that the earlier F_1 episode of deformation has had the greater influence on the present disposition of lithologies, and accounts for much of the extensive stratigraphical inversion over the area.

In this respect, sedimentary structures, where present, provide emphatic evidence of the direction of younging and have been invaluable in formulating the regional tectonic picture. In the northern part of the area the regional structure can be explained in terms of the refolding of a major early fold of unknown amplitude.

Account also has to be taken of 'tectonic level' so that a three-dimensional appreciation of the large-scale structure is essential. The magnitude of the folds is such that, in certain instances, a truly clear picture can only be gained by reference to the major structures present in other parts of Sørøy. This is particularly applicable when the changing styles and symmetry of folds related to strike direction are considered. The 'E.-W. strike belt', for example, is only partially represented on the writer's area and here it is necessary to refer to the geology of Central and S.W. Sørøy for a comprehensive interpretation of the regional structures.

Three major structures are present in this area, these being termed the Hønseby Fold, Langstrand Antiform and Hellefjord Synform. Of these, only the Hønseby Fold is an F_1 generation structure, the others being of F_2 age. Apart from these folds reference will be made to the F_1 Skarvfjord Synform and the complementary Skarv Antiform. In discussing the major structures a system of individual fold descriptions is adopted, although in most cases it is necessary to disgress in order to consider genetic implications and fold relationships. The structural aspects of the northern area lend themselves more easily to a generalised description, with the stratigraphical evidence here being vital in constructing the pattern of folding.

Throughout the chapter it will be necessary to consult the many geological sections drawn across the area, together with the structural map (Plate II) and stereographic plots of minor structural observations.

A. The southern area.

The northern limit of this half of the area is taken as an E-W line through Hønsebyfjord.

The Langstrand Antiform.

Dominating the structure of this southern area the Langstrand Antiform is responsible for the inverted stratigraphy east of the fold axial trace (see Plate II). West and south-west of the axial trace, strata are the right way up – current bedding evidence in quartzites – and dips are generally to the south-east.

The axial trace of the antiform is taken as the narrow zone of vertical or near-vertical dips, since the banding on the limbs dips at much the same angle in opposed directions. However, the axial plane of the antiform is not horizontal but dips gently in a S.W. to W. direction; strictly speaking therefore the axial trace should not be drawn through perfectly vertical strata, but through strata with a steep easterly dip. Because of the variability of dip values and the development of a more acute fold style towards the north, the zone of near-vertical dips represents an approximation to the axial trace of the Langstrand Antiform.

From Langstrand, where the antiformal axial trace is in Hellefjord Schist, the 'vertical zone' runs northwards traversing the succession. It follows the ridge of Upper Semi-pelite before swinging to the W.N.-W. on the south side of Hønsebydalen. This marked swing is largely a consequence of erosional features, in particular the high cliff of South Hønsebyfjord and Hønsebydalen which naturally produces the sharp trace swing of the shallowly inclined axial plane.

An examination of the F_2 microfolds and minor folds on either limb of the Langstrand Antiform indicates a perfect congruity of overturning with respect to the major fold. It should be noted that the F_2 conjugate folds of the 'E-W strike belt' are a special case and do not invalidate the observations demonstrable elsewhere. This evidence of congruous minor folds, together with the observed refolding (by F_2 structures) of tight early folds in the limestone and calc-silicate schists towards Langstrand strongly suggests an F_2 age for the Langstrand Antiform. Substantiating evidence is provided by the schistose amphibolite dykes and transgressive sills of proven post-F_1, pre-F_2 age, the schistosity of which is S_3, axial planar to the F_2 folds. The 3 m thick amphibolite sill unbroken for 4 kilometres in the Quartzite of 3 of this southern area is a good example of this; it exhibits a conspicuous S_3 schistosity and is clearly folded by the Langstrand Antiform. No post-F_2 sills or dykes have been observed.

The Hellefjord Synform.

The Hellefjord Synform is traceable from the Lundhavn area northwards to Otervikfjell. Further north it can be shown, on minor structural evidence, to extend to the Finfjord region and beyond the Skarvdal Fault the axial trace is found again near Veiviken. The fold displays perceptible changes of style although the axial plane is inclined constantly to the W or WSW. The eastern limb dips gently to the S.W., while the western limb shows steeper dips to the W. Further east, in and beyond the Skippernesfjord area, dips become sub-horizontal.

Everywhere, abundant minor F_2 folds exhibit a congruous relationship to the major synform, axial planes and the associated S_3 schistosity dipping gently in a W.S.W. direction. F_2 lineations are abundant, paralleling the fold axes. In a few localities, tight early folds refolded by F_2 are present. Clearly the Hellefjord Synform is an F_2 structure and it is noteworthy that the west limb of the synform and the east limb of the Langstrand Antiform are one and the same.

Stratigraphical evidence indicates that the Hellefjord fold is essentially a syncline; the rocks in the limb common to both the Langstrand and Hellefjord folds are inverted. The strata of the extensive eastern limb of the Hellefjord fold are therefore normally disposed. Limestones are noticeably absent from the eastern limb of the synform, and probable reasons for this non-repetition of the Falkenes Limestone Group are twofold:

(i) It can be noted that, owing to the near-horizontal nature of the normal limb together with an abundance of microscopic and mesoscopic F_2 folds (all overturned to the E.N.E.), it would appear unlikely that the limestone could crop out on the present-day erosion surface, other than by faulting. As yet no strata older than the Hellefjord Schist have been found in the extreme N.E. of Sørøy.

(ii) The absence of an eastward repetition of the Falkenes Limestone Group is possibly due to the presence of the major F_1 isocline, the Hønseby Fold. This, noted in the reconstructed profile (Fig. 57), occurs beneath the normal limb of the Hellefjord Synform. The presence of isolate, minor F_1 folds in the banding is a pointer to the proximity of a larger structure. Hence, the limestones could be envisaged as wedging out eastwards in the core of the tight F_1 fold beneath the thick pile of Hellefjord Schist.

Comparing the Langstrand and Hellefjord folds in this southern area, a difference of style is noticeable despite the compatible axial planar altitudes. The Hellefjord Synform is invariably more acute than the Langstrand Antiform although both the actual major fold closures have large radii of curvature. Accepting the thesis of similar age for these structures, the greater angular disparity between limbs in the case of the antiform is explicable as another example of lithological control. Whereas the antiform is almost wholly within massive quartzites, the Hellefjord Synform affects flaggy, relatively incompetent schists. Another factor influencing the style of folds appears to be the position of folds relative to the E.-W. and N.-S. strike belts. This affects both the Hellefjord and Langstrand folds. The E.-W. belt generally contains F_2 folds of less acute style than in the more strongly deformed N-S belt. When viewed on a regional scale a relationship between F_2 fold style and strike belts is more clearly apparent.

While the axial trace of the Langstrand Antiform transgresses various members of the stratigraphical succession, that of the Hellefjord Synform shows a general concordance with strike, transgressive only north of Hønsebyvatn. This can be explained in terms of observations of F_2 minor fold axes, 'b' lineations and axial planes which reflect to a large extent the attitudes of the bigger structures. With the Hellefjord Synform all 'b' linear structures approximate to the horizontal: therefore, on the plateau surface the axial trace parallels the banding. Further north, F_2 fold axes and lineations plunge at $12°–16°$ to the S.S.W. (see stereograms, Fig. 22) and the average strike of the axial planes changes accordingly. Consequently the erosion surface is such as to cause the axial trace of the synform to transgress the massive gneiss of Hønsebyvatn; this is seen in the cross-profiles (Plate III).

Structural observations from the Langstrand Antiform show that axes and lineations plunge, on average, 20°–22° to the S.-S.S.E. with local plunges above 30° in the south. Axial planes are generally more steeply inclined than those of the Hellefjord Synform due to the minor folds having a less acute style in the psammite. The transgressive nature of the trace of the major axial plane can thus be accounted for, since the general erosion surface cuts deeper into the antiformal core towards the north.

The closed outcrop pattern of the massive gneiss on Otervikfjell is shown to be a topographic feature, as the Hellefjord Synform axial trace is there west of the gneiss outcrop. Topography, together with near-horizontal dips also accounts for the faulted outlier of gneiss capping the ridges of Finfjordfjell. In this general Finfjordfjell area of Hellefjord Schist, minor F_2 folds show a constant overturning to the E.N.E. This is noticeable as far north as the Skarvdal Fault. S.S.E. from Veiviken a prominent F_2 fold of wavelength ca. 100 m and amplitude 50–70 cm can be traced for some 4–5 km. It is clear therefore that this is a continuation of the normal limb of the Hellefjord Synform, the minor folds being congruous to the major structure.

The Hønseby Fold.

On the upper limb of the Langstrand Antiform, normally disposed strata can be followed towards the Hønseby area with the aid of sedimentary structures, until the Hønseby Fold is encountered. This latter fold is a tight, near-isoclinal fold closing downwards, the axial plane of which is parallel or sub-parallel to the general banding. The fold, here a synform, is recognisable in cross-section only on the south cliff coast of Hønsebyfjord. Without the presence of current bedding, accurate mapping of the fold axial trace, would have been impossible since, on the plateau of Hønsebyfjell, only two exposures of the actual fold closure give any indication of the location of the axial zone.

Towards the west, this tight Hønseby Fold is refolded by later structures. These are F_2 folds of intermediate size, all overturned eastwards. In the extreme west along the Sandøfjord coast, the extension of the Skarvfjord Synform folds both the banding and the early Hønseby Fold. Towards the coast of Hønsebyfjord, the banding and axial plane of the Hønseby Fold steepen rapidly to near-vertical, indicating a

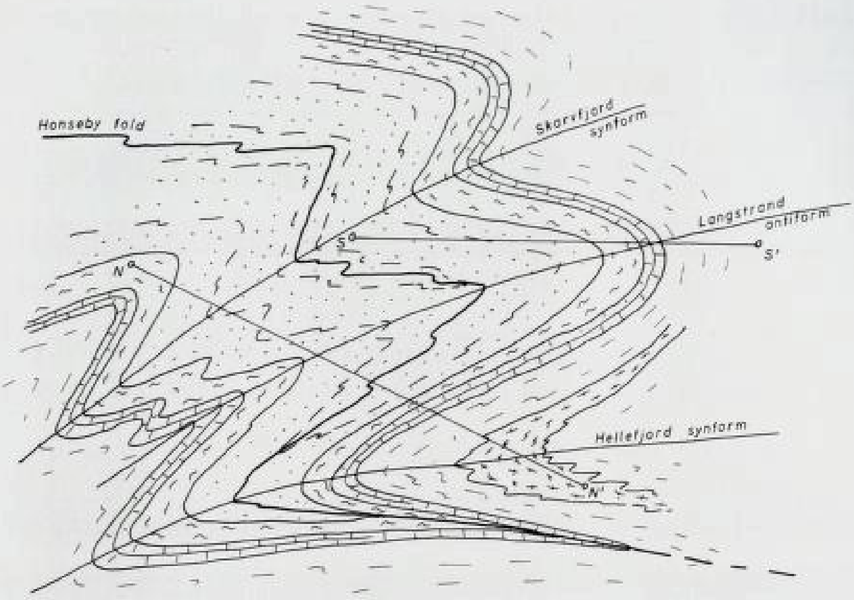

Fig. 57. Diagrammatic reconstruction of the major structures of N.E. Sørøy. N-N', WSW-ENE profile across north of mapped area: S-S', NNW-SSE profile across south of mapped area. Symbols as drawn on main map (Plate I).

proximity to the Langstrand Antiform axial region (see Sections Q-R and S-T, Plate III). The fold is then lost temporarily beneath the fjord.

The cumulative evidence thus clearly indicates that the Hønseby Fold is an early (F_1) structure. It is, in fact, an inverted anticline since older rocks lie in its core. Although it is not possible to comment, in absolute terms, upon the amplitude of the Hønseby Fold, it is none the less certain from various lines of evidence that recumbent folds of very considerable dimensions developed during the F_1 deformation episode. The amplitude of the Hønseby Fold, though indeterminate from the restricted evidence, may well be measureable in tens of kilometres.

The Langstrand Antiform and Hønseby Fold can be shown to extend north of Hønsebyfjord although the evidence in this belt of N-S strikes is less direct, owing to a general lack of sedimentary structures. Utilising the fundamental criterion of the proven stratigraphical sequence, however, it can be demonstrated that these two major folds are present.

From the foregoing discussion of the major structures of the southern half of the surveyed area, a comprehensible picture of major and minor fold age relationships has emerged. Second generation folds, namely the complementary Langstrand Antiform and Hellefjord Synform dominate the surface geology with the axial trace of the early Hønseby Fold detectable only in the north-west. Fold axial and axial planar attitudes of the major F_2 structures differ only slightly but nevertheless significantly. Differences are attributed primarily to lithology.

B. The northern area.

North of a line drawn from Hønseby to Basteidet, strike direction remains fairly regularly S.S.E. or S.E. Dips are also fairly constant towards the W or S.W., with local exceptions near Bastafjord. Northwest of Skarvfjordhamn near-vertical strata constitute the western limb of the Skarvfjord Synform. The predominant overturning of minor F_2 folds towards the E.N.E. with intervening and alternate zones of W.S.W.-overturned folds is extremely important in an understanding of the large-scale structures and deserves some detailed consideration here. At the same time, in interpreting the overall tectonic picture, it is essential to refer constantly to the basic stratigraphy.

The profiles drawn across the northern area show the constant direction of inclination of F_2 axial planes referred to in the minor structural chapter and also depicted in the various stereographic plots. Monoclinic structural symmetry is ubiquitous. North of the Skarvdal Fault, axial traces of several intermediate and larger F_2 folds can be demonstrated (Plate II). Several of these folds have been mapped solely on the S_1/S_3 relationship described earlier, this criterion being of considerable value in such an area of monotonous lithology.

In contrast to the southern area, sedimentary structural evidence of younging is rather localised being restricted to the immediate Hønseby and Skarvfjordhamn areas. Where present this indicates that the second generation folds, including the Skarvfjord Synform, are folding an inverted succession. Towards the east, along the Sakstinderne ridge and in the Storelv Schist and Falkenes Limestone belt of the interior, minor F_2 folds are again affecting an inverted sequence. In this area no sedimentary structures have been observed: the stratigraphical succession, however, is identical to and a continuation of that proven in the south and also near Skarvfjordhamn.

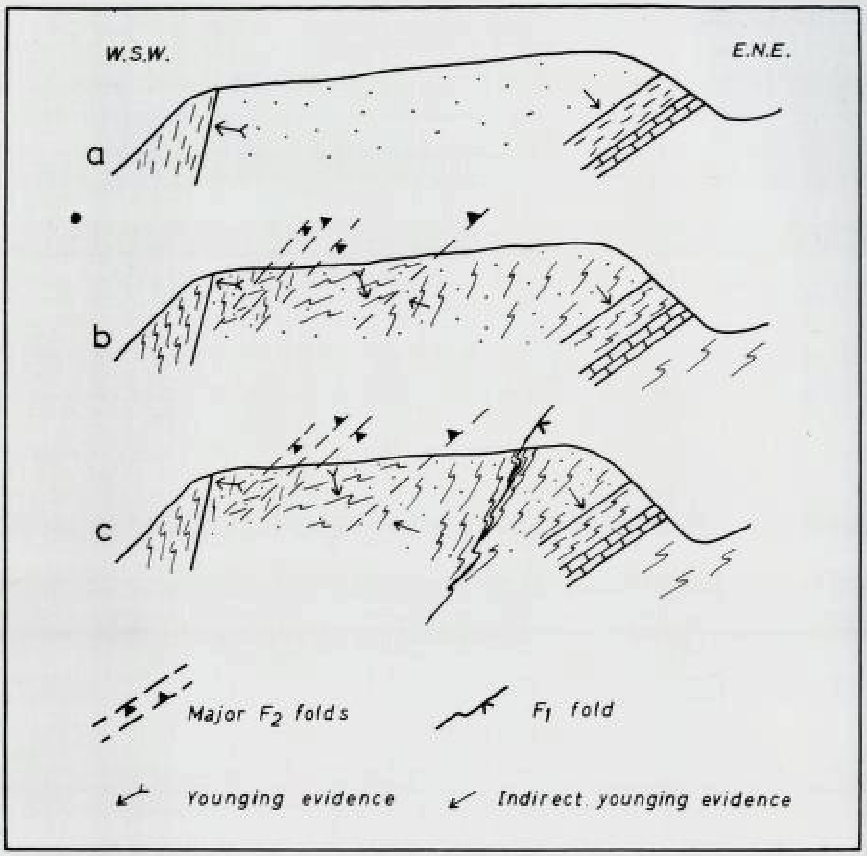

Fig. 58. Diagrammatic profile across northern area showing proof for the presence of an early (F_1) isoclinal fold (further explanation in text).

The similarity between the Sandøfjord and east-central areas in terms of the folding of an inverted succession ends there, however. Taking the Sandøfjord tract first, it can be shown that:

(a) On the steep, western, reverse limb of the Skarvfjord Synform, strata young towards the west.

(b) Minor F_2 folds on this same near-vertical limb are overturned towards the west, congruous to the major synform.

In the case of the east-central area of limestone, schist and highest psammite lithologies, the situation is as follows:

(a) Strata dip W.S.W.–S.W. but young towards the east.
(b) Minor F_2 folds are overturned to the S.W. indicating the reverse limb of a larger fold.

From these observations, it is clear that in the two separate areas, strata on the reverse limbs of major F_2 folds display a younging in opposed directions. This dissidence can only be resolved by invoking the northward extension of the tight Hønseby Fold, as depicted schematically in Fig. 58 and also on the structural map and serial sections. Direct sedimentary structural evidence is lacking north of Hønsebydalen but by extrapolation of the known stratigraphical order, the presence of the F_1 fold (the Hønsby Fold) can be demonstrated.

Tracing the common limb of the Langstrand and Hellefjord folds towards the north, the inversion of the lithologies is attributed to the second generation folding, as can be seen from an examination of the regional profile (Fig. 57). This shows quite clearly that the schist-limestone belt throughout its length from Langstrand to Finfjord, is always on the upper limb of the major F_1 Hønseby Fold, whatever the attitude of the banding. The succession in the north-west however, is located on the lower limb of the early Hønseby Fold. It is this first generation recumbent fold, therefore, which is responsible for the extensive tracts of inverted strata in this northern area. The inversion of the lower limb of the Langstrand Antiform is essentially an F_2 feature.

Area south of the Skarvdal Fault.

As noted previously the swing of the Langstrand Antiform axial trace down towards Hønsebyfjord from the southern plateau is due to a marked erosional feature. On the north side of the fjord and Hønsebydalen the land rises comparatively gently up to the plateau of Skarvfjordfjell.

In considering the plateau area extending north to Skarvdalen, an examination of F_2 minor and larger folds warrants a division into western and eastern sectors based on the direction of fold overturning. The boundary between these two sectors is essentially that of the axial trace of a major F_2 structure. West of this axial trace, F_2 folds constantly face E.N.E. towards an inferred antiformal structure; in the eastern sector, which includes the Storelv Schist and limestones, folds are overturned to the W.S.W. indicative of their location on the reverse limb of the same major F_2 structure. This latter reverse limb has been shown to be the lower limb of the Langstrand Antiform. The inference and conclusion is that the major antiform demonstrable on Skarvfjordfjell is the northward continuation of the Langstrand Antiform. The serial sections portray the structural picture adequately.

A previous discussion of the stratigraphical problems involved showed that on this plateau of constant W.S.W.-dipping strata, the presence of a major F_1 fold must be invoked to explain the downward younging of the upper western limb of the major F_2 antiform (Fig. 58). Direct proof of the exact location of the axial trace of this fold in terms of sedimentary structures is absent in this area. Several other lines of evidence, however, point to its approximate location on the lower limb of the F_2 antiform (Langstrand fold): the rational evidence is as follows:

(i) The Hønseby Fold is known to be refolded by F_2 folds including the Langstrand Antiform.
(ii) On the south side of Hønsebyfjord the synformal Hønseby Fold, there on the upper limb on the Langstrand Antiform, approaches the closure zone of the latter F_2 fold.
(iii) North of Hønsebydalen, at a lower tectonic level, the Hønseby Fold by virtue of its being refolded is now positioned on the lower limb of the Langstrand Antiform.

Therefore, on Skarvfjordfjell, the axial plane trace of the Hønseby Fold must clearly be located east of the axial trace of the Langstrand Antiform. This is illustrated in the serial sections and reconstructed regional profile (Fig. 57).

The gentle plunge of the axis of the Hønseby Fold in the approximate plane of erosion also makes it difficult to pin-point the F_1 axial trace, especially as the fold is of an isoclinal nature. Furthermore, the plateau surface is often a 'felsenmeer' of quartzite blocks due to frost-heaving. Localised anomalous dips in a small area N.N.W. of lake 170 are possibly related to the axial zone of the synform.

In this northern area the Langstrand Antiform shows significant changes of style, the upper limb now dipping to the W.S.W. so that the fold is appreciably more acute. It should be noted that the change from 'easterly' to 'westerly' dips on this upper limb is not so abrupt as first appears; westerly dips are locally present south of Hønsebydalen. Another important consideration concerning the northward increased acuteness of the major F_2 fold, is that the northern area is characterised by N.N.W.-S.S.E. strikes (the 'N-S strike belt'), whilst the southern area is largely composed of S.E.- or S-dipping strata. This is the eastern extremity of the extensive 'E-W strike belt' of central Sørøy. Differences of fold style related to the dominant strike trends have already been discussed, but it is important here to reiterate the observation of more intensive tectonic stretching and acuteness of fold style as being a major feature of the 'N-S strike belt'. Identical variations of style related to strike trend are demonstrable in the Breivikbotn area of S.W. Sørøy (B. A. Sturt, personal communcation).

In the Skarvfjordfjell area, minor F_2 folds show fairly constant plunges towards the S.S.E. in the order of 5–°15°. They reflect the attitude of the major Langstrand Antiform. On the lower limb of the F_2 antiform, fold trends locally deviate to 305° in pelites, with plunges to the N.W. Minor fold axial planes show a regular dip towards 235°–245°, averaging some 30° (see stereogram, Fig. 35), which is slightly less steep than equivalent F_2 axial planes in the southern area.

The progressive development of F_2 folds of increasing magnitude from south to north is an interesting feature of the upper limb of the Langstrand Antiform in this 'N-S strike belt'. This implies either increased fold development with depth or, more feasibly here, increasing fold development away from the 'E-W strike belt'. It will be seen that north of the Skarvdal Fault, intermediate-sized F_2 folds are more extensively developed.

In the zone of appreciable tectonic stretching, towards Sandøfjord, F_2 folds are particularly numerous and mappable with axial plane-banding relationships here confirming location on the upper limb of the Langstrand Antiform. Just to the west of the coastline the presence of the Skarvfjord Synform axial trace can be inferred; vertical strata on the island of Skarvholm are representative of the western limb of this synform (see section I–J, Plate III).

East of the limestone scarp the Hellefjord Schist constitutes the extensive area towards Bastafjord. The Hellefjord Synform is the only major fold affecting this lithology.

Area north of the Skarvdal Fault.

This northernmost area can be conveniently divided into:

1. The plateau of Saksfjell.
2. The peninsula of Finfjordnæringen.

Fig. 59. North Skarvfjordhamn cliff-face showing profile section through the Skarvfjord Synform. View looking almost due north.

(1) *Saksfjell.*

The problems of stratigraphy inherent in this region are identical to those explained for Skarvfjordfjell, since it is only the fault-controlled valley of Skarvdalen which separates the two plateaux. Briefly, the problem of different younging directions in the east and west is solved by extending the Hønseby Fold northwards in the lower limb of the Langstrand Antiform. Plates II and III and figures 57 and 58 illustrate the structures involved.

A major addition to the structural picture is the F_2 Skarvfjord Synform, represented in the cliff-section of North Skarvfjord (Fig. 59). This is seen to fold an inverted succession, proven by a local abundance of sedimentary structures. Again, the extensive stratigraphical inversion can be attributed to the location on the lower limb of the recumbent, F_1, Hønseby Fold. The axial plane trace of the Skarvfjord Synform can be followed from Skarvfjord northwards some 5 km to the Kamøsund coast. South of Skarvfjord, owing to the presence of the major fault, the axial trace is displaced some 400 m to the west and is submerged between Skarvholm and the mainland. It appears again near Hønsebynes.

In the extreme north-west the Skarv Antiform is the complementary structure to the Skarvfjord Synform. This is depicted on sections A-B and C-D, as well as on the reconstructed regional profile (Fig. 57). Between the axial traces of the Skarvfjord Synform and the Langstrand Antiform, two other major F_2 folds are present along with several lesser folds. These two structures, synform and antiform, are also represented on the Skarvfjordtind peak overlooking the south shore of Skarvfjord, but there they are of distinctly smaller amplitude. This is a further example of the increasing development of F_2 folds from south to north.

Everywhere on Saksfjell, F_2 axial planes dip towards 236°–248° at between 30° and 50°. The variations are significant since it can be shown that axial planes are generally steeper in the west than in the east. At the same time reverse limbs of F_2 folds display a distinct shallowing of dip from near-vertical in the west to only 30°–40° in west Finfjord .Taking Finfjordnæringen into consideration too, this trend is continued. There, in Hellefjord Schist and massive gneiss, axial planes dip at 10°–20° towards 236°. This eastward shallowing of axial plane dips is not merely a consequence of the S_3 fanning effect noted elsewhere around F_2 closures. Examining reverse limbs only, in the west F_2 axial planes dip at ca. 43° whereas in the W. Finfjord area the corresponding figure is 28°. Clearly the style of F_2 folds is tending towards increased overturning and, in part, increased acuteness towards the east.

Generally, F_2 fold axes in the Saksfjell area plunge in a S.S.E. direction at about 15°. In the Skarvfjord district, the trend is nearer southerly, averaging 172°. However, an important deviation can be noted in the broad zone of mixed pelite and psammite lithology, on the reverse limb of the Langstrand Antiform along Sakstinderne. There, F_2 axes and related linear structures plunge ca. 12° towards 310°. A similar occurrence is less conspicuous in the schist of the north-west. Although cross folding accounts for certain local variations of fold axial trend, the chief cause of this more significant variation (of ca. 20°) is thought to be the regional tendency for diverging B axes in different lithologies, referred to earlier.

In both the plateau areas of Saksfjell and Skarvfjordfjell, granite gneiss crops out in an irregular zone often more or less coincident with the axial trace of the major F_2 Langstrand Antiform. Although, at first, a relationship between the position of the gneiss and the fold core seems plausible implying a genetic relationship, this can be shown to be a fortuitous situation. The gneiss is clearly of pre-F_2 age, folded by the F_2 structures. It also occurs on the normal limb of an F_2 structure (not in a fold closure) south of Skarvfjordhamn and again south of Hønsebyfjord. An examination of the general outcrop pattern of the granite gneiss on Skarvfjordfjell, ignoring the extreme irregularity of outcrop, indicates a V-pattern, the apex pointing south. This corresponds with the strike observations on either limb of the fold, plunging as it does in a southerly direction. Moreover, the gneiss is shown to develop preferentially in the more pelitic bands in the Klubben Quartzite Group.

(2) *Finfjordnæringen.*

The serial profiles (Plate III) across this peninsula illustrate the stacking of several F_2 folds of near-recumbent attitude. As mentioned earlier, F_2 axial planes are inclined at between 10° and 20° to the SW; dips of banding on reverse limbs average only 25° to the W.S.W. or S.W. while F_2 normal limbs dip variably between S.E. and N.E. This is quite anomalous to F_2 fold attitudes noted elsewhere on the surveyed area and is representative of the maximum development of overturning. Minor F_2 folds are everywhere congruous to the larger structures.

South-east of Skipsvik the upper limb of the Hellefjord Synform (equivalent to the lower limb of the Langstrand Antiform) is recognisable, minor F_2 folds overturned in a S.W. direction. This picture can be traced as far as the Skarvdal Fault; south of the fault the lower limb of the Hellefjord Synform is present with NE-ENE overturned minor folds, a situation which presents further proof of the relative downthrow to the north of the fault. The axial plane trace of the synform can be found near Veiviken north of the fault, indicating a relative displacement of ca. 1.6 km on either side of the fault. This apparent displacement is clearly due to the near-recumbency of the major fold.

C. Summary and discussion.

The foregoing descriptions of major structures both on an individual and areal basis have established a picture of regional fold superposition which readily explains the existing outcrop pattern while conforming to fundamental stratigraphical principles. Briefly, the regional structure is postulated in terms of the refolding of an early fold of unknown amplitude, the Hønseby Fold, by structures of a later episode of deformation of which the Langstrand Antiform and Hellefjord Synform are major representatives.

Although the inverted sequence on the lower limb of the Langstrand Antiform is essentially a consequence of the F_2 tectonism, it can be shown that the widespread stratigraphical inversion in the northern area is an inherent property of the earlier folding. The pattern of F_1 structures prior to the second episode of folding is somewhat conjectural, though it is reasonably clear that the F_1 tectonism produced tight or isoclinal folds almost certainly of recumbent attitude, considerable magnitude and variable axial trend. The root zone of these early folds is thought to have been far to the W. or N.W. of present-day Sørøy; although evidence is scarce, partial verification of this is found in the mapping by Ramsay and Sturt (personal communication) of a major F_1 syncline which can be traced from the Brevik area in the S.W. across to Storelv and which closes towards the north in the 'EW strike belt' and towards the west near Brevik.

The second episode of deformation, following the acme of regional metamorphism, has yielded the majority of macroscopic and mesoscopic folds mappable in this part of Sørøy. From the prior analysis of minor and major structures, the constancy of axial plane trend and inclination together with the related fold overturning to the ENE stand out as the more salient features of this deformation. Tectonic transport parallels this direction, i. e. towards the ENE. Fold axial and tectonic 'b' lineations show a general plunge towards the S.S.E. It is clear therefore, that monoclinic symmetry of movement prevails, wherein the 'ac' plane and deformation plane are coincident and translation in this plane is normal to the 'b' axis (here towards the ENE) which is an axis of binary symmetry.

Complicating this initially relatively simple picture, however, is the tendency for F_2 linear and planar elements to show slight variations in the northern area but larger and significant variations in the south where a regional swing into the belt of E-W strikes is of major importance. Whereas in the northern area monoclinic symmetry is manifest, in the limited tract of E.W. strikes in the vicinity of Langstrandfjord both congruous and incongruous F_2 folds are encountered indicative of orthorhombic symmetry, though in detail triclinic symmetry is possibly prevalent.

Variable structural symmetry dependent on regional strike is also demonstrable in other parts of Sørøy. Significantly the 'E-W strike belt', which occupies a large part of central Sørøy, is characterised by the presence of F_2 conjugate folds reflecting an orthorhombic or lower structural symmetry. These features have been described by Ramsay and Sturt (1963) from Sandøfjord and noted also by Appleyard (1965) in the Dønnesfjord area. In the extreme south-west of Sørøy the Breivikbotn and Hasvik region constitutes a tract of N-S strikes and there monoclinic structural symmetry again obtains. The sense of overturning of F_2 folds is also to the east. Some 40 km S.W. of Sørøy in the Loppen–Kvænangen area mapped by Ball et al. (1963), strikes are again prevalently N-S and F_2 folds show a consistent E.N.E. overturning.

On a regional scale therefore, as in localised areas, a relationship between strike and F_2 structural symmetry is seemingly indubitable. As well as the orthogonal strike swing, an associated swing of F_2 fold axes may be observed. On first sight this would appear to suggest a later tectonic episode, of NW-SE trend, as being responsible for the F_2 fold swing, but this does not help to explain the related symmetry changes since conjugate structures are restricted to the 'E-W belt' while the 'N-S belts' are characterised by more intensive tectonic stretching and monoclinic undirectional F_2 structures. If a later deformation were responsible therefore, it would be extremely difficulty to explain the strict and adherent relationship of F_2 fold symmetry and overturning to strike and fold axial trend.

The marked regional swing of strike must therefore be regarded as a feature of either the first or second episode of deformation. Although potentially important areas of Sørøy have still to be mapped, the evidence to date would suggest that the swing is an inherent property of the F_1 episode, but a coeval F_2 development cannot be entirely ruled out. Regionally, as far as can be ascertained, the direction of F_2 tectonic transport appears to be towards an easterly point, that is, away from the central parts of the Cambro-Silurian geosyncline and towards the bounding foreland. Although it is perhaps too premature a suggestion in view of the present incomplete evidence, the F_2 fold axes of the 'E-W strike belt' could, however, be visualised as paralleling the regional F_2 'a' direction (even though locally they are 'b' lineations). The 'E-W belt' may then be regarded as a zone in which the regional monoclinic symmetry of movement has been relaxed: with paired folds having developed related to conjugate shears, the symmetry is then orthorhombic or even triclinic.

In many ways this situation is similar to that existing in areas described by Lindstrøm (1955, 1957, 1958) from the Swedish Caledonides and by Johnson (1956) from the Moine Thrust Belt of the North-west Highlands of Scotland. Both these areas are characterized by dominant lineations and fold axes aligned transversely to the major tectonic 'b' direction. Furthermore, linear structures normal to the margin of the fold belt (regarded as 'a' lineations) appear to occur quite consistently along the length of the Caledonian mountain chain in Norway (Kvale 1953). The tectonic pattern prevailing in this north-eastern part of Sørøy can now be considered in relation to these regional structural features, although at this stage conclusions can only be tentative on account of the tenuity of the evidence.

With regard to the Scandinavian Caledonides the central parts of the orogenic belt have lineations and fold axes trending roughly parallel to the orogen whereas the marginal areas with large overthrusts have prominent lineations and axes of small-scale folds which parallel the direction of movement (Kvale, 1953). The various areas mapped by Lindstrøm in Sweden include overthrust metamorphic complexes in which 'a' lineations and fold axes are the major linear element. This accords with the above-mentioned relationship of 'a' lineations to the orogen marginal zones in proximity to thrusting.

A significant observation concerning fold axial trends in the Kartimvare area of Swedish Lapland (Lindström 1957) is strikingly similar to observations on Sørøy. Lindström states, "... folds striking N-S or N.E.–S.W. are constantly overturned towards the E. or S.E. whereas the transverse folds are overturned to the N.E. or S.W. with no clear preference for either direction. This is good confirmation of the hypothesis that the transverse folds are parallel to the tectonic 'a'-axes or the main direction of transport".

This is a similar pattern to that indicated by observations of the directions of fold overturning on Sørøy in the N-S and E-W strike belts respectively. However, an important difference is apparent: this is that, while the distribution of E-W and N-S folds is random on Lindström's area, in the writer's area of N.E. Sørøy preferential fold overturning is confined to the 'N-S strike belt'. Inthis N-S belt as discussed earlier, F_2 folds are consistently overturned to the E.N.E. F_2 folds with inconsistent translation (either to the N. or S.), including conjugate folds, are confined almost exclusively to the 'E-W strike belt'. This is corroborated by the work of Ramsay and Sturt in central and S.W. Sørøy.

Another difference between Sørøy and the Swedish Caledonide areas of Lindström is that in the Swedish areas the regional transverse 'a' lineation is dominant and pervasive. On Sørøy however, a regional 'a' lineation is quite subordinate in the main 'N-S strike belt' although the local 'b' lineations and fold axes in the 'E-W strike belt' are essentially equivalent to the regional transverse 'a' element in terms of trend. It is obvious, therefore, that the structural picture on Sørøy is not entirely correlative with that of the Swedish thrust zone despite similarities of fold trends and overturning. Further to the south in part of the Nordland area of central Norway, Rutland (1959) has commented on the absence of pronounced linear structures parallel to the local movement direction. He then considered this area as representative of the central zones of the Caledonian geosyncline. Accepting this postulate, the Swedish and Nordland areas could therefore be taken as examples of the marginal (thrust) and central zones respectively of the Caledonian orogenic belt. The tectonic picture on Sørøy would appear to conform to neither of these patterns though containing certain elements of both, so that it could possibly be regarded as representative of an area part way between the central and thrust zones of the Caledonian mountain chain.

As analogy may also be made with the Moine Thrust Zone of the North-West Highlands of Scotland. There the major tectonic thinning and extension has been compensated by the formation of zones of tectonic thickening and folding about E-W axes, the folds indicating no consistent direction of translation (Johnson, 1956). The symmetry of these E-W folds is essentially orthorhombic, as compared with the regional monoclinic symmetry, and the folds parallel the major tectonic 'a' direction.

Sørøy is situated some 90–100 km N.W. of the Caledonian Thrust Front. A major thrust cuts the plateau surface of West Finnmark, however, revealing Pre-cambrian rocks in tectonic 'windows' only 30–40 km S.E. of Sørøy. Although thrusting is absent on the writer's area, it is quite possible that major thrusts may occur at a relatively shallow depth beneath the island. This is based on the postulated extension of the gently dipping West Finnmark thrusts westwards and north-westwards beneath Seiland and Sørøy.

From these various considerations and structural comparisons of selected areas in the Scandinavian and Scottish Caledonides and Sørøy, the main conclusions to be drawn are necessarily speculative. While many striking similarities of structural symmetry relative to fold translation are apparent between these areas, the Sørøy structural reconstruction would appear to be not so easily explicable in terms of its relationship to the regional structural picture.

That marked differences of F_2 fold symmetry dependent on the prevailing strike trend exist cannot be disputed, but the overriding cause of this apparently coeval yet variable symmetry might not be deduced until a fuller regional investigation has been completed. Since any attempt at explaining the inconsistent movement picture must be considered in a regional context this lack of structural information is a decided hindrance. One consideration is that regarding the regional B-axis as a direction of low compressive stress, unequal elongation in 'b' could produce appreciable departures of B-structures from the normal trend; factors having a bearing on such a deviation are numerous but include varying resistance to compression, possible proximity of the basement at no great depth and inhomogeneity of the supracrustal material.

Another possible explanation for the changing F_2 trend and symmetry also appeals to a resistance or obstruction to the regional compressive stresses. The great basic and ultrabasic complex of the Stjernøy–Seiland–Øksfjord and southern Sørøy area was almost certainly emplaced either immediately preceding or, in part, during the early stages of the protracted F_2 movement episode. Alkaline intrusions in S.W. Sørøy clearly post-date the main basic and ultrabasic rocks and were themselves emplaced during the initial stages of the second deformation episode (Sturt and Ramsay 1965). It is therefore possible that the F_2 folding, in deforming the metasediments, was partially deflected around this extensive plutonic complex. The transverse stresses thus set up can be envisaged as giving rise to structures of orthorhombic or lower symmetry, local movement being transverse or oblique to the regional direction of transport. Problems associated with this question of fold axial swing and symmetry variation may, however, not be resolved until the regional tectonic picture is more fully known.

III. Metamorphism

1. Mineral and textural relationships of the metasediments

Structural evidence has shown that the metasedimentary sequence has been subjected to two main episodes of deformation, the younger of which can be subdivided into a number of phases. The mineralogy and lithologies, ranging up to kyanite-sillimanite schists and granite gneisses, clearly indicate a fairly high grade of regional metamorphism (almandine-amphibolite facies), and from a study of textural and mineral relationships it can be shown that the fold movements were accompanied by metamorphic recrystallisation. The early-F_2 movements occurred under higher grade conditions than the F_1 deformation, while the late-F_2 movements were accompanied by retrogressive metamorphism. Furthermore, it is abundantly clear that the peak of the metamorphism was reached late in the static interval separating the two main generations of folding and continued into the earliest phase of the F_2 deformation.

On the basis of textures and mineral growth relative to the folding, the metamorphic history is best discussed as phases:

(1) The F_1 syn-tectonic metamorphism.
(2) The main, post-F_1, pre-F_2 (static) metamorphism.
(3) The early-F_2, syn-tectonic metamorphism.
(4) The post-S_3 (static) metamorphism.
(5) The late-F_2, retrogressive metamorphism.

There is no evidence to infer these stages as being discontinuous and indeed the metamorphism is regarded as a more or less continuous and progressive process of mineralogical and textural change. Metamorphic grade waned during the F_2 deformation, the last stage being marked by characteristic diaphthoretic features and minerals.

(1) *The F_1, syn-tectonic metamorphism.*

The earliest recognisable metamorphic fabric is that which developed concomitantly with the F_1 folds. This early fabric is represented by the primary schistosity, S_2, but because of the subsequent protracted and intense deformation and recrystallisations to which the rocks have been subjected, only scant evidence of this fabric remains.

In all lithologies the dominant and pervasive fabric is that of S_3. The chief indication of the existence of an earlier schistosity is the preservation of included grains within post-F_1 porphyroblastic minerals, this relict F_1 inclusion fabric usually being of extremely fine grain and distinctly finer than the groundmass fabric outside the porphyroblasts. Inclusions are generally of quartz with variable amounts of ore grains, micas, plagioclase and apatite. Commonly the included grains are elongate and exhibit a parallel dimensional orientation within the porphyroblasts (Fig. 60). Fineness of inclusion grain size varies with lithology; thus, grains enclosed by garnets in the phyllitic Hellefjord Schist, while contributing to the oriented relict fabric, may be almost unresolvable. It can be demonstrated that the fineness of grain size is not entirely an absorption feature. Neighbouring included quartz grains are seldom optically continuous and in many instances grain clusters provide additional evidence for the fine-grained nature of the preserved relict fabric.

In quartzites the prevailing schistosity is usually that related to F_2. Some biotite flakes, however, show a parallelism with the banding and in pelitic stringers the biotites often tend to be aligned with (001) subparallel or parallel to the banding. These micas have recrystallised mimetically following the primary schistosity rather than align themselves in S_3.

The maximum grade of metamorphism attained during the F_1 generation of folding apparently did not exceed that of the biotite zone. The meagre textural evidence, principally that seen within porphyroblasts, indeed points to low grade conditions as prevailing at this time with chlorite probably stable. As regards rock-types, phyllites were probably developed from calcareous mudstones and other argillaceous members of the succession. The extent of recrystallisation within the arenaceous rocks is impossible to judge but it can be noted that the sub-rounded clastic grains of quartz with some feldspar occurring in a 'metasandstone' band within the Storelv Schist NW of Skarvfjordhamn (Fig. 9) have retained their primary character despite the lengthy polymetamorphism.

(2) *The main, post-F_1, pre-F_2 (static) metamorphism.*

After the F_1 deformation but prior to the initiation of the second generation of folding, static conditions obtained during which time the bulk of the porphyroblastic minerals were developed. In this, the most constructive stage of the regional metamorphism, garnet, staurolite and kyanite overgrew the earlier fabric while later in this period and during early-F_2 the growth of sillimanite signifies the attainment of the highest metamorphic grade and P-T conditions. In the calcareous lithologies diopside and amphiboles crystallised as porphyroblasts.

The pre-F_2 static phase is almost certainly a continuation of the F_1 metamorphism necessitating a progressive rise of temperature and entailing a gradual coarsening of the textures in the schists. A delimitation of particular metamorphic zones is impracticable, however, since garnet is ubiquitous and both kyanite and sillimanite are found throughout the area in favourable lithologies. Evidence substantiating the pre-F_2 growth of these porphyroblasts is outlined below. The major criteria for dating the mineral crystallisation are provided by textures and the fact that these porphyroblasts are often deformed by F_2 movements. In several cases growth of particular minerals continued into the F_2 deformation episode.

Biotite

Porphyroblastic biotite is most abundantly present in some impure banded limestones and graphitic schists. Within the limestones, biotite flakes up to 1.4 mm across occur in thin layers rich in quartz and silicate minerals and are noted from widely distributed localities – from Langstrandfjord to Finfjord. The biotites are randomly oriented, invariably deformed and exhibit a marked uneven extinction, while the groundmass schistosity, S_3, is deflected around them. An inclusion fabric is rarely present. Flake margins are often ragged, the biotite breaking down to quartz and matrix biotite with scapolite developing and occurring as a corona around some flakes. In a graphitic schist from Finfjorddalen, biotite porphyroblasts are deformed by the main F_2 folding and show incipient chloritization. These biotites contain fine dust trails which are representative of the bedding.

Garnet.

From detailed examinations of the textural relationships of the inclusion fabric within garnet porphyroblasts and the external groundmass fabric, it can be demonstrated that garnet growth commenced in this static period and in many cases continued well into the lengthy F_2 episode of deformation. The present description is restricted to those garnets developed prior to the F_2 folding; zoned garnets containing static and kinematic crystallisation evidence are discussed later.

The relationship of the inclusion trails within garnet porphyroblasts to the fabric of the groundmass schistosity is an important criterion in recognising the age of crystallisation. Garnets containing a non-rotational arrangement of inclusion trails of tiny, often elongate quartz grains are interpreted as pre-tectonic (here, pre-F_2) when the inclusion fabric is oblique to the schistosity of the groundmass (e. g. Rast 1958, Sturt and Harris 1961, Johnson 1962). Idioblastic garnets in the Hellefjord Schist frequently exhibit this textural relationship, the S_3 schistosity spindling the porphyroblasts (Fig. 60). It is significant that the groundmass grain size is appreciably larger than that of the relict inclusion fabric, the interpretation being that the garnet has grown over the earlier finer grained S_2 fabric and has subsequently been rotated during the development of the coarser grained S_3 fabric. Thus, in any one schist band the inclination of the included S_2 fabric to the groundmass S_3 may vary depending on the degree of rotation during F_2. As a rule, however, the obliquity of the included trails remains fairly constant within any one thin-section. In some cases the trend of the inclusion fabric is normal to that of S_3.

With the recrystallisation of the groundmass, development of S_3 and rotation of garnets associated with planar-slip, quartz may be observed in the pressure-shadow areas adjacent to porphyroblasts. This is more common in the coarser schists and gneisses. Transverse tensional fractures are developed in many garnets, a further indication of their later subjection to deformation. In the small, 1–2 mm garnets of the pelitic Hellefjord Schist, fractures are extremely rare possibly because the finely poekiloblastic nature of the garnets was not conducive to fracturing. Another possible explanation is that the biotite-rich phyllitic schist was sufficiently incompetent as to inhibit the development of tensile fractures. In the slightly coarser, calc-siliceous bands of the Hellefjord Schist, garnets crystallised in this static phase often display an irregular, almost skeletal outline. They are poekiloblastic but in-

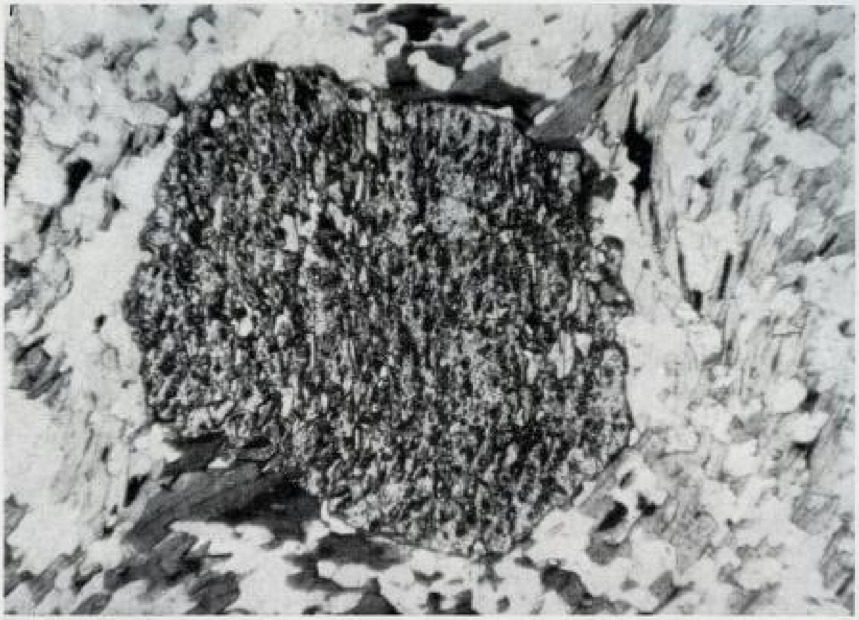

Fig. 60. Pre-F_2 garnet; parallel dimensional orientation of inclusion fabric, oblique to S_3 (trending 'NE-SW'). Hellefjord Schist. Plane polarised light, x48.

clusions are fewer and of slightly larger grain size than those in the more pelitic schists. This is a largely a reflection of the varying F_1 fabric.

The coarser, kyanite-sillimanite schists also contain garnets which often exhibit a rectilinear arrangement of a finer grained inclusion fabric, though sometimes this earlier fabric may only be seen in thin-sections cut parallel to the F_2 'b' direction. Garnets are frequently observed to be elongated in this F_2 'b' direction (Fig. 61) and fractures are present in both 'ac' and 'bc' planes, less noticeably where garnets are highly poekiloblastic. The S_3 groundmass fabric is again deflected around the garnets.

Evidence from the psammites indicates that garnet growth was predominantly static and pre-F_2. Sporadic garnets are present in many thin-sections of massive quartzite but are more abundant in micaceous psammites and semi-pelites: pelitic ribs within a psammite are sometimes studded with garnet porphyroblasts. Garnets are rarely idioblastic in mixed pelite-psammite lithologies; they usually contain only a few

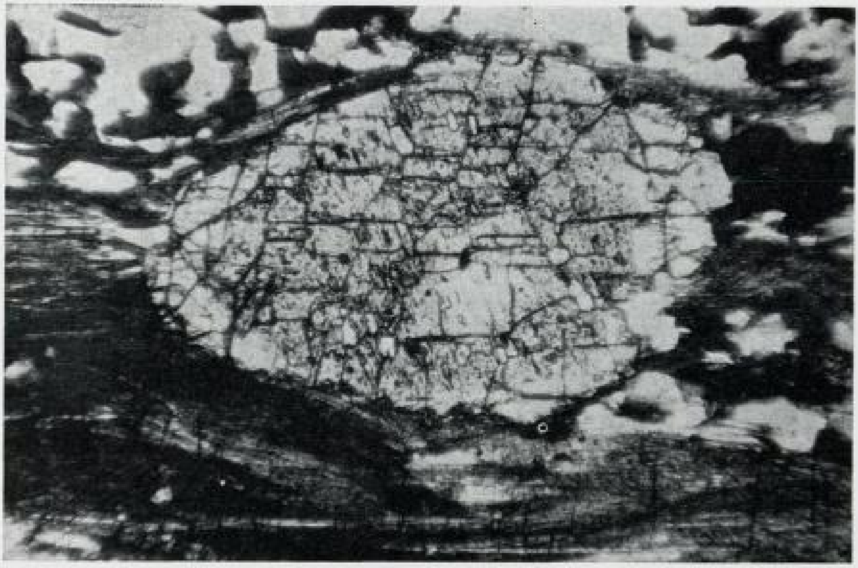

Fig. 61. Pre-F_2 garnet containing early rectilinear inclusion fabric. Fibrolite abundant in lower half of photo shows transverse fractures. Thin-section cut parallel to F_2 'b'. Schist from Falkenes Limestone Group. Plane polarised light, x55.

quartz grains but where a sieve-structure is present the internal fabric is normally less coarse than that of the groundmass but coarser than that observed in garnets from other more pelitic lithologies. Many garnets developed in the psammitic lithologies are fractured, often considerably, with the fracture direction normal to the groundmass schistosity and F_2 'b'. S_3 usually spindles the garnets. In psammites from the Skarvfjord–Hønseby area, garnets have been broken down (sometimes comminuted) by strong F_2 axial planar slip and shear and drawn out as trains of granules and grains along the S_3 schistosity. Different stages in this process can be noted.

Garnets with an internal arrangement of clear sectors separated by narrow zones rich in fine inclusions and dust granules (Fig. 63) are present in schists from the Falkenes Limestone Group at Lundhavn and Langstrand. These are similar to the sectoral garnets described by Harker (1932), Rast and Sturt (1957) and Powell (1966). Each of the larger 'clear' sectors exhibits an exceptionally fine, parallel arrange-

Fig. 62. Fractured pre-F_2 garnet with S_3 penetrating between the two segments. Hellefjord Schist. Plane polarised light, x30.

ment of minute, rod-like particles, possibly of carbonaceous dust. This alignment is always normal to the garnet margin and crystal faces. In the marginal zone of the garnet a concentration of graphitic inclusions and opaque dust is frequently present.

Harker (1932) regarded this sectoral zoned arrangement of included particles as polysynthetic twinning and suggested that the inclusions are situated along the traces of the edges of the growing rhombdodecahedra. Rast and Sturt (1957) expand on this hypothesis in stating that, under static conditions, "the purely crystallographic factor of velocity of growth determined those sectors where growth was fast and inclusions were preserved and those where the slowly growing mineral allowed the dissolution of inclusions". Thus, along the edges of the crystal, with dodecahedral surfaces converging towards the centre, growth can be envisaged as being faster than on the faces.

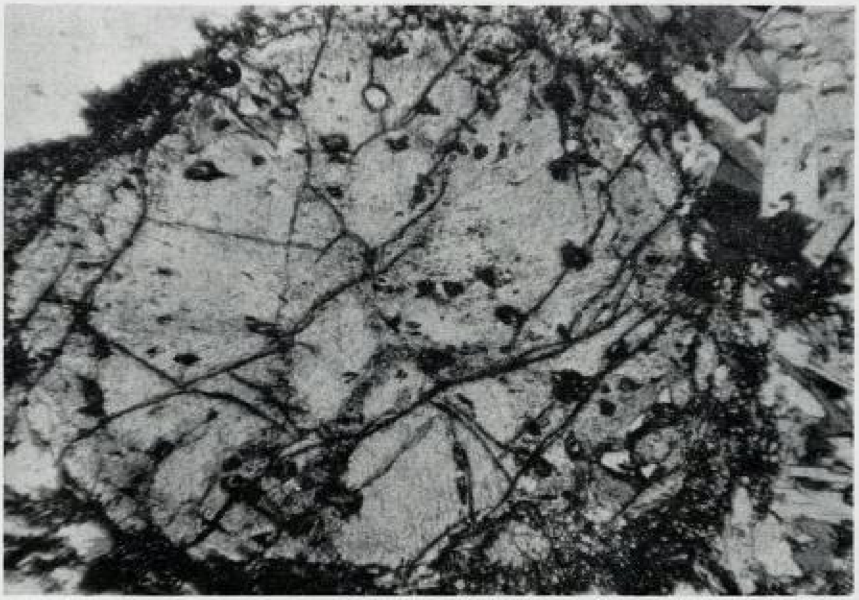

Fig. 63. Garnet showing sectoral arrangement of inclusions. Schist from Falkenes Limestone Group. Plane polarised light, x55.

Staurolite.

Staurolite has been noted from only 5 thin-sections within the Storelv Schist and Falkenes Limestone Group from widely scattered localities. In one specimen of coarse Storelv Schist, from Skarvdalen, staurolites up to 1.5 cm are present and one 'skew' penetration twin – with a (232) pyramid as twin plane – is distinguishable. In thin-section staurolites rarely attain 3 mm; they are occasionally twinned and usually poekiloblastic, the inclusion fabric being distinctly finer grained than that of the groundmass schistosity. Internal inclusion trails (sometimes of ore grains and opaque dust) are invariably straight and often oblique to the external S_3.

Evidence that the growth of staurolite pre-dates F_2 is seen in the bending of crystals (with resulting undulatory extinction) and their fracturing and partial replacement by the S_3 fabric. In a tourmalinised schist from the Falkenes Limestone Group, a pronounced tensional fracture pattern is seen in staurolites and garnets containing identical pre-S_3 inclusion trails. The garnets have a narrow (0.1–0.2 mm) marginal zone devoid of inclusions, which is not always continuous. In a few garnets staurolite is included in this peripheral zone, suggesting that at least the outer zone of clear garnet post-dates staurolite crystallisation. This is in accordance with the general observation that, of the porphyroblastic minerals, garnet has the longest period of crystallisation (see Fig. 72) and is often found with a zonary structure indicative of successive growth stages. Staurolite, on the other hand, is thought to have been stable over quite a restricted range of P-T conditions. In this instance therefore, garnet growth continued after the development of staurolite and since kyanite is also present in this particular schist, staurolite quite possibly has reacted with quartz with increasing temperature to produce the kyanite and additional almandine garnet:

$$\underset{\text{staurolite}}{3\,FeAl_4Si_2O_{10}(OH)_2} + \underset{\text{quartz}}{2\,SiO_2} \rightleftharpoons \underset{\text{almandine}}{Fe_3Al_2Si_3O_{12}} + \underset{\text{kyanite}}{5\ Al_2SiO_5} + 3\,H_2O$$

Kyanite.

This mineral is found as porphyroblasts in many thin-sections of Storelv Schist and schists within the limestone group. It also occurs in thin pelite bands within the massive quartzites both on Blåfjell and east of Skarvfjordhamn. Clearly, an original aluminous sediment is favoured as a locus for kyanite growth.

Criteria for recognising the main growth stage for kyanite as being late in this static pre-F_2 metamorphic phase are essentially identical to those outlined for staurolite and garnet. Deformation features are similar though bent and fractured kyanites are often more conspicuous. In some thin-sections kyanite plates show a definite preferred orientation, the crystallographic 'c'-axis parallel to the F_2 'b' direction. Rarely, kyanite plates are deformed by several kink-bands trending across the length of the crystal either normally or obliquely (Fig. 64). These are probably related to the later phase of F_2 movements. Near Langstrand, kyanites in a graphitic schist are strongly deformed by late-F_2 microfolds.

Although evidence of its age relative to other porphyroblasts is meagre, kyanite growth was dominantly pre-F_2 as evinced by inclusion and deformation features. A finer grained inclusion fabric is however less common in kyanite than in either garnet or staurolite. In the tourmalinised schist mentioned earlier, in one thin-section kyanite occurs as an aggregate of grains within which is enclosed a small garnet. This and other similar evidence would tend to suggest that kyanite postdates the garnet in this instance. Staurolite-kyanite relationships are nowhere seen. Mention can be made here of replacement of kyanite by post-S_3 muscovite in some schists within the limestone group: all stages in this replacement can be traced.

Sillimanite.

The occurrence of sillimanite is largely confined to schists of the Falkenes Limestone Group, but the mineral has also been observed in two specimens of Storelv Schist – one from Langstrand, the other from near Skarvfjordhamn – and in a coarse quartz-kyanite-garnet schist from Finfjordnæringen.

In hand-specimen sillimanite is prominent as small white or cream-coloured 'knots' which are visibly flattened in the S_3 plane and drawn out in the F_2 'b' direction. Microscopic examination confirms that the sillimanite needles and fibres have a marked preferred orientation. Typically the mineral occurs as swarms or trains of acicular crystals having a felted or quasi-fluxional arrangement; at times a feathery pattern is discernible. In thin-sections cut normal to the F_2 'b' lineation, sillimanite is seen as a profusion of minute, square- and diamond-shaped end-sections giving a good positive figure.

A common observation is that the sillimanite fibres are clearly associated and intergrown with biotite and it would appear that the mineral is developing primarily at the expense of biotite. The amount of sillimanite or fibrolite intergrown with biotite is quite variable but frequently the latter shows only as a trace of brownish pleochroism within an extensive fibrous felt of sillimanite, similar to features described by Tozer (1955), Hietanen (1956) and Chinner (1961). Transverse fractures, both 'ab' and 'ac' with respect to F_2 are common both in individual sillimanite needles and in sheaves of oriented fibrolitic sillimanite (Fig. 65). They also transsect partially digested biotite enclosed within the felted mass of fibres and can be observed to pass into garnet porphyroblasts. The precise origin of the sillimanite is somewhat equivocal, since from a chemical point of view the derivation of sillimanite wholly from biotite seems rather unlikely, a topic discussed at some length by Chinner (1961) who also stated that "the impression that sillimanite replaces biotite may therefore be illusory". Further considerations of this matter are reserved for the summarizing discussion.

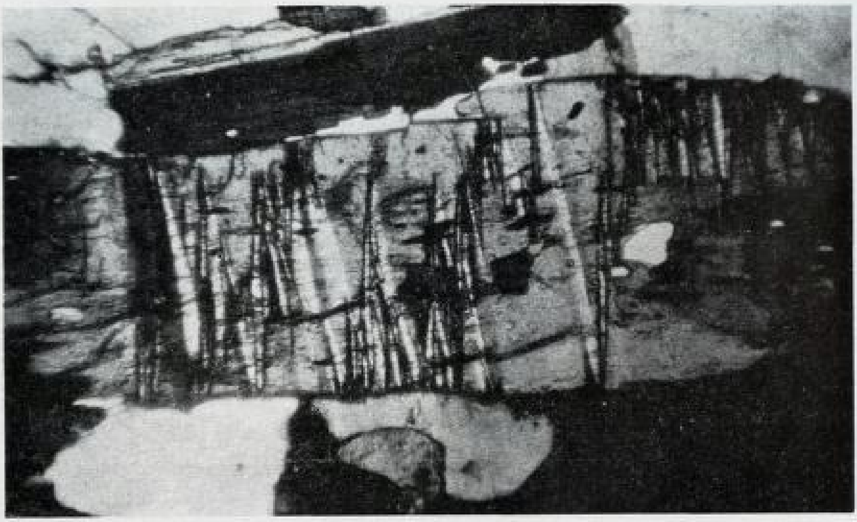

Fig. 64. Kyanite crystal with transverse and oblique kink bands. Storelv Schist. Crossed polarised light, x50.

Fig. 65. Fibrous sillimanite (fibrolite) showing transverse fractures. Thin-section cut parallel to F_2 'b'. Schist from Falkenes Limestone Group. Plane polarised light, x60.

Clear textural relationships between sillimanite and kyanite are difficult to trace though two examples of kyanite being replaced marginally by sillimanite have been noted. Kyanite, and more commonly garnet, may be almost totally sheathed by the fibrolite needles. In some garnets, marginal zones which are otherwise inclusion-free contain needles of sillimanite oriented similarly to those outside the garnet, a feature which suggests the younger peripheral zone of garnet to be post-sillimanite in crystallisation.

Diopside.

Both the Hellefjord Schist and calc-silicate schists of the Falkenes Limestone Group are diopside-bearing lithologies. Limestones also contain this mineral. Diopside occurs chiefly as xenoblastic porphyroblasts, often quite ragged and with variable replacement by amphibole and late scapolite and calcite. The S_3 schistosity is usually deflected around the crystals. In some calcic schists the pyroxene may be present as small, stubby prisms or in a granular form. No strict linear element has been observed. Twins with (100) as twin-plane are only rarely encountered but in some of the larger porphyroblasts, polysynthetic twinning is discernible.

Various stages in the uralitisation of diopsides are demonstrable but the process of replacement is seldom so advanced as in the metagabbros. Garnets, normally skeletal and spongy in calc-siliceous schists, occasionally enclose diopside or uralitised diopside. This would indicate that garnet growth post-dates that of pyroxene, although uralitisation has seemingly occurred both pre- and post-S_3.

Amphibole.

Amphibole, namely hornblende or actinolite, occurs extensively throughout the Hellefjord Schist and in the amphibolites and metagabbros. Within the Falkenes Limestone Group tremolite is the dominant amphibole: cummingtonite occurs in one quartz-garnet schist.

A strong preferred linear orientation of the 'c'-axes of long prisms and needles of actinolite or actinolitic hornblende in the F_2 'b' direction is characteristic of the flaggy Hellefjord Schist. Thin-sections cut normal to F_2 fold axes show a preponderance of lozenge-shaped amphibole end-sections, the S_3 schistosity being deflected around the small porphyroblasts. A planar orientation of amphibole is also discernable with

Fig. 66. Diopside partially replaced by hornblende. Hellefjord Schist. Plane polarised light, x55.

(100) more or less parallel to S_3. The actinolite may or may not contain inclusions, while the margins are sometimes corroded by the S_3 fabric; quartz and biotite are the chief replacive minerals. Uralitisation of diopside is frequently observed (Fig. 66) and in a few instances the uralite is a paler green colour than the bulk of the surrounding actinolite/hornblende.

Tremolite, or tremolite-actinolite, is present in the limestones and associated phyllites and schists. In phyllites, tremolite often occurs as knots or sheaves up to 1.5 cm across, the groundmass schistosity spindling the almost fibrous colourless or pale green mineral agregate. Inclusion content varies enormously. In some highly poekiloblastic tremolite knots the inclusion trails, although oblique to the S_3 trend, are indistinguishable texturally from S_3. Some marginal flexing of the inclusion trails is observed, however, such that the tremolite growth can be envisaged as continuing into the period of F_2 deformation from the pre-F_2 static phase.

In limestones, colourless tremolite is frequently present as scattered laths which, together with phlogopitic biotite, help to define the schistosity. Occasionally, tremolite individuals up to 8 mm across occur in limestones; these are markedly spongy containing abundant inclusions of calcite and sporadic quartz grains. In one limestone specimen from Langstrand the groundmass calcite fabric contains much granular ore and opaque dust which is only very rarely present in the tremolite porphyroblasts.

Cummingtonite has been observed in only one hand-specimen, that of a quartz-amphibole garnet schist from east Finfjord. This mineral also displays a preferred dimensional orientation, while optical properties include: baxial positive figure; colourless, non pleochroic; elongate sections length-slow; $z \wedge c = 20°$. Twinning on the (100) twin plane is common.

Feldspar.

While the feldspathisation features producing the various gneisses are dealt with later, it can be mentioned here that plagioclase porphyroblastesis along with some microcline crystallisation was initiated at the very end of this static period prior to the F_2 tectonism. Many feldspars have been granulated and fractured during the F_2 deformation.

(3). *The early-F_2, syn-tectonic metamorphism.*

Extensive recrystallisation affecting the fabrics of all lithologies is the dominant textural feature of this second deformation period. As a result of this the component minerals are bigger and the general fabrics coarser grained than those produced during the F_1 deformation, the prime factor in the coarsening of the fabrics being the increasing metamorphic grade during the pre-kinematic static phase. Porphyroblastic minerals developed extensively during this latter phase were deformed or re-aligned while some porphyroblastesis continued into the kinematic period.

Quartz and biotite show fairly outstanding recrystallisation features. Biotites exhibit a distinct planar orientation in the plane of the newly developed S_3 schistosity. This is a ubiquitous feature in the Hellefjord Schist and various psammites and semi-pelites. In many thin-sections, S_3 is markedly oblique to the recognisable banding; in some micaceous quartzites abundant scattered biotites are regularly oriented parallel to the F_2 axial planes. As noted earlier, the biotites in pelitic laminae are often mimetic after the earlier S_2 fabric, though now of larger size.

In calc-siliceous schists, biotite and amphibole depict the S_3 schistosity, the latter invariably developing a marked dimensional orientation paralleling the F_2 fold axes. The small S_3 biotites are sometimes seen to be replacing the actinolitic hornblende and while some of this replacement is quite probably syn-kinematic, post-S_3 alteration is not ruled out.

Quartz was extensively recrystallised during this early-F_2 period, a feature substantiated by petrofabric analyses of psammites; this shows the dominant type of quartz (0001) orientation to be a peripheral 'ac' girdle normal to the F_2 fold axes. With one minor exception there is a lack of evidence of a fabric which can be related to the F_1 deformation. Quartzes are not infrequently strained, a feature indicative of post-crystallisation deformation.

Evidence of post-growth rotation of minerals is displayed by porphyroblasts of garnet, kyanite and staurolite which have a rectilinear non-rotational pattern of the internal inclusions arranged obliquely to the fabric of the matrix. In three thin-sections of Hellefjord Schist, the inclusion trails within garnets show a weak sigmoidal pattern (Fig. 67). Garnets having an S-shaped inclusion pattern are usually interpreted as syn-tectonic (Rast, 1958, Sturt and Harris, 1961), indicating rotation during growth. With the Hellefjord Schist garnets, however, the curvature of the inclusion trails is confined to a narrow (0.5–0.3 mm) marginal zone. Moreover, the fine inclusion fabric is of constant grain size from core to margin and is essentially a preserved S_2 fabric. It would appear that, while the bulk of the garnet has grown during the pre-F_2 static conditions, the peripheral zone of curved inclusion fabric has developed during the early stages of the F_2 deformation. This peripheral zone also contains fewer inclusions, probably indicating a slower velocity of growth, and has a slightly higher relief than the core.

The rotated garnets are interesting in that, in any one thin-section, the better developed S-shaped inclusion trails are seen in the biggest garnets. The sigmoidal shape gradually disappears with decreasing size so that the small garnets have a rectilinear inclusion fabric. This appears to be simply a consequence of the sectioning of the garnet porphyroblasts; those cut centrally are naturally of maximum size and would illustrate maximum rotational features, where the axis of rotation parallels the F_2 'b' direction. Garnets cut marginally show little, if any, discernible effects of rotation. Again, in any one thin-section containing rotated garnets the sense of rotation is constant. This complies with theory since on any fold limb characterised by planar slip (paralleling S_3), relative movement should be constant.

Syn-kinematic garnet recrystallisation is seen in the zoned garnets from schists within the Falkenes Limestone Group. In these garnets a

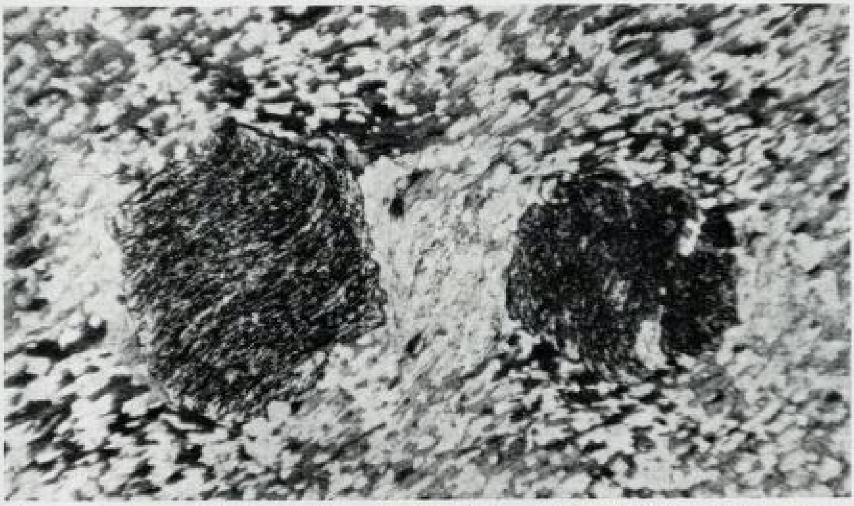

Fig. 67. Garnets with sigmoidal trend of inclusion fabric. Hellefjord Schist. Plane polarised light, x30.

marginal zone containing very few inclusions surrounds the bulk of the garnet which has grown statically over a rectilinear F_1 fabric. Some marginal zones contain sillimanite needles which trend into the sillimanites of the groundmass fabric; in the innermost part of the marginal zone of garnet, the sillimanite needles swing round, and are sometimes bent, into parallelism with the core S_2 fabric. The outer zone of garnet crystallisation was thus probably syn-tectonic, overgrowing the inner static core.

Only one kyanite crystal has been found containing an S-shaped inclusion fabric whereas several retain an earlier rectilinear fabric which is often oblique to S_3. In Fig. 68 the kyanite containing the sigmoidal pattern of internal trails is spindled by the S_3 micas and has associated pressure-shadows of recrystallised quartz. Extinction within the kyanite is only slightly uneven – a range of $2^1/_2^\circ$ was detectable.

Immediately N.W. of Skarvfjord, kyanite-quartz segregatory pods are present in the severely tectonised Storelv Schist. Kyanites up to 15 cm in length are developed in these tectonic lenses and are presumed to have grown during the early-F_2 deformation. Nearby, at Rødberg-odden, similarly developed garnet-biotite lenticles contain garnets measuring up to 8 cm in diameter: these show good rhombdodecahedral

Fig. 68. Kyanite (K) showing sigmoidal trail of quartz inclusions. Groundmass fabric is that of S_3. Q-quartz; F – fibrolite. Schist from Falkenes Limestone Group.

form but are quite impure and crowded with inclusions. They are thought to have developed in favourable low-pressure sites during the intense F_2 deformation.

Sillimanite crystallisation, initiated late in the pre-F_2 static phase, continued into the early part of this protracted kinematic period. Evidence to this effect is seen firstly in the development of some fibrolite from S_3 biotite and secondly in the presence of sillimanite needles in the S_3 fabric which is deflected between the two halves of a fractured and separated garnet.

Some tremolite growth continued into this syn-tectonic phase, while actinolite and hornblende segregations within limestones and calc-silicate schist bands are indicative of an F_2 development. Limestones are composed predominantly of recrystallised calcite with phlogopite or phlogopitic biotite depicting the rude schistosity. Small tremolite laths also parallel this S_3 plane and in some cases produce a lepidoblastic texture.

The processes of feldspathisation and granitisation which were initiated prior to the F_2 deformation can be shown to have continued into this kinematic phase. In the Skarvfjordhamn area, an example of the late granitisation of a mixed semi-pelite/psammite sequence depicts relict F_2 fold closures in quartzite bands, the sheared limbs and semi-pelite bands having been converted into a granite gneiss. Veins of aplite and pegmatite emplaced along or sub-parallel to the axial planes of F_2 folds have also originated during this F_2 deformation episode.

The normal metamorphic feldspar, seen as polygonal or equant grains with boundaries from straight to sutured, has clearly recrystallised during this tectonic phase. Plagioclase composition ranges from An_{24} to An_{39}, but significant variations are discernible within the metasedimentary lithologies. The plagioclase of quartzites and allied rock-types show a variation from An_{24}–An_{32}, in comparison with a range An_{30}–An_{39} for the calcareous Hellefjord Schist. Schists containing kyanite and sillimanite (mostly those within the Falkenes Limestone Group) have plagioclase in the range An_{28}–An_{33}. Progressive granitisation, with gneisses as the ultimate product, yields a more sodic plagioclase feldspar. Microcline is an important constituent mineral in many arkosic psammites, occurring as small fairly equant grains in the groundmass fabric. The familiar cross-hatched twinning may be poorly represented.

Tourmalisation of a garnet-kyanite-staurolite schist near Skarvdalcol is a distinctive syn-F_2 feature and quite unique in this area. Its known outcrop extent is restricted to two exposures some 35 m apart along the strike. Tourmaline enrichment of the schist is irregular, at best a dense closely knit aggregate of crystals enclosing relics of the schist fabric, chiefly partially degraded garnet and highly altered plagioclase. The tourmaline is strongly pleochroic from yellow to olive greens or browns and often exhibits a zonal structure. Some crystals contain abundant granules of ore or graphite. Although the tourmalines overgrow the S_3 fabric they usually exhibit imperfectly developed tensional fractures, parallel to those noted in the garnets and staurolites. This would seem to verify the syn-kinematic (early-F_2) age for the tourmalinisation, a process which would appear to be associated with the early-F_2 granitisation.

In the tourmaline-rich segregations, iron ore (magnetite) laths – some 3 mm in length – are numerous and appear to be associated with the tourmalinisation. They are sometimes flexed by the late-F_2 movements.

Tourmaline is a common accessory mineral in many thin-sections of Hellefjord Schist, particularly the less calcareous horizons. It occurs as minute, olive-green prismatic grains (rarely > 0.1 mm) which are always aligned within the S_3 schistosity; hexagonal end-sections constitute $> 80\ \%$ of the tourmaline grains found in most thin-sections cut normal to the F_2 'b' direction, a feature strongly indicative of their preferred orientation parallel to the F_2 fold axes. While this tourmaline

Fig. 69. Brecciated and comminuted limestone from minor slide zone. Fragments mainly of calcite crystals with some quartz and tremolite. Plane polarised light, x25.

may possibly represent original recrystallised grains, that occurring in some pegmatised Storelv Schist is almost certainly of metasomatic origin.

A mylonite or calc-mylonite which owes its formation to early-F_2 sliding is present in a 20–60 cm thick zone near the top of the Lower Limestone. The typical calc-mylonite is composed of finely comminuted calcite, at times cryptocrystalline with streaky laminae containing dusty ore or graphite. Fragments of calcite crystals, quartz, tremolite, phlogopite and schist lenticles are quite abundant and up to 4 mm across (Fig. 69). Cataclastic textures and structures are ubiquitous in this lithology. Calcite fragments exhibit glide lamellae many of which are bent, some showing an S-shape and others fractured with the granulated matrix penetrating along the fractures. Marginal granulation is commonly seen. Quartzes are intensely strained, sometimes with a biaxial figure, and Boehm deformation lamellae are quite conspicuous. Fractures are present in quartz and in tremolite fragments along which matrix carbonate has penetrated, while phlogopite may be flexed through $> 90°$.

(4) *The post-S_3 (static)metamorphism.*

The S_3 schistosity, developed during the early-F_2 phase, is frequently found to be succeeded by the crystallisation of certain porphyroblastic minerals and, in turn, many of these post-S_3 porphyroblasts show deformation features ascribed to the latest, brittle phase of the F_2 tectonism. Two main criteria helpful in the recognition of post-S_3/pre-late-F_2 crystallisation are:

(i) The continuation of the trend of the groundmass schistosity, S_3, through porphyroblasts without any deflection of the constituent minerals.

(ii) The abrupt truncation of groundmass trends by porphyroblasts around which there is no spindling (deflection) of the S_3 mica fabric.

Minerals which crystallised in this post-kinematic static phase include muscovite, tremolite and garnet. Scapolitisation, uralitisation and some chloritisation may also be attributed to this metamorphic phase. Comparing the post-S_3 mineral assemblage with those of the pre- and syn-early-F_2 crystallisation phases it is at once apparent that this static post-S_3 episode is one of waning metamorphic grade.

Three modes of occurrence of muscovite are referable to post-S_3 crystallisation though one is quite probably continuous from the preceding kinematic phase. These are:

(a) The development of irregularly shaped porphyroblasts in psammites.

(b) The replacement of kyanite (and less commonly sillimanite) by large muscovite flakes.

(c) The bleaching or muscovitisation of biotite.

With regard to the first occurrence, late muscovite in quartzites is present as ragged flakes up to 2 mm across. These are commonly markedly oblique to the S_3 fabric enclosing small biotites (or partially chloritised biotite) which parallel the groundmass trend. The larger muscovite flakes are frequently highly poekiloblastic and of irregular outline owing to the enclosure of groundmass quartz and feldspar. It appears that the irregular shape of the muscovites has been partly conditioned by the

Fig. 70. Kyanite (high-relief) partially replaced by muscovite. Some biotite (dark grey) also present. Schist from Falkenes Limestone Group. Plane polarised light, x55.

outlines of the grain boundaries of quartz and feldspar. Some muscovite is vermicular, either wholly or partially, the vermicules being of quartz.

Kyanite is replaced by muscovite (Fig. 70) in schists from the Falkenes Limestone Group and also in some specimens of Storelv Schist. In many instances the mica appears to be pseudomorphous, occurring as long flakes containing optically continuous relics of the original kyanite. In other cases several small muscovite flakes, of random orientation, are replacing kyanite. All stages in the breakdown may occasionally be seen within one thin-section and some muscovites may also enclose grains of quartz, tourmaline and ore as well as small flakes of groundmass biotite. Fine slivers of ore along muscovite cleavages may result from a partial breakdown of biotites. In a few sillimanite-bearing schists, muscovite flakes which overgrow the schistosity at random appear to enclose trains of acicular sillimanite. One schist contains knots of fibrous sillimanite which are partially replaced by a post-S_3 generation of muscovite; gradations in this replacement are observed.

The third occurrence of post-S_3 muscovite, probably partly syn-tectonic, is that derived by a bleaching of biotites, essentially a process of muscovitisation. This is noted in some psammite and schist specimens. In the latter, lepidoblastic muscovite containing abundant ore granules and slivers of ore along the cleavage planes frequently encloses streaky biotites in a stage of replacement by muscovite. This biotite is often a distinctly pale brown colour with ore granules and iron oxides (mostly limonite) adjacent to the remnant flakes. Many muscovite individuals have oxidation stains along cleavages and sometimes mantling the flakes, while occasionally the iron oxide occurs around quartz grains.

Garnet post-dating the S_3 schistosity occurs mainly in amphibolite dykes but also in some of the schists of the Falkenes Limestone Group. In both these occurrences only the outer zone of zoned garnets is post-S_3, crystallisation of this outer zone being envisaged as continuing after the cessation of rotation and planar-slip. Small garnets, devoid of inclusions and truncating S_3 micas, characterise parts of the coarse Storelv Schist and would appear to be of post-S_3 development. In some thin-sections, however, an extraordinary fine-grained fabric occurs in the cores of these garnets, the outer zone remaining clear. In this case it is plausible to regard the initial garnet growth as being pre-F_2 with slow crystallisation and accretion occurring throughout and after the development of the S_3 fabric.

A striking post-S_{-3} mineral growth is that of small, idioblastic tremolite in a phyllite from the Finfjorddalen limestone scarp. The tremolite occurs as laths up to 4 mm in length in a narrow zone of phyllite adjacent to a diopside-tremolite boudin. The fine-grained phyllitic fabric continues without any deflection through the randomly oriented tremolite porphyroblasts (Fig. 71). Inclusions within the tremolite are principally small, slightly elongate, quartz grains with sporadic, partially digested phlogopitic biotite sometimes visible. As well as enclosing the groundmass schistosity these porphyroblasts also abruptly truncate the S_3 fabric.

It is probable that some uratilisation has continued during this immediately post-S_3 phase of waning metamorphism. Evidence for this is often ambiguous even within any one thin-section, but the inclusion of undeflected S_3 biotites is suggestive of a late amphibole growth which has also extended out from the boundaries of the original pyroxene crystals.

Scapolite is a common post-S_3 mineral in calc- siliceous schists and impure limestone. It replaces both amphibole and diopside and less commonly plagioclase, enclosing S_3 biotites and some quartz. In impure

Fig. 71. Post-S_3 tremolite porphyroblasts. Tremolites clearly overgrow and truncate the S_3 fabric. Phyllite from Falkenes Limestone Group. Plane polarised light, x40.

limestones scapolite porphyroblasts may occur up to 3 mm across, preserving the S_3 fabric as trails of small biotite flakes. Secondary calcite, present in many calcic schists, occurs in certain Hellefjord Schist specimens in irregular form enclosing biotites, quartz and ore grains of the S_3 fabric. It is frequently present in association with diopside, uralite and scapolite and appears to post-date all these minerals. Calcite occurring in the nodes of boudins and pressure shadows adjacent to porphyroblasts of garnet in certain schists probably crystallised initially during the early-F_2 phase though some growth in these loci quite feasibly continued post-kinematically.

Chlorite, replacing biotite and garnet, is mostly found as a late diaphthoretic mineral particularly where seen associated with late-F_2 brittle structures. Some chloritisation features may date from this post-S_3 static metamorphic phase, particularly chloritised biotite flakes which are flexed by later strain-slip movements. In the tourmalinised schist previously described, radial or fan-shaped aggregates of pale grey-green chlorite (some 'fans' up to 4 mm across) occur in the tourmaline segregations. The chlorite is probably ripidolite. It is clearly associated with and is replacing tourmaline, but also corrodes and partly replaces garnet. Plagioclase grains enclosed within the tourmaline-chlorite segregations are now almost completely altered to a grey, cloudy, cryptocrystalline mass. The chlorite of this tourmalinised schist is sometimes deformed by numerous kink-bands which trend at high angles to (001). Simple bending of the (001) cleavage provides further evidence of its later deformation. Rarely, a discontinuous mantle of small muscovite flakes is present around garnets within this chlorite segregation and these also exhibit late-F_2 deformation features.

(5) *The late-F_2, retrogressive metamorphism.*

This represents the ultimate stage of the metamorphic history and took place simultaneously with the late-F_2 brittle phase of deformation. A mineralological breakdown of the former higher temperature assemblages is the significant feature of this metamorphic phase; various mechanical deformative effects are products of this dislocation metamorphism.

Chloritisation of biotite and garnet and less commonly amphibole is a major diaphthoretic process, though usually of restricted occurrence. Chlorite may replace and pseudomorph garnet. Preferential chloritisation of biotite is present along several irregular fractures and late-F_2 strain-slip cleavages ($\leq$ 0.2 mm wide) in some thin-sections of Hellefjord Schist, and where dodecahedral pre-F_2 garnets are situated along or near to such dislocations they too are partially or completely chloritised (Fig. 72). Complete pseudomorphing of the garnets by chlorite (sometimes with grains of clinozoisite) is better seen along the strain-slip cleavage, the chlorite pseudomorph retaining the F_1 quartz fabric. It is notable that chloritisation proceeds initially along the fractures within the garnets. Clinozoisite also occurs along the fractures and strain-slips.

Many thin-sections of psammite contain biotites which exhibit incipient chloritisation along their margins and cleavages. A curious localised chloritisation of biotite is present in one garnet-bearing aplite (spec. LR. 39), in this case the chloritised biotites being restricted to a 1 mm-wide aureole around the garnet porphyroblasts. This feature is particularly noticeable near the contact between aplite and psammite. In the metasediment a 2 mm-wide biotite-rich 'basic-front' is quite conspicouous: where garnets are situated along the aplite-psammite contact the chloritisation of biotites in a zone around the garnet individuals is prominently displayed.

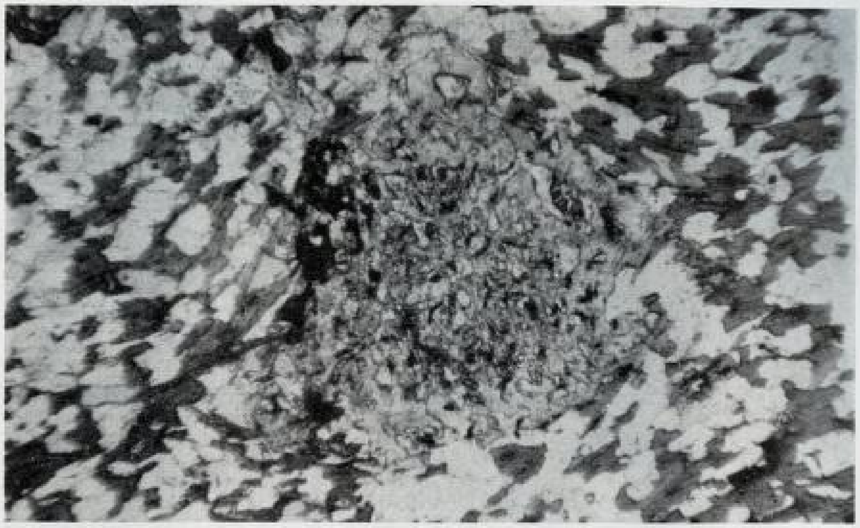

Fig. 72. Chloritised pre-F_2 garnet retaining S_2 inclusion fabric. S_3 trends 'ENE-WSW'. Plan polarised light x48.

Alteration of plagioclase to sericite or saussurite is mostly confined to this late metamorphic phase and possibly also partially post-dates the movements. Within late-F_2 shear zones and fault-breccias, plagioclase is not infrequently altered to a grey, cloudy, cryptocrystalline mass.

Intense cataclastic features invariably accompany the late-F_2 shearing and faulting. Within shear zones examined in thin-section, numerous sub-parallel planes of more intense shearing and granulation interdigitate with thin lenses of incompletely comminuted metasediment or individual mineral grains. Flexed sericite flakes are sometimes aligned along the shear planes, while in many examples of shears and shear zones chlorite is absent but both biotite and amphibole are bent and granulated respectively. Magnetite occurs frequently along shear zones as irregular grains, patches or threads; it is sometimes partly altered to haematite or limonite.

Severe straining of quartzes, bending, rupture and microfaulting of feldspar cleavage and twin planes, and the fracturing and granulation of amphibole are among the cataclastic phenomena associated with fault-breccias. Garnets may be granulated, biotites chloritised and plagioclase masked by a cloudy breakdown product. Ramifying veinlets of finely granulated material enclose less highly deformed mineral grains and metasediment or gneiss fragments. Ore grains are often present and have crystallised during this late-F_2 episode. Magnetite, altering to haematite, is common and occasional veinlets of limonite (altered from magnetite?) may be observed, together with clinozoisite. Calcite occupies previously existing cavities in certain fault-breccias and occurs as veins up to 10 cm thick along some joints in the Hellefjord Schist. Quartz and zeolite can also be found as films along joint surfaces.

In summary, the late-F_2, retrogressive, metamorphic episode is characterised by the mineralogical and physical breakdown of higher grade metamorphic mineral assemblages. It clearly represents a low-temperature dynamic or dislocation metamorphism.

Concluding discussion.

The examination of the evidence provided by mineralogical and textural relationships in the metasediments demonstrates that five major crystallisation and recrystallisation phases of the metamorphism can be delimited on the basis of their relations with tectonic episodes. It is not thought that any appreciable break occurred between any of these phases since they are themselves partly overlapping and are associated with either the absence or advent of tectonic activity: crystallisation of several minerals extended from one phase to another (Fig. 73) so that the metamorphism is to be regarded as a continuous and generally progressive process of textural, structural and mineralogical change. Despite variable crystallisation periods and the difficulty of recognising clear-cut textural relationships between certain minerals, e. g. staurolite and kyanite, the order of appearance of zonal index-minerals corresponds with that characterising the Barrovian-type facies series (Barrow 1912, Miyashiro 1961), namely (chlorite→) biotite→ garnet→ staurolite→ kyanite→ sillimanite, as indicated in figure 73.

In this Langstrand–Finfjord area of Sørøy it has not proved possible to trace zones of differing metamorphic grade as the several index-minerals are fairly evenly distributed in the various lithologies. Kyanite and sillimanite are restricted to pelites and occur throughout the area. The presence of kyanite in thin pelite bands within the massive quartzites (both in the north and south of the area) would seem to indicate that originally aluminous sediments have facilitated the growth of this mineral. Within the Storelv Schist, kyanite is especially abundant in

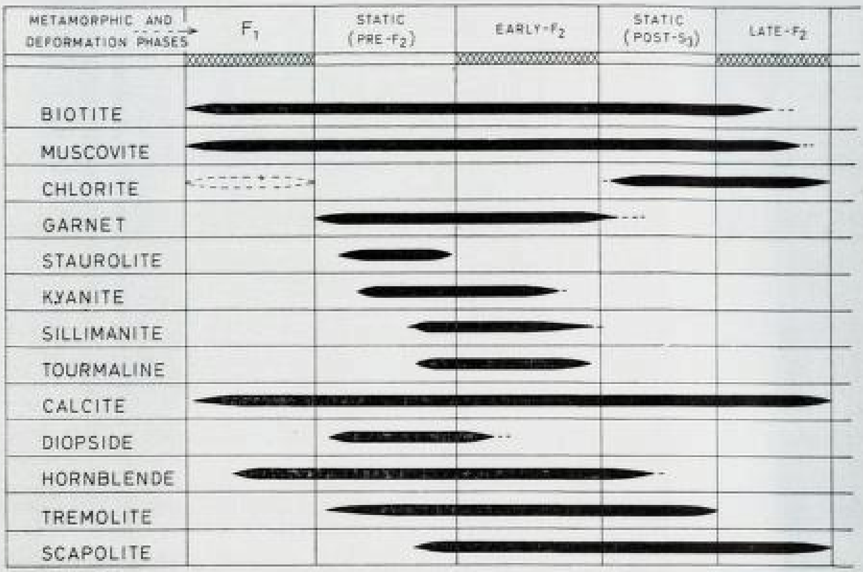

Fig. 73. Diagram showing relations between the structural time-scale and metamorphic mineral growth (metasediments).

the north-west where it is thought to be related, in part, to the locally intense movements of the early-F_2 period.

Sillimanite and kyanite coexist in many schist specimens collected over the area and kyanite is regarded as having persisted into the sillimanite zone. Examples of local persistence of kyanite into this latter zone with both minerals occurring in the same rock are not uncommon (e. g. Wyckoff, 1952, Hietanen, 1956), and Clark et al. (1957) have pointed out that the polymorphic transition from kyanite to sillimanite tends to be rather sluggish. In the present area, sillimanite development from kyanite is rare however (only two examples found); instead, sillimanite or its fibrous variety fibrolite, appears to be developing at the expense of biotite, as described earlier. An actual breakdown (and replacement) of biotite involving an internal rearrangement of silicon and aluminium ions has frequently been cited as a not uncommon process in sillimanite derivation (e. g. Deer et al., 1962). Epitaxial nucleation of sillimanite on biotite has been proposed by Chinner (1961), in this case involving the transfer of alumina and silica from unstable kyanite through the medium of a fluid phase. The applicability of this latter process of sillimanite nucleation in the case of the Sørøy

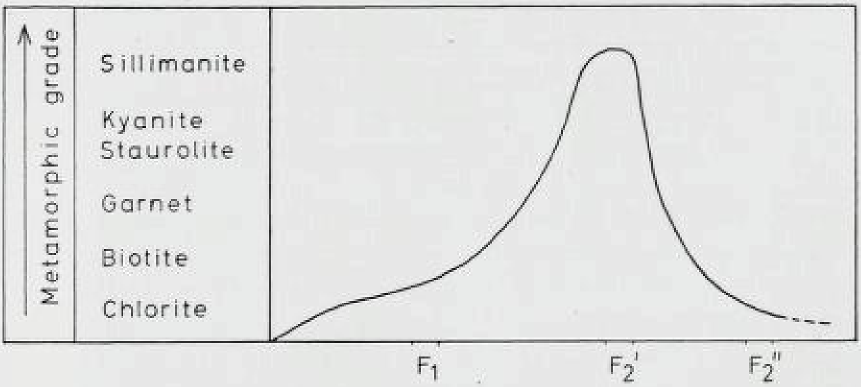

Fig. 74. Diagram to show the variation of metamorphic grade with time in this part of Sørøy.

ocurrences would, in the absence of mineral and whole-rock analyses, seem rather conjectural but not an improbable suggestion. The breakdown of staurolite could also have provided some of the silica and alumina required. The irregular nature of sillimanite development throughout any one hand-specimen and the general rarity of kyanite in those rocks containing an abundance of sillimanite are similar features to those described by Chinner (op. cit.), while the many shear-planes developed during the early-F_2 movements would have facilitated the movement of the requisite fluid phase.

Garnet is present in all the metasediments, sporadically developed in the massive quartzites but frequently profuse in pelites. Its previously noted polyphase crystallisation history is such that it can be regarded as having remained stable throughout the highest grade metamorphic conditions and into the waning post-S_3 phase. Particularly fine examples of post-S_3 garnet growth are to be found in the amphibolite sills (described later). Staurolite, on the other hand, occurs infrequently and textural observations suggest that it crystallised entirely statically, prior to the F_2 movements. It would appear to be present in disequilibrium with both kyanite and the later growth of garnet.

The metamorphic facies controlling the mineral parageneses in the Langstrand–Finfjord metasediments appears to be that of the high-grade part of the almandine-amphibolite facies, namely the sillimanite-almandine-orthoclase sub-facies of Turner and Verhoogen (1960) and Winkler (1965). Since, however, muscovite occurs in several of the pelitic schists and as orthoclase is relatively poorly developed, it may be preferable to assume a paragenesis in the slightly lower sillimanite-almandine-muscovite sub-facies delimited by Turner and Verhoogen (op. cit.) but disregarded by Winkler (op. cit.).

The initial phase of regional metamorphism associated with the F_1 folding is regarded as having been of fairly low grade – greenschist facies conditions – with chlorite probably stable along with muscovite and biotite. Such a mineral association would be indicative of the quartz-albite-epidote-biotite sub-facies. A full study of this metamorphic phase is, however, prohibited by the near-complete eradication of F_1 fabrics by the later metamorphism and tectonism. During the period following the F_1 deformation metamorphic grade rose progressively to a peak marked by the growth of sillimanite late in this static interval. Garnet, staurolite, kyanite, diopside and amphiboles developed as porphyroblasts during this post-F_1, pre-F_2 phase. Concomitant with the acme of the metamorphism a partial feldspathisation and granitisation of the metasediments was initiated. Growth of garnet, kyanite and sillimanite continued, in certain lithologies, into the early stages of the main-F_2 episode of folding (Fig. 73), while further granitisation and the emplacement of pegmatites and aplites also dates from this kinematic phase of the metamorphic history. Local tourmalinisation of schists and an early pulse of scapolitisation in the more calcic lithologies are metasomatic phenomena associated with the granitisation.

That kyanite requires greater pressure for growth than sillimanite at a given temperature is generally accepted and indicated in phase diagrams for the system Al_2O_3–SiO_2. The position of the non-variant point for the polymorphs andalusite, kyanite and sillimanite has been the subject of some recent controversy (cf. Clark et al. 1957, Kharitov et al., 1963, Winkler, 1965), and although the kyanite-sillimanite phase boundary is of principle interest here the location of the triple-point has a decided bearing on estimates of the physical conditions of metamorphism. Winkler (op. cit.) argues strongly against the positioning of the non-variant point at 300°C/8 kb. (Bell, 1963) or 390°C/9 kb. (Kharitov et al. 1963) and suggests 570°C/ca. 7.5 kb. as a more likely point. Recent experimental work by Althaus (1967) broadly agrees with this, his values being 595°C/6.6 kb.; the unvariant curves from this paper are shown in Fig. 75. It will be noted that the boundary of Winkler's suggested stability field of staurolite + quartz barely touches the P-T conditions of Althaus's triple-point but the inconsistencies of

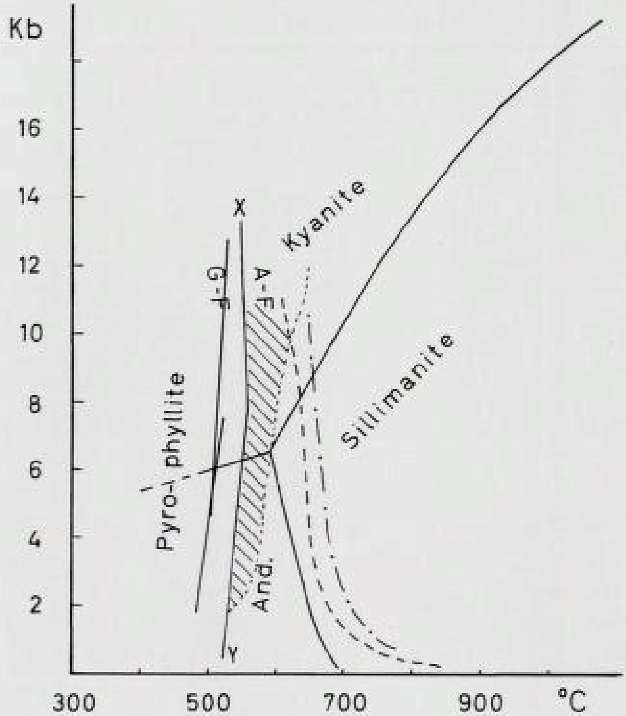

Fig. 75. Diagrammatic representation of Al_2SiO_5 phases. Triple point and unvariant equilibrium curves from Althaus (1966). Line X-Y – boundary between greenschist facies (G-F) and amphibolite facies (A-F). Shaded area – presumed stability field of staurolite + quartz (from Winkler, 1965): — — — — — Tuttle and Bowen's (1958) 'beginning of melting' curve; —·—·—·— 'beginning of anatexis' curve.

this relationships cannot be discussed here. The kyanite-sillimanite unvariant curve is of immediate interest and in view of the coexistence of these two minerals in the Langstrand area without any visible marked disequilibrium and since the initial stages of migmatisation are observed, physical metamorphic conditions of 625°–675°C/ca. 7.5–8 kb. could be conjectured as possibly having obtained during the development of the higher grade mineral assemblages and gneissic rocks of this area.

A general waning of metamorphic grade followed the early-F_2 phase, the major porphyroblastic minerals of this fourth, post-S_3 phase being muscovite, tremolite, garnet and scapolite. The continuing retrogression of the metamorphism then culminated in the diaphthoresis and cataclasis associated with the late-F_2 phase of brittle deformation.

2. Gneisses and pegmatites

A variety of gneissic and pegmatitic rocks are present within the area. In the accounts which follow the salient petrographic and textural features are preceded by only brief notes on structure and distribution.

A. The biotite-hornblende granitic gneiss.

This is the massive gneiss cropping out within the Hellefjord Schist from Lundhavn to Otervikfjell and again along the Finfjordnæringen peninsula (Plate I). The thickness of the sheet-like gneiss generally varies from 25–35 m; along Finfjordnæringen 150 m cliffs of gneiss are present but the true thickness is masked by a profusion of F_2 folds. In the south, on the eastern limb of the Hellefjord Synform, a wedge of sparsely feldspathised metasediment occurs in the middle of the gneissic sheet, the lower half of the gneiss being garnetiferous. The gneiss is markedly foliated and exhibits feldspar augen up to 1.5 cm, rarely 2 cm, across; clots of quartz/feldspar are prominently aligned parallel to F_2 'b' axes. Quartz, plagioclase and microcline together amount to > 75 % of the mineral content (Tables 4, 6 and 7).

Biotite is the chief mafic constituent, the (001) cleavage having a preferred orientation paralleling the foliation. Hornblende, rarely exceeding 8 % of the mode and occurring as ragged prisms $\leqq 4$ mm in length, is often poekiloblastic, the inclusions being mainly small quartz grains. Pleochroism is moderate: α yellow green, β green and γ medium to dark green: $z \wedge c = 26°–28°$.

The plagioclase is a basic oligoclase, An_{26}–An_{30}, in the more massive gneisses. Porphyroblasts up to 6 mm have been observed with biotite deflected around the augen and individual crystals. Sericitisation features may be present, initially either in the cores or along cleavage traces; clinozoisite grains are sometimes developed as a breakdown product. Zoning is uncommon but normal where present. Bent twin lamellae are sometimes prominent.

Microcline frequently displays Carlsbad twinning: microperthite is relatively uncommon occurring as film-microperthite. Where granitisation is advanced a later generation of perthite may sometimes be observed.

Garnet occurs as skeletal individuals, with tensional fractures normal to the foliation quite common. Sphene and apatite are the most notable

Table 3. *Chemical analyses of the 'granitisation' series L.285 A_F.*

	A	B	C	D	E	F
SiO_2	60,91	59,34	59,79	64,31	65,92	60,33
Al_2O_3	18,42	17,17	16,19	15,56	15,98	16,42
Fe_2O_3	0,73	0,77	0,61	0,63	1,15	1,77
FeO	5,40	6,15	6,44	5,94	4,51	7,24
MnO	0,12	0,13	0,11	0,11	0,05	0,14
MgO	2,53	4,81	2,97	1,38	0,50	0,54
CaO	5,74	3,47	6,10	4,73	3,24	4,84
Na_2O	2,03	1,70	2,34	2,39	2,86	3,32
K_2O	2,61	3,42	1,79	2,89	3,40	2,45
H_2O+	1,21	2,50	1,63	0,79	0,90	0,97
H_2O-	0,08	0,10	0,08	0,04	0,09	0,09
TiO_2	0,79	0,86	1,08	1,13	1,08	1,36
P_2O_5	0,18	0,17	0,21	0,21	0,24	0,25
	100,75	100,59	99,34	100,11	99,92	99,72

CIPW Norms.

	A	B	C	D	E	F
Q	19,92	18,06	18,60	23,82	26,70	16,86
Or	15,57	20,02	10,56	17,24	20,02	15,01
Ab	16,77	14,15	19,39	20,44	24,63	27,77
An	27,52	16,68	28,63	21,68	14,18	22,24
Co	2,14	4,69	—	0,51	2,35	0,10
Hy	14,48	21,24	16,77	11,95	6,71	10,77
Mg	0,93	1,16	0,93	0,93	1,86	2,55
Il	1,52	1,67	2,13	2,13	2,13	2,74
Ap	0,34	0,34	0,67	0,67	0,67	0,67

Analyst: Mrs. J. Banham, Bedford College, London.

L 285 A and B Calc-silicate with incipient feldspathisation.
L 285 C Feldspathised calc-silicate schist.
L 285 D Medium-grained garnet-hornblende gneiss.
L 285 E Granitic gneiss.
L 285 F Biotite-hornblende gneiss.

Table 4. *Modal compositions of the 'granitisation' series L. 285 A–F.*

	A	B	C	D	E	F
Quartz	21,4	23,6	21,0	28,6	27,8	29,0
Plagioclase	27,9	33,3	38,7	29,8	34,2	34,5
Microcline	10,2	6,8	3,9	15,1	17,5	8,4
Hornblende	20,0	13,3	15,6	3,3	2,6	7,0
Chlorite	15,1	18,3	15,8	13,0	9,5	x
Biotite	0,4	—	—	3,3	6,7	16,9
Garnet	—	—	—	2,7	—	1,3
Sphene	2,0	1,9	2,8	1,9	1,2	2,2
Clinozoisite	2,3	2,3	1,5	1,4	0,3	0,2
Apatite	0,3	0,3	0,5	0,5	0,2	0,5
Ore	0,2	0,2	0,2	0,3	x	x
Diopside	0,2	—	0,1	—	—	—
Allanite	—	x	0,1	0,2	—	—
Zircon	—	—	—	—	—	0,1
Plagioclase composition	$An_{39.5}$	An_{39}	An_{39}	An_{35}	An_{27}	An_{28}
Hornblende $z \wedge c$	18°	18°	20°	22°	28°	28°

Specimens as in Table 3.
x — Present in minor amounts only.
All plagioclase determinations on U-stage.

accessory minerals. Allanite may be zoned, twinned and surrounded by a corona of subhedral epidote.

Felsic constituents are arranged in a lenticular manner between the often segregatory mafic minerals, and the inequigranular nature of the gneiss is a reflection of its origin since progressive feldspathisation can be demonstrated in a series of specimens ranging from ungranitised amphibole schist to the typical granitic gneiss (see Tables 4, 6 and 7). Together with the porphyroblastesis, a virtually complete recrystallisation of the rock has been effected. The quartzo-feldspathic groundmass is itself inequigranular. Grain boundaries are generally only slightly interlocking with the exception of those of the microcline. Quartz grain boundaries vary from straight to irregular or diffuse; suturing is uncommon. Undulose extinction is a characteristic feature of the quartzes.

Table 5. *Trace elements – 'granitisation' series L. 285 A–F.*
X-ray fluorescent spectrographic determinations.

	A	B	C	D	E	F
Zr	120	193	310	402	385	540
Y	48	52	62	94	73	84
Sr	157	150	310	280	220	278
Rb	85	147	58	71	75	108
Ba	670	825	530	530	580	670
Cr	100	94	87	44	34	30
Zn	268	154	160	104	112	203
Cu	12	11	26	24	tr.	n.d.
Ni	83	75	56	30	23	17
Nb	tr.	10	tr.	9	12	25
V	124	137	117	82	77	69

Values expressed as p.p.m.
tr. — only a trace present
n.d. — not determined.
All Specimens as in Table 3.
Analyst: Dr. B. A. Sturt, Bedford College, London.

Symplectitic intergrowth of vermicular quartz in host plagioclase feldspar produces a common myrmekite texture which can be observed in all stages of development. Myrmekite is normally seen in association with microcline (Fig. 6) and the more advanced stages of this texture are found in the microline-rich sectors of the gneiss. It is manifest that the myrmekitic plagioclases are being absorbed or replaced by the growing potassic feldspar. Not uncommonly relics of this plagioclase occur within microcline porphyroblasts, and examples are present where the K-feldspar has totally replaced the plagioclase crystal leaving the quartz vermicules as a relict myrmekitic texture within the later porphyroblast. An interesting feature of some myrmekite lobes is the development of a rim of more sodic plagioclase at the margin of the oligoclase grain. These rims rarely exceed 20 μ in thickness and may be discontinuous. Where Becke lines are observed the rim always has a lower R. I. than the parent oligoclase, suggesting a possible sodic-oligoclase or even albitic composition. Twin lamellae extinction

Table 6. *Modal compositions of the 'granitisation' series D. 286 A–F.*

	A	B	C	D	E	F
Quartz	32,8	25,9	28,1	29,5	24,1	39,5
Plagioclase	32,8	32,0	29,1	35,1	29,9	35,2
Microcline	0,1	13,6	16,7	17,8	25,4	20,7
Biotite	19,5	19,3	19,4	15,3	14,2	1,6
Hornblende	6,1	5,9	2,5	—	4,0	—
Diopside	5,0	—	—	—	—	—
Garnet	—	—	1,5	—	0,8	0,3
Muscovite	—	—	—	—	—	2,5
Sphene	0,6	1,5	1,7	1,1	1,1	—
Clinozoisite	0,1	1,3	0,2	0,3	0,3	0,3
Apatite	0,2	0,3	0,5	0,5	0,3	—
Allanite	x	0,1	—	—	x	—
Scapolite	2,5	—	—	—	—	—
Ores	0,2	x	0,4	0,2	0,1	x
Zircon	x	x	0,1	0,1	x	—
Zeolite	—	0,2	—	—	—	—
Chlorite	—	x	—	—	—	—
Plagioclase composition	$An_{38.5}$	An_{38}	An_{36}	An_{31}	An_{28}	—
Hornblende z $\wedge$ c	24°	24°	26°	—	28°	—

D. 286 A–Calc-silicate schist: D. 286 B-D — schists with progressive feldspathisation (D-gneissic): D. 286 E — coarse hornblende-biotite gneiss: D. 286 F–Pegmatitic gneiss.
x — Present in minor amounts only.
All plagioclase determinations on U-stage.

positions become reversed on passing from the normal basic-oligoclase into the sodic rim.

Plagioclase porphyroblasts are developed at the expense of the pre-existing quartzo-feldspathic mozaic. Microcline is normally absent in the calc-siliceous schists and is essentially a product of the granitisation. It is seen that the K-feldspar has invariably developed later than the porphyroblastic plagioclase, many examples of microcline or microcline-microperthite replacing and digesting plagioclase being observed. As the granitisation series D. 286A-E (Table 6) is traced along the strike,

Table 7. *Modal compositions of feldspathisation series.*

	1 LR.24	2 LR.25	3 LR.26	4 LR.87	5 LR.88	6 LR.89	7 D.97
Quartz	23,08	28,31	31,54	27,06	31,66	32,87	23,2
Plagioclase	35,55	37,10	30,54	32,63	33,56	31,03	30,2
Microcline	1,73	5,36	21,84	1,44	10,34	15,23	21,7
Biotite	22,76	15,25	10,73	18,92	18,80	16,71	14,3
Hornblende	11,76	9,08	3,76	13,92	3,42	2,88	8,6
Diopside	—	—	—	4,72	—	—	—
Garnet	0,86	0,28	0,13	—	0,06	—	0,9
Sphene	3,40	2,21	1,03	0,65	1,66	1,14	0,7
Apatite	0,40	0,36	0,11	0,13	0,26	0,11	0,2
Clinozoisite	0,05	0,39	0,27	0,03	0,02	—	—
Ores	0,19	0,14	0,04	0,28	0,22	0,05	—
Chlorite	—	0,26	—	x	—	—	—
Allanite	—	—	—	—	—	x	0,2
Scapolite	—	—	—	0,23	—	—	—
Plagioclase composition	An_{39}	An_{27}	An_{21}	An_{39}	An_{34}	An_{30}	An_{26}
Hornblende z ∧ c	20,5°	23°	20°	20°	24°	22°	26°

x — Present in minor amounts only.
LR. 24–26 – Feldspathisation series.
LR. 87–89 – Feldspathisation series.
D. 97 – Massive hornblende-biotite granitic gneiss.
Plagioclase determinations on U-stage.

it is noted that, initially, plagioclase porphyroblasts are developed far in excess of K-feldspar. The plagioclase is andesine, An_{38-39}, becoming progressively more sodic with increasing porphyroblastesis and granitisation until oligoclase, An_{27-28}, is attained. This change of plagioclase composition is in agreement with the increase of Ab. content with granitisation noted by Engel and Engel (1958) in the Adirondack Mountains. Moreover, a proportionate increase of Na_2O with granitisation is noted in the chemical analyses of the series L. 285A-F from Langstrand (Table 3).

Fig. 76. Myrmekite partly engulfed by large microcline. Biotite-hornblende granitic gneiss. Crossed polarised light, x35.

Apart from distinct porphyroblasts, microcline also occurrs less preminently as amoeboidal growths in the coarsely grained gneissic groundmass fabric, wrapping around and including grains of quartz and plagioclase. Microcline also appears in segregations along with quartz; grain boundaries between microclines are slightly interlocking, seldom sutured; quartz is always strained. Discontinuous – as distinct from undulose – extinction in quartzes occurs where grains are divided into sectors sub-grains, each of which displays its own undulose extinction: no true fractures can be detected between the sub-grains however. K-feldspar is also present as microperthite which appears to represent a slightly later generation of this feldspar.

From the various evidence of field occurrence, relationships and textural studies, a replacive origin for this (garnet-) biotite-hornblende granitic gneiss is postulated. The chief criteria favouring such a genesis are:

(a) Progressive feldspathisation and porphyroblastesis as seen in the field (Fig. 77 and 78).

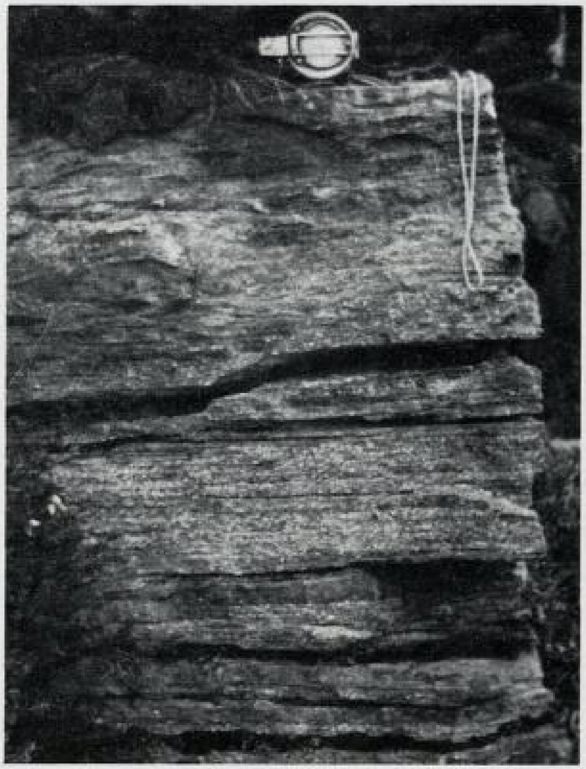

Fig. 77. Partially feldspathised Hellefjord Schist – from the wedge of meta-sediment within the (divided) gneiss sheet, 1.5 km N. of Hellefjord.

(b) Increase of grain size in transitions from schist to gneiss, sometimes along the strike.
(c) Stringers and bands of metasediment are undisturbed within the gneiss and show various stages of recrystallisation and feldspathisation.
(d) Perfect conformity of the gneiss to the banding in the metasediments.
(e) Considerable variation of grain size of a mineral species.
(f) Compositional variation of plagioclase with increasing feldspathisation.
(g) Variations in the modal compositions.
(h) Notable increase of potash feldspar with transition from schist to gneiss and with increasing coarseness of general grain size.

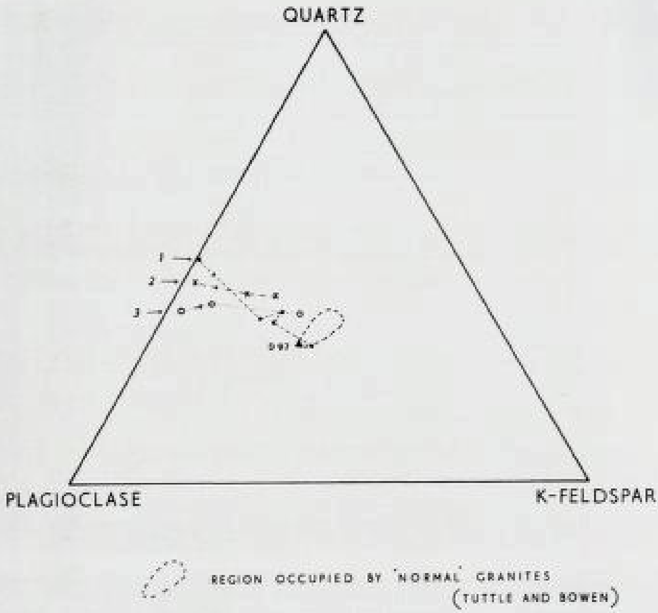

Fig. 80. Plot of the modal analyses of the granitisation series D. 286A-E (No. 1), LR. 87–89 (2) and LR. 24–26 (3). Specimen D 97 (massive gneiss) also included.

The formation of the gneiss was clearly initiated prior to the main F_2 movements but it is demonstrable that part of the K-feldspar growth continued syn-kinematically: larger K-feldspars not infrequently contain the S_3 biotite. That the granitisation and metasomatic processes were closely associated with the highest grade or grades of regional metamorphism appears incontrovertible.

Chemical and modal analyses of the granitisation series L. 285A–L. 285F (taken from the gneiss locality near Langstrand and not wholly identical to the massive gneiss further NE) show several features of interest (Tables 3, 4 and 5) despite the presence of diaphthoretic phenomena and the fact that the samples were collected across the strike. Specimen L. 285E represents the most typical gneiss. Significant general increases in silica and soda are noted whilst alumina, magnesia and lime show decreases with granitisation. Potash varies irregularly but shows an overall increase, agreeing with the modal increase of microcline. A possible relationship is that of MgO with chlorite; magnesia decreases proportionately to the decrease of modal chlorite. Of the

Fig. 78. Biotite-hornblende granitic gneiss with metasedimentary banding still prominent. This represents a more advanced stage in the granitisation process than that depicted in Fig. 77. S.W. Lundhavn.

Fig. 79. Massive biotite-hornblende granitic gneiss deformed by F_2 folds. Locally, granitic veinlets occur along the minor fold axial planes. Veiviken.

microcline together constitute over 85 % of the modal composition (Table 8) with muscovite the only other essential mineral.

Plagioclase, usually the dominant feldspar in the several variants of gneiss examined, is an oligoclase (An_{16-24}).[1] An irregular cloudiness is seen in many plagioclase grains and sericite may be developed. Sericite often displays a preferred orientation along cleavages or even certain twin lamellae. Where plagioclase is in contact with potash feldspar, a sodic rim is frequently present. Myrmekite is irregularly developed within plagioclase crystals adjacent to microcline.

Microcline generally forms xenoblastic grains of variable size ranging from isolated grains in relict quartzite ribs to porphyroblasts sometimes containing partially digested plagioclase. Porphyroblasts (up to 3 mm) are uncommon, the usual microcline grain size in the order of 1 mm; this is particularly so in microcline segregations. Microcline microperthite is infrequently developed but tends to be closely associated with those gneissic specimens which show the most intense quartz deformation features. This accords with Chayes (1952) observations and suppositions that shearing stress promotes the exsolution of albite from the host potash feldspar into optically resolvable microperthite.

Quartz occurs interstitially but most notably in lenticular segregations wherein anhedral grains may exceed 5 mm. It is always strained sometimes to the extent of producing an anomalous biaxial interference figure; Boehm deformation lamellae may be discerned. Sub-grains are distinguishable, again lacking fractures at their boundaries suggesting the presence either of annealed boundaries or a stage of lattice dislocation prior to actual fracture. Suturing of quartz grain boundaries is uncommon, there being a large variation in boundary shape.

The general inequigranular texture is accentuated where quartz is large and in lenticular segregations: feldspars also occur in roughly lenticular segregations paralleling the foliation. Microcline, where segregatory, develops a granoblastic mozaic with straight grain boundaries showing good triple junctions (often subtending angles of 120°). Elsewhere microcline boundaries are irregular, as are those of oligoclase and quartz grains. Suturing of grain boundaries is rarely extensively developed in these gneisses. The planar orientation of muscovite, and less commonly biotite, defines the foliation as do the elongate aggregates of quartz and feldspar. Muscovite may occasionally develop a vermi-

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[1] U-stage determinations.

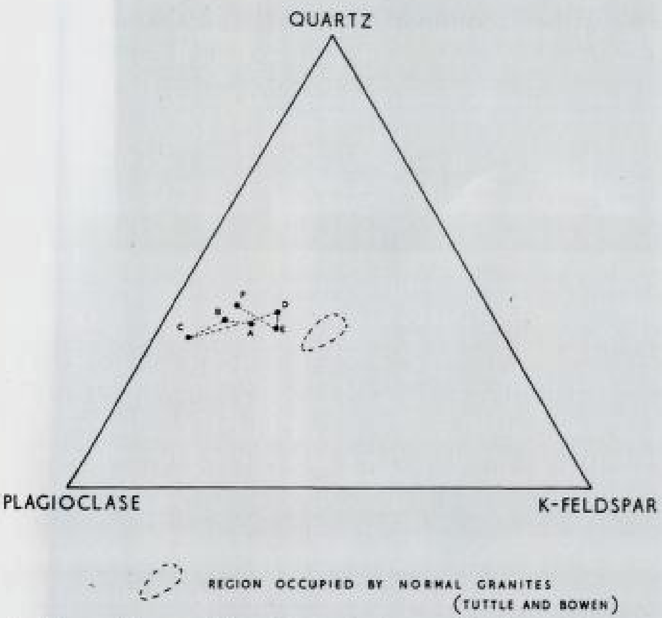

Fig. 81. Plot of the modal analyses of the granitisation series L. 285A-F.

trace elements, decreases of Ni, Cr and V are notable while Zr shows a marked rise towards the granite gneiss.

The series D. 286A–E, LR. 24–26 and LR. 87–89 are better indicators of the granitisation. A triangular variation diagram of these series (Fig. 80) depicts significant trends towards the 'normal' granite field, the coarse gneiss D. 286E and solitary specimen D. 87 lying on the margin of this field. While the members of series L. 285 A–F are more irregularly distributed (Fig. 81) a statistical trend indicates movement towards the 'normal' granite.

B. Granite gneiss.

This is a leucocratic, coarse-grained, quartzo-feldspathic gneiss occurring as concordant partings or irregular lenticular bodies within the Klubben Quartzite Group, mostly in the northern half of the area. Gradational boundaries are typical of this gneiss and often a preferential feldspathisation of the pelitic intercalations in the metasedimentary sequence can be observed: relatively unaltered quartzitic ribs are frequently present within the massive gneiss. Quartz, plagioclase and

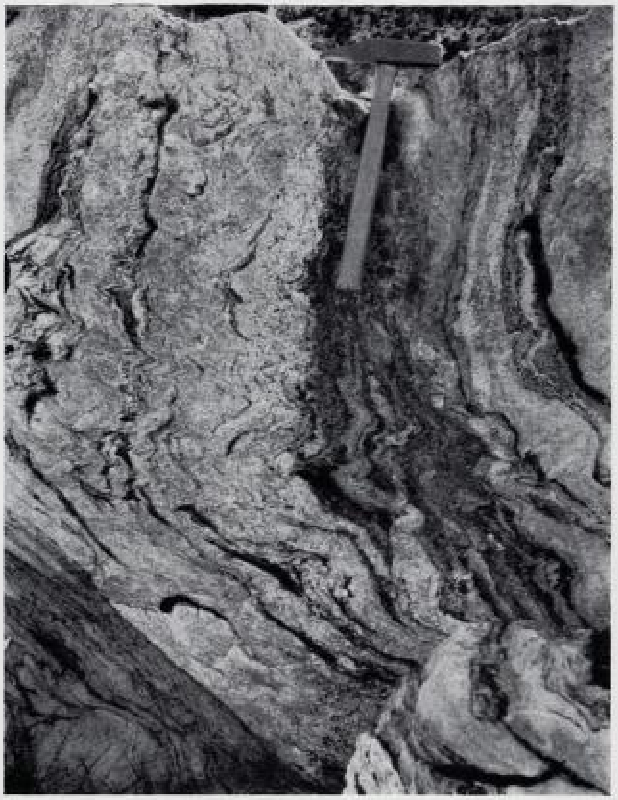

Fig. 82. Granite gneiss in mixed psammite-pelite sequence with relics of metasediment quite prominent in the gneiss. Minor F_2 folds abundant. Shore of S.E. Hønsebyfjord.

Fig. 84. Tourmaline pegmatite with narrow aplitic margin. Hellefjord Schist below. 200 m NE of Hellefjord.

Table 8. *Modal compositions of granite gneisses.*

	1 L. 38	2 SR. 106A	3 SR. 334	4 SR.643	5 SR. 110	6 SR. 204	7 LR. 43	8 SR. 109	9 SKR. 3	10 SKR. 4
Quartz	32,7	33,4	26,4	38,2	32,8	27,4	27,9	29,9	40,6	21,7
Plagioclase	48,4	47,3	17,7	18,8	31,0	50,9	37,2	40,8	25,1	36,0
Microcline	14,4	15,1	52,2	36,8	23,0	13,1	28,2	17,7	22,4	33,4
Muscovite	1,4	3,9	3,8	3,0	10,4	6,8	5,4	2,8	3,1	1,4
Biotite	2,9	—	—	—	1,1	—	—	7,9	8,3	7,3
Garnet	—	0,3	—	2,8	1,7	0,5	—	0,8	—	—
Apatite	0,2	—	x	0,1	0,1	1,3	0,7	x	0,4	0,1
Ores	—	—	—	0,2	—	—	—	0,1	0,2	0,1
Tourmaline	—	—	—	—	—	—	0,6	—	—	—
Plagioclase composition	An_{23}	An_{25}	An_{24}	An_{23}	An_{22}	An_{22}	—	An_{24}	—	—

x — Present in minor amounts only.

1–8 – Various gneisses : 9 – Slightly feldspathised mica quartzite : 10 – Granitised mica quartzite (granite gneiss).

Plagioclase determinations on U-stage.

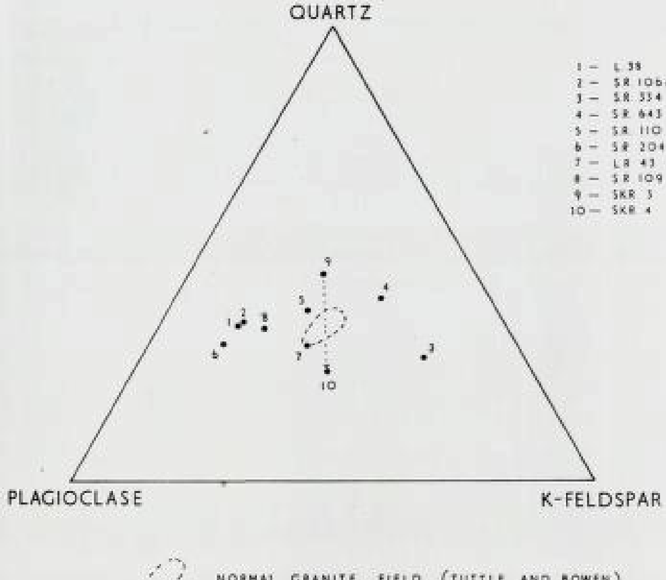

Fig. 83. Plot of the modal analyses of granite gneisses. Arrow indicates granitisation trend between two specimens (see Table of modal compositions).

cular form with quartz the guest mineral. The quartz intergrowth is often initiated along cleavages of the mica so that a semi-trellised pattern may be developed. Either the whole or a part of the intergrown quartz is optically continuous and often this continuity extends into quartz grains adjacent to the mica.

Garnet where present is rather skeletal. It has been severely deformed and often comminuted by the F_2 deformation into long trains of garnet grains and granules in the plane of the foliation. Its crystallisation, initiated prior to the onset of feldspathisation, probably continued into the granitisation period.

Both field and petrographic evidence lend substantial support to the view that the granite gneiss has originated by a process of granitisation of the metasediments (Fig. 82). On a megascopic scale various transitions from psammite (with pelite bands) to granite gneiss are observed, with a notable preferential granitisation of the less quartzose horizons. In thin-section progressive granitisation features and textures are demonstrable. Quartzite ribs, devoid of any appreciable feldspathisation, indicate their relative resistance to the process of metasomatic diffusion. The granitisation of the metasediments occurred immediately prior to the onset of the F_2 deformation, and is known to have continued locally into the early-F_2 phase.

C. Pegmatites and aplites.

Both these rock-types occur throughout the area although they are never extensively developed. In some pegmatitic bodies the border zones are aplitic. Field evidence shows that aplites and pegmatites were formed before and during the F_2 tectonism.

1. *Pegmatites.*

The pegmatites are extremely coarse-grained quartzo-feldspathic rocks usually containing muscovite. Accessory minerals include biotite, tourmaline, apatite, garnet and zircon while beryl, magnetite and chlorite occur only rarely. Tourmaline is common locally, sometimes as prismatic crystals up to 3 cm in length. In the marginal zone of a pegmatite near Hellefjord, tourmaline individuals tend to be oriented normal to the sharp, concordant margin (Fig. 84).

Normally the pegmatites are concordant with the banding, though occurring as pinch-and-swell structures, boudins and tectonic inclusions. Those associated with the Lower Limestone sometimes have a sheet-like form and may be cross-cutting, but it can be seen that this is mainly a tectonic effect. Pegmatites also occur along minor slide planes (Fig. 44). Thin pegmatite dykes of a totally discordant nature are sometimes observed. These may change from concordancy to discordancy within one exposure (Fig. 85) and are noted to be non-dilational – the metasediment banding is not offset by the emplacement of the pegmatite dyke. This implies volume-by-volume replacement, as opposed to forceable intrusion, as the mechanism responsible for the pegmatite emplacement (King, 1948).

The Veiviken pegmatite body contains many lenses and 'rafts' of metasediment. These are never visibly disturbed or rotated and quite often display marginal basification features; in some instances a narrow garnet-rich zone is present at the junction with the pegmatite. In pegmatites in east Saksfjord, abundant streaks, aggregates and lenticles of biotite occur in the marginal zones, a feature noted on a lesser scale in other pegmatites.

Coarseness of grain in the pegmatites varies appreciably. On the Saksfjord east shore, one feldspar measured 47 cm (Fig. 86). This crystal contains slivers of quartz along two cleavage directions. Occa-

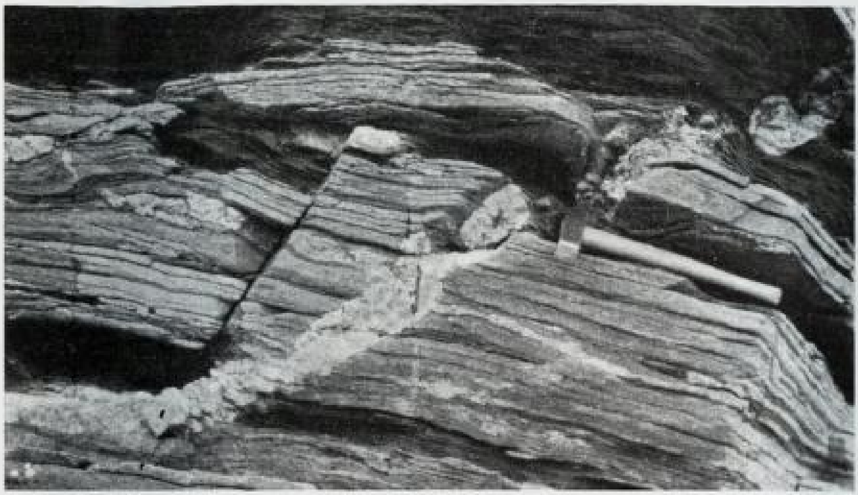

Fig. 85. Non-dilational pegmatites in Hellefjord Schist. S.E. Finfjord.

Fig. 86. Large feldspar in pegmatite. East coast of Saksfjord.

sionally, runic textured pegmatites are observed (Fig. 87); this has the appearance of a close intergrowth of feldspar and quartz, the latter showing shapes controlled by three cleavage directions in the host feldspar.

Considerable variability of textural and mineral relationships from specimen to specimen is typical of the pegmatites. Grain boundaries, for

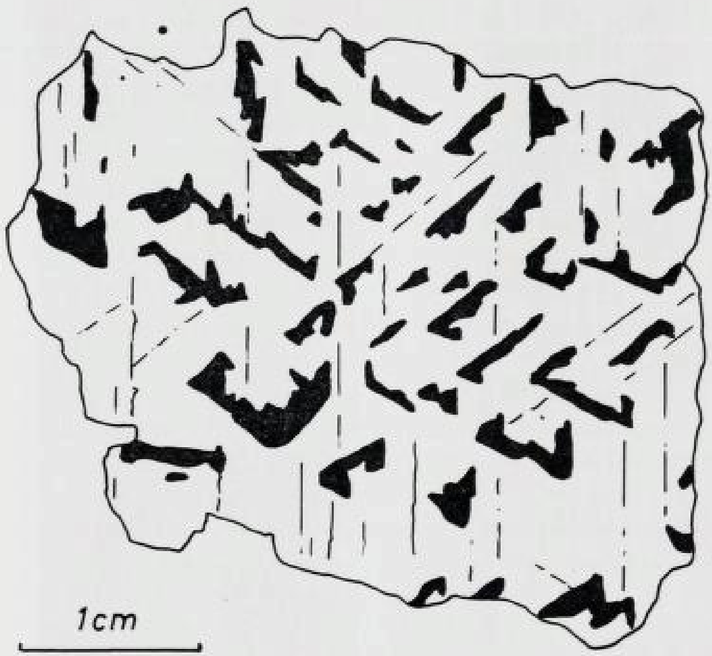

Fig. 87. Runic texture in pegmatite – quartz (black) in microcline (sketched from hand-specimen). 200 m N.W. of Veiviken.

example, vary in irregularities and curvature and may ultimately be complexly sutured. Quartz and oligoclase (An_{14}–An_{20}) are the main minerals with muscovite and K-feldspar generally subordinate. Plagioclase appears to post-date muscovite growth – relict fragmentary flakes of muscovite within large oligoclase grains may display optical continuity denoting that they are parts of an originally larger flake. Plagioclase breakdown products are similar to those noted in the gneisses.

Microcline is generally later than plagioclase and while never being markedly poekiloblastic, contains all other minerals as inclusions. Myrmekite texture is common in plagioclase adjacent to K-feldspar and lower R.I. albitic rims are frequently developed. Microcline microperthite is present in a few pegmatites.

Deformation features are common throughout the pegmatites. Such features include strongly strained quartzes, fractured and bent plagioclase twin lamellae and the presence of microscopic shear planes. Micas are often strongly deformed with twin and translation gliding indicative of lattice deformation. Quartz-quartz boundaries may be variably granulated. Three textural varieties of quartz are distinguished – small relatively equant groundmass grains, large interstitial quartz and the later recrystallised granulate type (e. g. in mortar structure).

The pegmatites are regarded as replacement bodies produced as the result of alkali metasomatism during high grade metamorphism. Evidence for this assumption is seen in a number of features, principally the development of 'seed' lenticles of pegmatite in schists with further pegmatisation giving concordant streaks and bands, and secondly the non-dilational nature of cross-cutting pegmatite dykes. Other criteria favouring 'pegmatisation' as opposed to intrusion include: (i) relict banding and schistosity preserved within the pegmatite margins – this may be traced by, e. g., strings of garnets which grow preferentially in pelite-rich bands in the metasediment: (ii) diffuse local pegmatisation within gneisses: (iii) progressive feldspathisation and microscopic evidence of gradational uneven contacts and (iv) marginal basification features in the adjacent host rock. The alkali metasomatism is thought to have operated in two waves with potash slightly later than soda. This is identical to the features recognised in the granitisation processes described earlier.

Field evidence also shows the pegmatites to have been emplaced both prior to and during the F_2 deformation, but even those occurring along F_2 fold axial planes were appreciably deformed by later pulses of this protracted deformation episode. Boudinage-type deformation of pegmatites occurring along fold axial planes normally implies post-crystalline movement but it is not impossible that some such features are similar to Ramberg's (1956) 'concretion pegmatites', in which pinch-and-swell structures have developed contemporaneously with the growth of the pegmatite.

2. *Aplites.*

Aplites, compositionally related to pegmatites but distinctive texurally, are found principally as buff-coloured veins or thin dykes and sills. As with the pegmatites, aplites are both pre- and syn-F_2 in age.

Feldspar, either oligoclase (An_{14}–An_{18}) or microcline, is the dominant mineral; one feldspar may be present to the near-exclusion of the other while quartz is often quite subordinate. Accessory minerals include muscovite, tourmaline (schorlite), apatite, garnet, zircon, biotite and magnetite. An equigranular texture is the prominent feature of the

Fig. 88. Non-dilational aplite veins parallel to axial planes of F_2 folds. The banding in the psammite can often be traced through the aplites. 600 m SSW of Hønsebynes.

aplites, porphyroblasts being uncommon. The normal feldspar grain size is ≤ 1 mm. Many of the textural and mineral relationships observed in the gneisses and pegmatites are present in the aplites, notably myrmekite, albitic rims and vermicular muscovite. Feldspar grain boundaries vary from straight to interlocking, and in many specimens the fabric is polygonal with triple junctions subtending angles of ca. 120°.

Within the psammites, all gradations from mildly feldspathised metasediment to typical aplite can be demonstrated. The syn-F_2 dykes or veins of aplite are observed to be of a non-dilational nature (Fig. 88), there being no offsetting of the metasedimentary banding. Thus, both pre- and syn-F_2 aplites have developed by a volume-by-volume replacement mechanism similar to that described for the pegmatites.

3. *Greisen.*

The occurrence of greisen associated with the pegmatites is extremely localised. It is present within and marginal to the pegmatite following the outcrop of the Lower Limestone, but then only as diffuse areas or veinlets. Greisen consists solely of quartz, books of white mica and some tourmaline. A transition into pegmatites is traceable.

D. Microcline-garnet lenticles.

Several red-weathering lenticular bodies of garnetiferous microcline schist or gneiss are present within the psammites of Saksfjell, their boundaries being of a markedly gradational nature. The groundmass fabric of this rock is granoblastic with equant quartz and feldspar and scattered oriented micas. Lensoid segregations of microcline are common and represent a later generation of K-feldspar crystallisation.

Garnets are usually ≤ 2 mm in diameter and contain many small inclusions of quartz, microcline and plagioclase (An_{20}–An_{22}) which are thought to represent the fabric of the metasediment at the time of garnet growth. The outer zone of garnets may be broken down and replaced by the larger grain groundmass fabric of quartz, microcline and green Fe-rich biotite. Three generations of microcline are therefore present.

Field and petrographic evidence clearly points to a metasomatic origin for these rocks the localisation of microcline enrichment perhaps related to an original local preponderance of K-feldspar – many horizons within the quartzite sequence are arkosic.

Another variant of this microcline enrichment on the Saksfjord plateau is seen in the form of 0.3–1 m thick sills. Garnets are more profuse than in the above-described gneiss, ranging up to 1 cm in diameter, and enclose an abundant and finer grained pre-garnet fabric. A significant observation is that microcline is totally absent from this inclusion fabric while sphene is not infrequently present. Microcline is common in the groundmass but is clearly late. Quartz-microcline lenticles are also abundant in the more strongly feldspathised lithology and both these minerals may occur as pressure shadows around garnet. Plagioclase occurs throughout the groundmass fabric and traces of degraded hornblende are quite common.

Both field and mineralogical observations indicate that this lithology represents a feldspathised garnetiferous amphibolite sill, the presence of appreciable plagioclase and some hornblende and total lack of microcline in the garnet inclusion fabric being critical in this interpretation.

E. Quartz veins, pods and segregations.

Secondary quartz is found in a variety of forms – veins, boudins, lenticles, rods and at the nodes of boudins, to mention a few. The vast majority of quartz veins and segregations have been deformed during the F_2 fold episode, as evidenced by both field and petrographic observations, and it would appear that the origin of much of this quartz is related primarily to the period of highest metamorphic grade. Such quartz is visualised as being derived as metamorphic segregations sweated out from the metasediments during the episode of high temperatures. That quartz also crystallised segregatorily during the early-F_2 phase is seen in the presence of pressure shadows of this mineral around rotated porphyroblasts. It occurs, too, as veinlets along F_2 axial plane shears and minor slides.

In some instances pegmatite veins show a trend towards quartz-rich fractions, the ultimate development of such veins containing only sporadic feldspars.

3. The basic rocks

A. Metagabbro sheets.

Lenticular or irregular bodies of coarse amphibolite of outcrop length from ca. 1 km down to 2–3 m and maximum thickness 18m are fairly common within the Hellefjord Schist. They are regarded as tectonic lenses of a pre-existing gabbro sheet or sheets related to the Storelv Gabbro of central Sørøy (cf. Stumpfl and Sturt, 1965). Typically the rock is dark, medium- to coarse-grained and exhibits a notable amphibole lineation (F_2) and crude schistosity (S_3): mesoscopic F_2 folds are present locally. Marginally the metagabbroic bodies are of finer grain and appreciably sheared. The coarser central parts of these bodies often display a knobbly weathered surface, the white feldspar outstanding and the mafic minerals weathered out as pits. Contact metamorphic effects are not seen, this being attributed to the intensive prolonged deformation and recrystallisation and the improbability of such bodies being in an original juxtaposition with the surrounding pelites.

Amphibole and plagioclase constitute the bulk of all specimens examined. Sphene, ilmenite (or Ti-magnetite) and apatite are commonly present while diopsidic augite, rutile, clinozoisite and biotite occur rather variably: accessory minerals include chlorite, scapolite, calcite, allanite, zircon, pyrite and limonite. Quartz, while absent in the coarser metagabbro, occurs in subordinate amounts in the sheared marginal zones.

The amphibole is hornblende, present as prismatic crystals up to 4.5 mm in length. Pleochroism is weak to moderate; α pale yellow-green, β yellow-green, γ green or brownish green: $z \wedge c = 22^\circ$–25°. A preferred lineation of crystallographic c-axes parallel to F_2 'b' is conspicuous. In the coarser lithology, hornblende is often present as crystal aggregates with the intervening felsic areas containing only sporadic amphibole. Intergrowth of hornblende and plagioclase at times displays a recognisably relict ophitic or sub-ophitic texture; this igneous texture is better seen where pyroxene is present.

Plagioclase generally forms a mozaic of fairly equant grains rarely > 1 mm across except in porphyroblasts. It is andesine of composition range An_{36}–An_{42}. Sporadic sodic labradorite (An_{48}–An_{54}) occurs in the form of degenerate igneous feldspar (Fig. 89), while later porphyroblastic plagioclase is in the range of sodic andesine, An_{30}–An_{35}. In Fig. 89 the early partially sericitised feldspar is being replaced by the recrystallised polygonal plagioclase fabric; the degraded plagioclase is here An_{48} and the later feldspar An_{39}. Normal zoning is not uncommon in the plagioclases, anorthite content decreasing by $\leq 6\,\%$ from core to margin in some crystals.

Biotite, which may be absent in the coarser metagabbro, increases in amount towards the schistose marginal lithology. It is pleochroic from very pale yellow to orange- or foxy-brown. The biotite is formed primarily at the expense of amphibole and various stages in this replacement are noticeable; initially the biotite develops along cleavage traces or at the margins of hornblende laths.

Diopside or diopsidic augite is xenoblastic and rarely occurs in an unaltered state. It is extensively replaced by amphibole. The pyroxene is colourless and biaxially positive with $c \wedge \gamma$ 40°–42°. Schiller inclusions, including thin laminae of ore (? ilmenite), are common these having often been retained by the replacive amphibole. A notable observation is that the pyroxene-bearing amphibolites are restricted to the schistose marginal zones of the metagabbro bodies.

Sphene is ubiquitous. It occurs as grains, grain clusters or trains of granules (synneusis texture) which parallel the foliation. Frequently sphene is present as coronas around ilmenite or sometimes rutile. Rutile may nucleate around ilmenite so that occasionally the growth sequence

Fig. 89. Relict igneous plagioclase (An_{48}) surrounded by recrystallised polygonal plagioclase (An_{39}). Metagabbro, Finfjorddalen. Crossed polarised light, x25.

ilmenite→rutile→sphene may be observed. Rutile is usually present only in the quartz-deficient amphibolites.

Clinozoisite is mostly confined to the marginal amphibolites and usually occurs as vermicular masses replacing hornblende, sometimes pseudomorphing the amphibole prisms. It may also grow out from the hornblende into the andesine. Occasional examples of clinozoisite rimming diopside can be found. In thin-sections containing vermicular clinozoisite, the plagioclases are often cloudy or incipiently saussuritised. Clinozoisite also occurs as discrete crystals in feldspar-rich pods. A third occurrence is that of subhedral grains with cores of allanite.

Scapolite, present as an alteration product of hornblende, diopside or andesine, occurs in the schistose marginal rocks. It appears to have developed mainly from plagioclase and is observed in two forms:

(i) irregular, highly poekiloblastic, symplectitic intergrowths with feldspar; grains up to 1 mm.
(ii) a mozaic of fairly equant grains; rarely $>$ 0.6 mm.

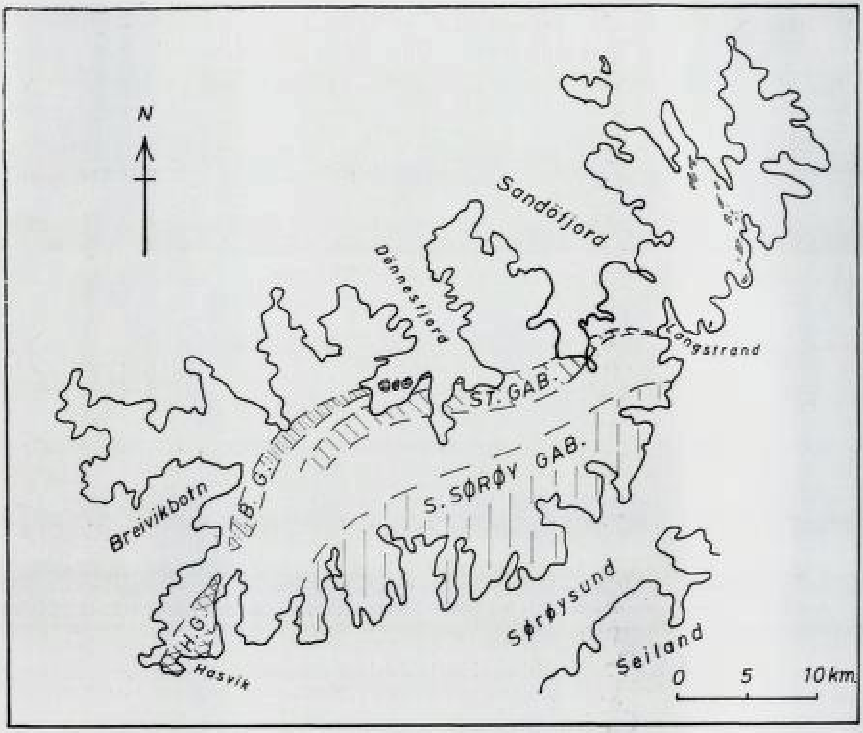

Fig. 90. General distribution of the gabbroic rocks on Sørøy. B.G. – Breivikbotn Gabbro; St. Gab – Storelv Gabbro; H.G. – Hasvik Gabbro. S. Sørøy Gabbro drawn from N.G.U. 1 : 1,000,000 map.

Calcite is a late product of the breakdown of diopside, hornblende, plagioclase and possibly scapolite and also occurs in the sheared marginal lithology.

Hornblende lineation is notably less prominent in the coarser metagabbros than in the schistose amphibolitic margins. Nevertheless, the larger prisms commonly reveal a parallel arrangement of the (100) planes containing the crystallographic b-axes, this contributing to the foliation. Mechanical breakdown and recrystallisation of these larger hornblendes is observed although this tends to be marginal and not affect the whole crystal. The recrystallised fragments are invariably aligned with their c-axes parallel to the tectonic F_2 'b' direction. Zoned amphiboles have been noted in marginal amphibolites, particularly those containing abundant vermicular clinozoisite, in which the normal hornblendic core of a prismatic crystal is surrounded by an almost colourless non-pleochroic outer zone. No extinction angle difference can be detected between core and periphery. Often hornblendes are 'bleached' immediately adjacent to vermicular clinozoisite.

Mineralogical and textural examination has shown that these lenticular foliated amphibolites were originally gabbroic in composition. Although relatively little evidence of the original igneous nature of these bodies remains, it can be recognised in the relict sub-ophitic texture, traces of the igneous feldspar (now variably saussuritised) and the localised presence of pyroxene with schiller inclusions. The similarity of mineral content and textural and field relationships of the various lenticles of metagabbro would suggest that they have had a similar history and origin. They have been metamorphosed, strongly recrystallised and deformed by the F_2 folding, and are regarded as tectonically transported and sheared derivatives of a pre-existing gabbro sheet or sheets.

On comparing the Storelv Gabbro of central Sørøy (Stumpfl and Sturt, op. cit.) with the metagabbro lenses of the writer's area, many features of mineralogy and textures are common to both. The Storelv Gabbro displays strongly sheared and schistose margins and towards the east the sheet thins out (Fig. 90) near Lottre on the west of Langstrandfjord where it is seen as hornblende schists and hornblende-biotite schists. It is therefore highly probable that the lenticular metagabbroic bodies of the present area represent sheared tectonic lenses of the Storelv Gabbro.

B. Basic sills and dykes.

Concordant or gently transgressive amphibolite sheets varying from 15 cm to 3.5 m in thickness are common within the Klubben Quartzite Group (Fig. 8). They are schistose (S_3) and of medium-fine grain, clearly deformed by F_2 folds and may be richly garnetiferous (Figs. 30 and 43): many of them have finer grained biotite-rich margins. Shearing affects the sheets to varying degrees, with F_2 axial plane shears frequently causing them to be disrupted (Fig. 42).

Garnets are of variable occurrence and size, gradations from amphibole schist to biotite schist usually being correlated with the increasing presence of garnet. Although no one garnet distribution pattern within sills or dykes can be said to be typical, a noteworthy pattern is the presence of abundant small garnets in the marginal zones with sporadic large garnets (up to 4–5 cm) in the central part. In many sills however, garnets are distributed quite irregularly.

All gradations in the mineralogy and textures have been observed between the hornblende schists and the garnet-biotite schists. Hornblende schists are characterised by a nematoblastic textural arrangement of plagioclase and amphibole with the latter preferentially aligned parallel to the F_2 'b' direction. Common accessory minerals include sphene and iron ore, the ore grains dominantly magnetite or Ti-magnetite. Garnet, apatite, clinozoisite, ilmenite, rutile and allanite are of variable occurrence. The amphibole, up to 1.5 mm in length, is only moderately pleochroic in shades of pale green and is frequently seen to be replaced by biotite.

Plagioclase has a compositional range of An_{27}–An_{36} although andesine of An_{40} has been recorded. It normally occurs as equant, polygonal or slightly elongate grains rarely > 0.4 mm, with triple junctions abundant, although irregular plagioclase segregations devoid of ferromagnesian minerals are noted in some sills. A third occurrence of plagioclase is that of relict, degraded phenocrysts in the central parts of the prominent 3–4 m thick transgressive sill north of Lundvatn. These are up to 5 mm across, twinned on the albite, Carlsbad and pericline laws and strongly altered to clinozoisite, calcite, sericitic aggregates and, rarely, zeolite. Zoning is present in the recrystallised polygonal plagioclase but is usually of reverse type; this is thought to be closely connected with rising temperatures during recrystallisation, the more sodic cores recrystallising at lower temperatures and anorthite content increasing with rising temperature during the metamorphism.

Biotite becomes conspicuous in these schists towards the sill margins, at times almost totally replacing the hornblende. Simultaneously the main texture changes from nematoblastic to lepidoblastic, and porphyroblastic garnets are more readily developed. Apatite content increases with the decrease of amphibole. Plagioclase is again an oligoclase or sodic andesine and sometimes exhibits incipient sericitisation, a feature absent in polygonal plagioclase of the hornblende schist. In certain sills and dykes of the east Sandøfjord area biotite has completely replaced hornblende in the groundmass fabric but important evidence of the mineral paragenesis is preserved within porphyroblastic garnets. In specimen SR 836 X from near Rødbergodden the cores of garnets contain abundant inclusions of ore and small quartz and plagioclase grains, whereas the outer zones of garnets contain much sphene often with nuclei of ore. Similar granules and lozenges of sphene (with cores of ore) are abundant within the biotite schist groundmass. Of major significance are the sporadic inclusions of hornblende within the garnet cores. From this evidence it is fairly clear that, during the early growth stages of the garnets, the sill groundmass contained amphibole and ilmenite. The outer zone of garnet has grown subsequently to the extensive metamorphic recrystallisation and replacement of amphibole by biotite (essentially the development of S_3) and the alteration of ore grains to sphene.

Porphyroblasts of almandine garnet, generally from 4 mm–8 mm but in rare cases up to 5 cm in diameter, are idioblastic and poekiloblastic. A zoned appearance is quite common with the core containing abundant small inclusions of several minerals: the outer zone frequently of slightly higher relief usually contains fewer and larger inclusions, inclusion mineralogy varying from sill to sill. Although the zones are disparate in their inclusion content, it seems likely that garnet growth has been continuous. Between the quartz-studded core and the sphene-impregnated outer zone of the garnets mentioned from specimen SR. 836 X there is a narrow, $\leqq$ 1 mm wide zone devoid of all except ore inclusions, and it would appear that this narrow zone represents a stage of slow velocity of garnet growth wherein dissolution, absorption or expulsion of quartz and other inclusions occurred. Since the garnet core is essentially of pre-F_2 age and the peripheral zone post-S_3, the crystallisation of this narrow intervening zone was probably concomitant with the widespread recrystallisation accompanying the early-F_2 deformation.

With many garnets the outer zone or skin is invariably discontinuous and it can be seen that this has grown at the expense of the biotite of the lepidoblastic S_3 fabric outside the porphyroblasts (Fig. 91). This schistosity was deflected around the pre-F_2 (core) garnet. Sphene, ore and apatite grains oriented within the schistosity retain this orientation when appearing as inclusions within the outer garnet zone.

An interesting garnet phenomenon in a dyke just north of lake 145 is that of a complete separation of the outer zone into two equal parts or 'hemispherical skins'. The garnets are up to 5 cm across and are good rhombdodecahedra, except for the split which is infilled with the groundmass schist and recrystallised quartz. Moreover, there is in all cases a displacement of one 'hemisphere' relative to the other along the plane of separation; this displacement may be anything up to 2 mm.

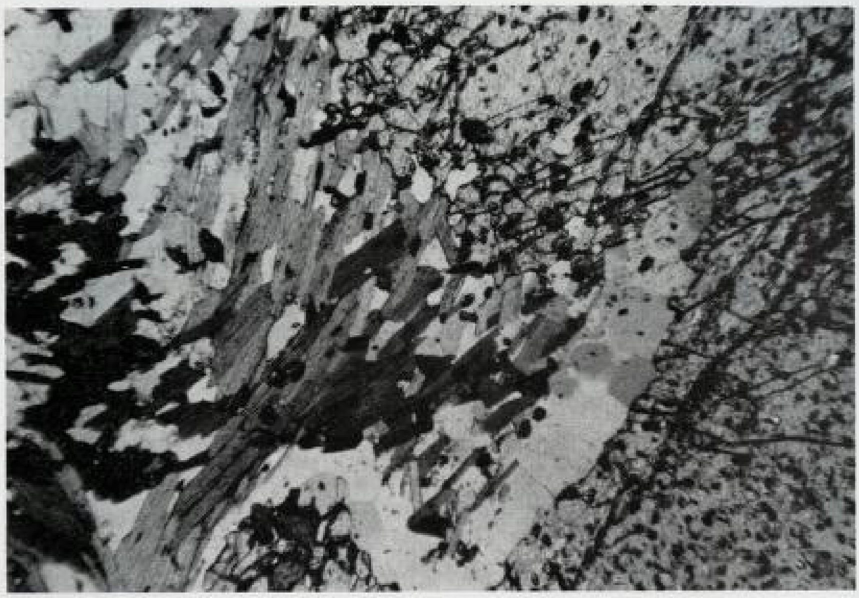

Fig. 91. Outer zone of garnet overgrowing the groundmass S_3 fabric. Garnet inner zone seen at far right of photograph.

The plane of separation is parallel or sub-parallel to the S_3 schistosity. An X-ray investigation[1] of one of these zoned garnets showed that crystal orientations in core and skin agree within 1° or 2°. The cell dimension of the core showed $a = 11.6184 \pm 0.0015$ A and the skin $a = 11.6125 \pm 0.0015$A, the difference being experimentally significant (maximum likely error is 0.002A). The difference in composition implied by the unit cell data appears to be confirmed by the difference of relief noted from core to margin in thin-sections of these garnets.

Of the remaining minerals quartz is relatively abundant in the garnet-biotite schists occurring either as groundmass grains or in pressure shadow areas adjacent to garnets. Chlorite is uncommon but may be seen partly or wholly replacing biotite flakes. In one amphibolite sill, scattered sheaves of radiate chlorite are found to be replacing hornblende. Garnet sometimes shows initial stages of breakdown, partly to chlorite. A metamict variety of allanite occurs quite commonly, producing strong pleochroic haloes in both biotite and hornblende. Rarely, a fine film of epidote may rim the allanite.

[1] By Dr. R. J. Davies of the British Museum, London.

In summary, the basic sills of this part of Sørøy are all deformed by F_2 structures and display a conspicuous S_3 schistosity. Their intrusion, clearly pre-F_2 and prior to the main regional metamorphism was effected either at the very end of the early, F_1 movements or immediately after the cessation of that period of deformation (Fig. 92). In parts of S.W. Sørøy, two distinct generations of basic sills and dykes have been recognised – (i) pre-F_2 and schistose , (ii) post-F_2 and non-schistose (see Sturt and Ramsay, 1965, Fig. 71). Differences of mineralogy within the sills of the writer's area are related to the degree of F_2 shear deformation that has affected the rock. Local degraded plagioclase phenocrysts represent relics of the original igneous texture. Garnet crystallisation is shown to have commenced prior to F_2 and extended into the F_2 period, often post-dating S_3.

C. Other basic rocks.

A basic rock-type occurring near the cliff base at Gamnes consists of a dark green amphibole and is characterised by rather diffuse, gradational boundaries into the adjacent calciferous semi-pelite. Scattered large porphyroblasts of amphibole appear in the metasediment and within only 2–3 m a rapid basification occurs in which a change from semi-pelite to ultramafite is discernible; the ultramafite is not more than 5 m thick. The amphibole (actinolite or actinolitic hornblende) is of pale yellow-green colour in thin-section, only faintly pleochroic and occurs as long laths, ragged prisms and sheaf-like aggregates; $z \wedge c = 18°$. Sphene grains are abundant throughout, often with cores of rutile. Plagioclase is uncommon, tending to occur in segregations; it is strongly sericitised with calcite and scapolite as breakdown products. Away from the basic mass, metasediment and ultramafite are intergrown with biotite progressively replacing amphibole. Quartz is now an essential mineral. It is thought that this occurrence of ultramafite represents some form of basification phenomenon more or less synchronous with the wide-spread regional metamorphism: no evidence for an igneous origin could be detected.

South-east of Hellefjord along the cliff coast of Vardas and Nordnes, F_2 axial plane shear zones are present along which sheared basic material in sometimes observed. These are 2–3 m thick and contain ramifying quartz and quartz-tremolite veinlets. Most typically the lithology is a schistose mélange of abundant hornblende with appreciable metasediment and degenerate garnets, the latter containing partially uralitised diopside. Diopside is notably absent from the groundmass, so that it would appear as if the garnet has shielded the pyroxene from rapid breakdown during the shearing. The lithology is regarded as a sheared mélange of calc-silicate schist with variable admixture of amphibolite, possibly metagabbro.

On the plateau nearby, a rock of similar appearance but whose relationships to the F_2 folding are indeterminate is found to contain porphyroblasts of tremolite-actinolite up to 1 cm across. The prevalent schistosity spindles the amphibole which is frequently flexed with variable extinction. Strongly schillerised diopsidic augite occurs as small isolate relics within large amphiboles. Optical continuity of these relics denotes the pre-existence of pyroxenes of ca. 2.4 mm. Quartz is absent, save for large late grains, while rutile may be quite prominent. The absence of sphene is also noteworthy. Plagioclase is included within the bigger amphiboles; it is calc-andesine, An_{48}. In the groundmass polygonal plagioclase is in the range An_{38}–An_{41}. The cumulative evidence would appear to favour an igneous origin for this rock-type, which has suffered uralitisation, porphyroblastesis and later F_2 shearing.

4. Summary of the structural and metamorphic history

It has been shown in the foregoing account that there has clearly been a close relationship between the sequences of tectonic and metamorphic events affecting the Eocambrian/Cambrian rocks of N.E. Sørøy. The integration of studies of the structures and textures of the metamorphic rocks has enabled the writer to subdivide the metamorphic history of the area into a number of phases which are either syn-kinematic or static in relation to the polyphase deformation.

The area comprises several large-scale second generation folds which deform a major early recumbent fold. The second episode of folding was protracted and consists of two distinct phases, the major second generation folds having been formed early in this period. Brittle deformation characterises the later phase of the F_2 episode of folding. The highest grade of metamorphism, that of the upper almandine-amphibolite facies, was established late in the static interval separating the F_1 and F_2 periods of folding and extended into the early-F_2 syn-kinematic phase. Subsequent to this the metamorphism waned and the late-F_2 brittle movements are distinctly diaphthoretic. Evidence of the

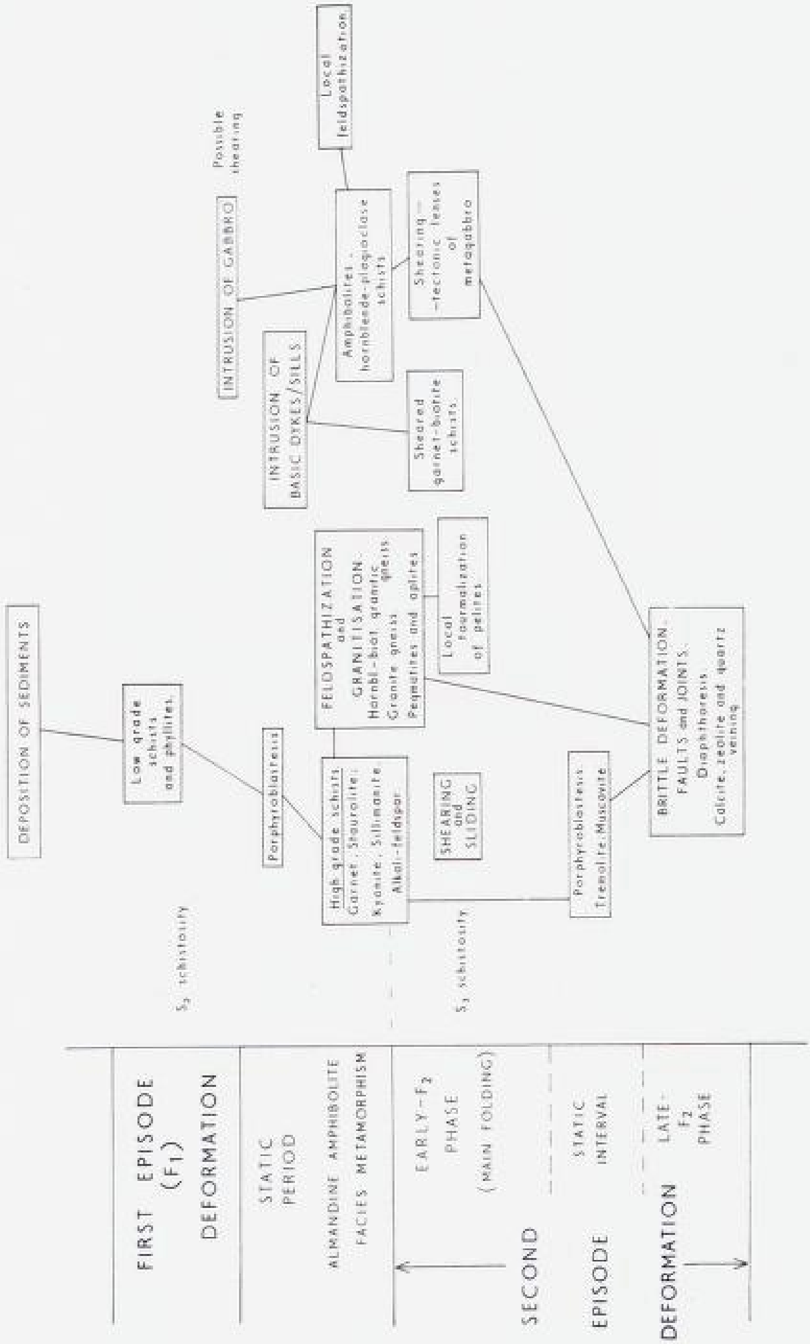

Fig. 92. Flow-diagram depicting the main metamorphic, metasomatic and structural events.

grade of metamorphism attained during the F_1 movements is rather exiguous, being restricted to that presented by the inclusion fabrics of porphyroblastic minerals. This indicates that low-grade metamorphism, essentially of greenschist facies, accompanied the early deformation. Ensuing these early movements a rise of metamorphic grade is apparent from the progressive porphyroblastesis.

The structural and metamorphic history can be outlined as a sequence of distinct though often closely related events, depicted in Fig. 92 and summarised below:

(a) The deposition of sediments, initially in a 'shelf-sea' environment passing upwards into a rhythmic sequence indicative of deeper water conditions, possibly in a developing geosynclinal trough.

(b) The first period of folding. This produced recumbent folds of considerable magnitude of which the Hønseby Fold is representative. The earliest phase of regional metamorphism – of probable greenschist facies – accompanied this episode of folding.

(c) The intrusion of a gabbroic sheet (or sheets). This occurred either late in the F_1 period of folding or immediately succeeding the cessation of movement; the F_2 deformation has erased evidence of the exact age of intrusion.

(d) The intrusion of basic dykes and sills. These post-date F_1 and are everywhere deformed by F_2 structures. They exhibit a conspicuous schistosity (S_3).

(e) The development of porphyroblastic minerals in the pre-F_2 static period. Initially, biotite and garnet porphyroblasts were developed, with amphibole and diopside in the calcareous rocks. The growth of staurolite, kyanite and sillimanite late in this period signified the attainment of the highest metamorphic grade (upper almandine-amphibolite facies).

(f) Feldspathisation and granitisation of the metasediments coeval with the stage of maximum metamorphic grade and continuing into the early-F_2 kinematic phase.

(g) The early-F_2 phase of folding. All the major F_2 folds were formed at this time. In general, F_2 folds are everywhere overturned towards the E.N.E. showing a monoclinic symmetry of movement pattern and implying an E.N.E.-directed tectonic transport. In the south, where strikes and fold axes are approximately E.-W., there is no uni-directional sense of fold overturning and the symmetry of the deforming movements is apparently orthorhombic.

High-grade metamorphic mineral growth continued into the early stages of the F_2 movement episode together with granitisation and pegmatisation of the metasediments, essentially an incipient migmatisation. A localised tourmalinisation of schists is associated with certain pegmatites. Both gabbro and basic sills were converted into amphibolites. Appreciable shearing, stretching and minor sliding accompanied the folding – many sills and dykes were converted into biotite schists while the gabbro was drawn out into elongate tectonic lenticles of amphibolite. Segregations of quartz-kyanite and garnet-biotite were developed in the more strongly deformed metasediments.

(h) A post-S_3 static phase. Certain minerals, notably tremolite, muscovite, scapolite and garnet, were developed as porphyroblasts overgrowing the S_3 schistosity.

(i) The late-F_2 brittle deformation with associated retrogressive metamorphic phenomena. Most of the faulting and jointing is of late-F_2 age.

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Sammendrag

Langstrand–Finfjord området på Sørøy karakteriseres av en metasedimentær lagserie bestående av kvartsitter, pelittiske og semi-pelittiske skifre og kalksteiner. Seriens stratigrafiske systematikk lar seg basere på de hyppige forekommende sedimentærstrukturer som er bevart i de massive psamittene. Mens den laveste delen av serien hører hjemme i eokambrium så er det meget mulig at de øverste (kanskje største) deler er avsatt i kambrium. Dette antydes ved funn av eksemplarer av Archæosyathider (under kambrisk alder) ved B. A. Sturt i en kalkstein i et område sydvest på øya som ennå ikke er kartlagt i detalj. Det har vært rapportert at denne kalksteinen forekommer innen den øverste delen av Klubbenkvartsittgruppen.[1]

Det underste ledd i serien, Klubbenkvartsittgruppen består hovedsakelig av massive eller båndete kvartsitter som lokalt oppviser fine og hyppige eksempler på kryss-skiktning. Den påfølgende Storelvskifer og Falkenes-kalkstein-gruppen synes å representere grunnvannsomgivelser ("shelf-sea"); en overflate med tørkesprekker i en uren kalkstein antyder at det til sine tider har hersket helt uttørrede forhold. De rytmisk avsatte Hellefjordskifrene med lokalt utviklet undervannskredlitologi antyder en fordypning av avsetningsbassenget, mest sannsynlig under initialstadiene for dannelsen av en geosynklinal.

Ved studier av primære strukturer kan det vises at lagene er inverterte over en stor del av området.« Inversjonen må tilskrives en makroskopisk overblikket ("recumbent") fold som tilhører den første deformasjonsperiode. I det minste to hovedfoldgenerasjoner har innvirket på metasedimentene, men den siste foldeepisoden var i seg selv nokså kompleks og er delelig i to, kanskje tre, faser. Større folder som tilhører den første fase i den andre deformasjonsperioden dominerer det tektoniske bildet og er hovedårsaken til lagenes nåværende stilling. Senere faser i den samme perioden inneholder også strukturer av kryssfoldtypen og lokalt foldinger bundet til en mer oppbrukket deformasjon og forkastning.

På denne delen av Sørøy er mesoskopiske F_1 folder relativt få i antall og viser hovedsakelig en tett og isoklinal stil: foldeaksene er svært variable i retning. Den regionale skifrighet henger sammen med denne tidlige foldeperioden. F_2 foldene er hyppige, fremkalt i alle størrelsesordener og deformerer tydelig de tidligere strukturer. De varierer i stil

[1] Sturt et al., Norsk geol. Tdsskr., Bd. 47, 1967, p. 256.

fra åpne til tette eller nesten isoklinale strukturer. Over store deler av området har F_2-aksene og lineasjonene en NNV-SSØ-retning, men i syd manifesterer det seg en dreining over til ØNØ-VSV-retning. Mens de NNV-rettede folder viser monoklin struktursymmetri, viser de ØNØ-rettede ingen felles retning i overveltningen og er karakterisert ved orthorombiske eller svært ofte, triklin symmetri. Det er en tydelig overgang i foldekarakterer og symmetri fra den ene orthogonale strøkretning til den annen, slik at den fremherskende bøyning ikke kan tilskrives en deformasjon senere enn F_2. Mulige årsaker til de forskjellige dannelser av F_2 folder blir diskutert og sammenligninger gjort med andre områder i den kaledonske fjellkjede.

Teksturstudier av metasedimentene i forhold til de to tektoniske episodene tillater en videre oppdeling av deres metamorfe historie i et antall faser. Hver av disse er karakterisert ved krystallisasjon av bestemte mineraler og ved en større eller mindre modifikasjon av den generelle bergartsstrukturen. Det kan ikke påvises noen diskontinuitet mellom noen av de forskjellige fasene. De bevegelser som forårsaket F_1-isoklinalfordelingen ble sannsynligvis bare fulgt av en lav grad av metamorfose, grønnskiferfacies. Metamorfosegraden øket så i den statiske periode forut for F_2-deformasjonen, med porfyroblastene av granat, staurolitt, kyanitt og sillimanitt i de pelittiske skifre, og tremolitt og diopsid i de kalkige metasedimentene. Den høyeste grad av regionalmetamorfose (øvre almandin-amfibolittfacies) ble etablert mot slutten av det statiske interval og strakte seg også inn i denne siste episode, og det kan også påvises at den kommer senere enn den intialkinematiske fase av F_2-episoden. Krystallisasjonen fortsatte derfor inn i denne siste periode, selvom metamorfosegraden avtok betraktelig etter dannelsen av S_3-skifrigheten. Diaftoretiske fenomener kjennetegner den siste fasen av sprø deformasjon.

Samtidig med kulminasjonen i metamorfosen oppsto det en lokal granittisering i metasedimentene. Denne fortsatte til dels inn i de tidligere stadier av den andre deformasjonsperioden. To hovedtyper av gneiser eller granittiske gneiser ble dannet, pegmatitter og aplitter finnes også. Både plagioklas og mikroklin porfyroblastene er vanlig utbredt, selvom en kalimetasomatose gjennomgående er senere enn bølgen av natriumøkning. En lokal turmalinisering av pelittene ved Finfjord er nært forbundet med dannelsen av pegmatittlegemer. Disse er vanligvis også turmalinførende.

Basiske bergarter er lite hyppige i området, og opptrer som metagabbro i linseformede legemer og amfibolitter i lagerganger og ganger. Metagabbrolinsene er antatt å være tektoniserte derivater av Storelv-gabbroen, beskrevet av Stumpfl og Sturt (1965 fra den sentrale del av Sørøy. Bergarten er tydelig deformert av F_2-foldene og antas i alder å være fra den siste del av F_1-deformasjonen eller følger umiddelbart etter F_1-bevegelsene. Lagergangene av amfibolitt som lokalt er omvandlet til granat-biotittskifer under F_2-deformasjonen, er tydelig intrudert etter den første folding, men før metamorfose-kulminasjonen. Relikte magmatiske strukturer er påvist i enkelte partier av disse bergartene.

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Note added in proof

Following investigations in the north-easternmost part of Sørøy in 1967, the Hellefjord Schist has been renamed 'Hellefjord Schist Group' and is now regarded as a sequence of alternating pelagic sediments and distal turbidites (Roberts, N. G. T., in press). Furthermore, the Falkenes Limestone Group has been discovered in the core of a major near-isoclinal F_1 fold between Kjøtviken and Vasvik, upthrown on the north side of a major E-W fault. The fold is anticlinal, overturned towards the NNE, and is thought to be either an extension of the Hønseby Fold or a separate F_1 structure. Minor linear structural observations show this F_1 fold to have an ESE-WNW axial trend. Axes of F_2 folds in this same area plunge towards a south-south-easterly point.

Plate I. Geological map of the Langstrand area, Sørøy, Northern Norway.
Plate II. Geological structures of the Langstrand area, Sørøy, Northern Norway.
Plate III. Geological sections across the area.

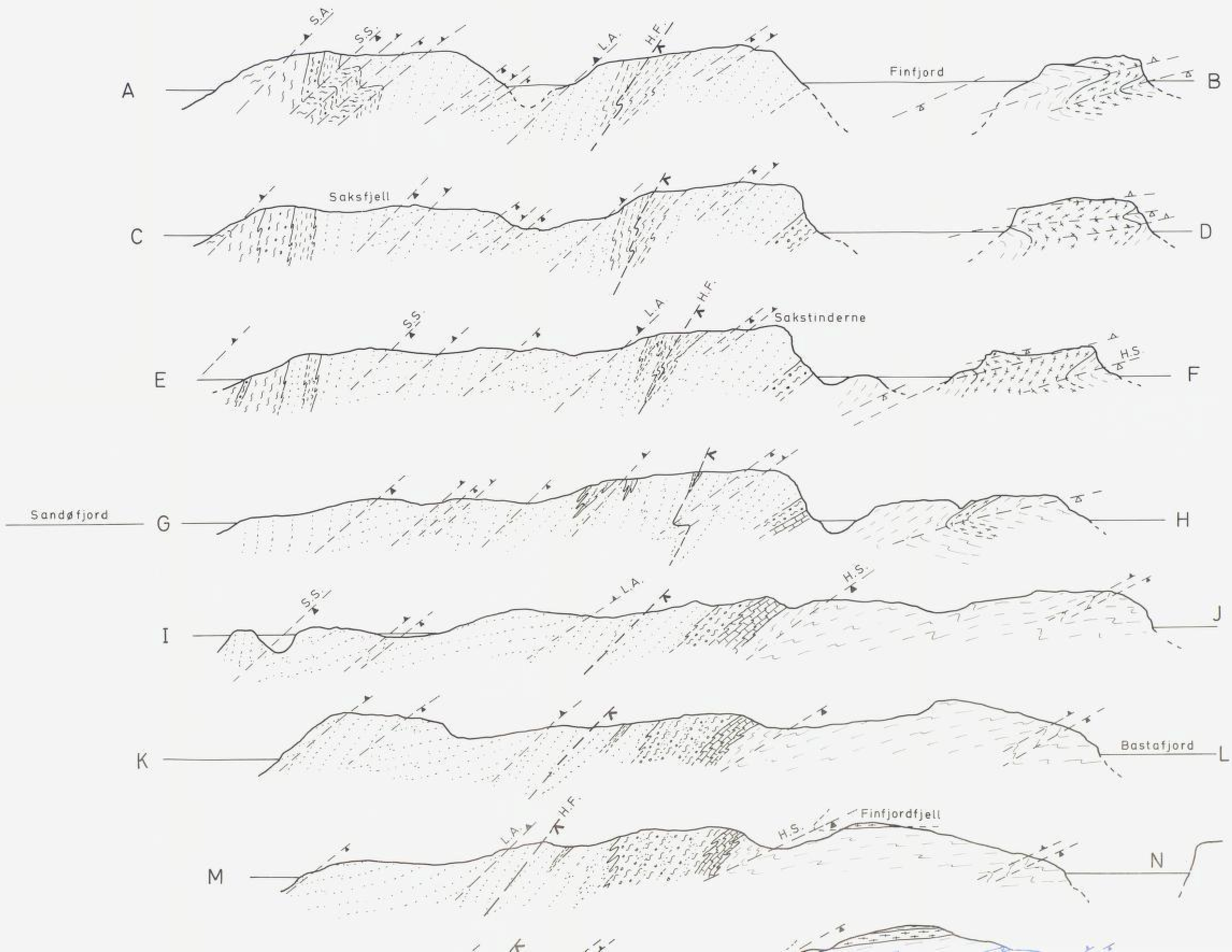
GEOLOGY OF THE LANGSTRAND AREA, SØRØY, NORTHERN NORWAY.

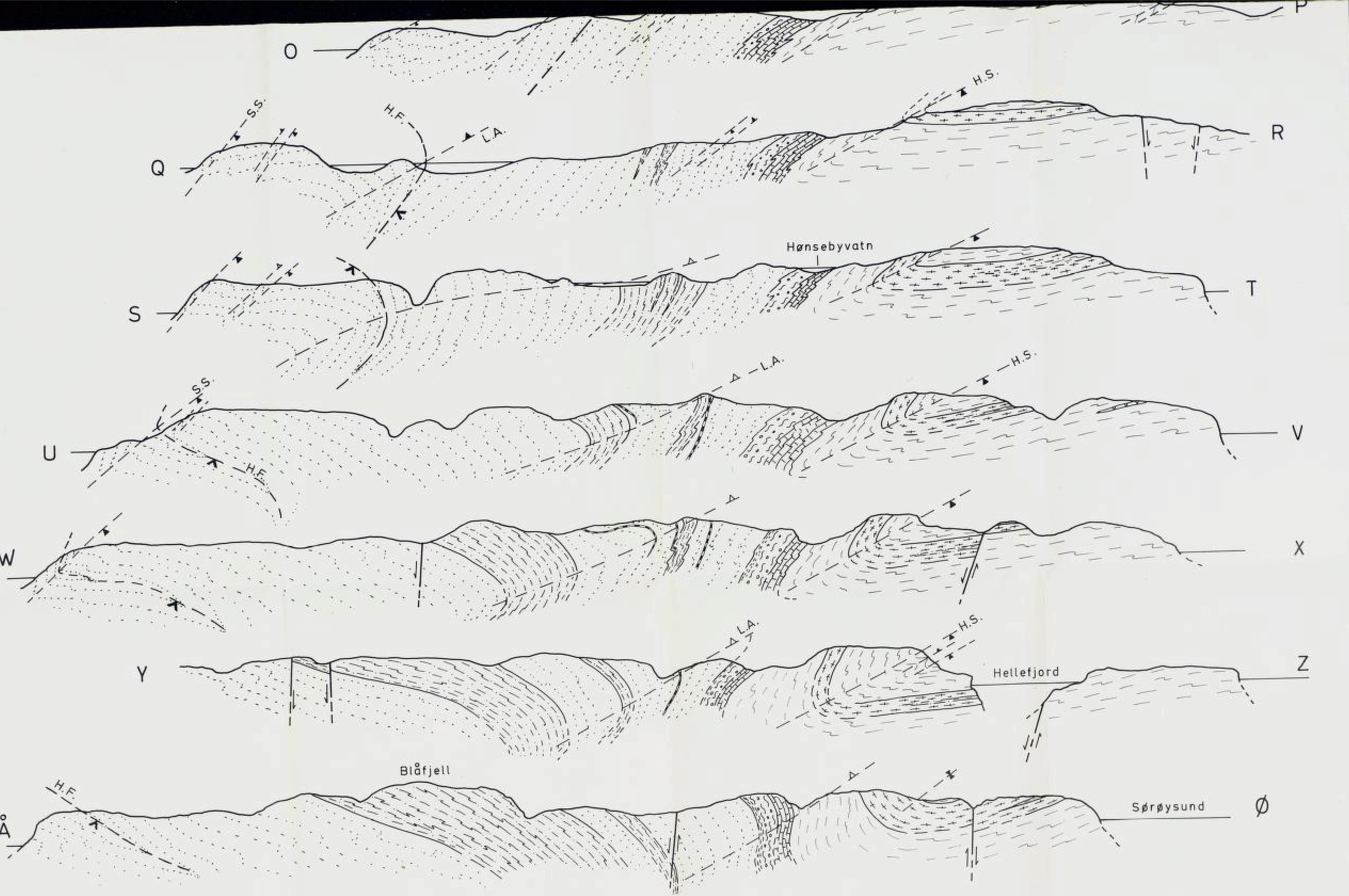
D. ROBERTS

Grasdalsnes
Finfjorddalen
Bastafjord
Finfjordfjell
Hønsebynes
Hønsebyfjord
(170)
Hønsebydalen
Otervikfjell
Vest-strømmen
(145)
Hønsebyvatn
Blåfjell
Lundvatn
Lunddalen
Skippernesfjord
Veirbukten
(135)
(137)

Langstrandfjord
Langstrand
Gamnes
Sandbukten
Lundhavn
Veirfjell
Hellefjord
Nordnes
SØRØSUND
LEGEND
(Not to scale)
HELLEFJORD SCHIST
Garnetiferous calc-silicate schist and phyllite
FALKENES LIMESTONE GROUP
Impersistent quartzite
Two limestones and garnet-kyanite-sillimanite calc-silicate schists
STORELV SCHIST
Biotite-muscovite schist
Garnet-muscovite schist, sometimes with kyanite and staurolite
TRANSITIONAL GROUP
KLUBBEN QUARTZITE GROUP
Quartzite 3
Upper semi-pelite
Quartzite 2
Lower semi-pelite
Quartzite 1
Lithological boundary
Transitional lith. boundary
Inferred lith. boundary
Dip and strike of bedding (banding in gneisses)
Vertical strata
Horizontal strata
Current bedding – direction of younging
Faults
Granite gneiss
Biotite-hornblende granitic gneiss
Pegmatite
Microcline-rich lenticles (gneiss)
Metagabbro
Amphibolite sills and dykes
1 0 1 2 km
1000 500 0

SECTIONS ACROSS THE AREA





H.F. — HØNSEBY FOLD
L.A. — LANGSTRAND ANTIFORM
H.S. — HELLEFJORD SYNFORM

S.S. — SKARVFJORD SYNFORM
S.A. — SKARV ANTIFORM

HORIZONTAL AND VERTICAL SCALE

0 1 km.

GEOLOGICAL STRUCTURES OF THE LANGSTRAND AREA, SØRØY, NORTHERN NORWAY.


SKARVFJORD SYNFORM
ANTIFORM
HELLEFJORD SYNFORM
HØNSEBY FOLD
Hønsebyfjord
HØNSEBY FOLD
Vest-strømmen
LANGSTRAND ANTIFORM
Hønsebyvatn
LEFJORD SYNFORM
Bastafjord
Skippernesfjord
K
M
O
Q
S
U
W
Å
Y
L
N
P
R
T
V

Langstrand
Gamnes
Sandbukten
Lundhavn
Hellefjord
SØRØSUND
Z
Ø
STRUCTURAL SYMBOLS
Bedding (banding) — S_1
F_1 STRUCTURES
F_1 fold axes
S_2 schistosity
S_3 schistosity
Fold axes
Banding/schistosity intersection
Mineral lineation
Mullions
Boudinage elongation
Quartz rodding
Cross folds
Transverse boudin axes
Late-F_2 microfold axes
Slide striae
F_2 STRUCTURES
Axial plane trace of major F_1 fold.
Axial plane trace of F_2 antiforms
Axial plane trace of F_2 synforms
Approximate axial plane trace of F_2 fold (Langstrand Antiform)
Various combinations of symbols are shown, e.g. —F_2 fold axes, boudinage and quartz rodding.
Faults
Slides
G— —H etc. — lines of section
Sedimentary structures (current bedding, direction of younging.)
1 O 1 2 kms.
1000 500 0
D. Roberts 1964