THE STRUCTURAL AND METAMORPHIC DEVELOPMENT
OF THE BERGEN ARC/BERGSDALEN FORELAND
GNEISSES OF OSTERøy, S.W. NORWAY

by

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VOLUME I

Thesis submitted for the Degree of
Doctor of Philosophy

Bedford College (University of London) February 1980
Dedicated to the memory of
Mark Howard Johnson
ABSTRACT

North Østerøy comprises autochthonous and para-autochthonous basement which has been reworked during the orogenic development of the adjacent Bergsdalen and Bergen Arc nappes.

The divisions of the basement, the Southern and Northern Gneiss Units, are composed of acid gneisses, paragneisses, migmatites, orthoamphibolites and granites. Three sedimentary complexes are identified, the first (circa 1500 Ma.) resembles the Fjordane complex. The second is correlated with sediments in the Bergsdalen Nappes (older than 1000 Ma.), and with the Samnanger complex in the Bergen Arcs. The third sedimentary sequence is the Ashgillian, Holdhus sequence of the Bergen Arcs.

The rocks of the Southern Gneiss Unit are derived from, and emplaced over, the Northern Gneisses during four deformations, each of which is separated by a sedimentary interval. The deformation involves 'mylonitic' processes acting at high temperatures and high strain rates.

The first deformation forms part of the Svecofennian orogenic cycle. The second and third form part of the Sveconorwegian cycle and produce folds with north-easterly trending axial traces. The third is responsible for the emplacement of the primary Bergsdalen Nappes and contrary to current opinion these Nappes are thought to root in the south east.

The final deformation is part of the Caledonian orogenic cycle and involves the development of the Bergen Arcs and their emplacement over the Bergsdalen Nappes. In the early stages of this cycle, structures formed on an east-north-east Bergsdalen trend, but later, non-cylindrical folds with north-westerly trending axial traces developed, giving rise to the Bergen Arc trend. Of the two deformations which produce the Bergen Arc and Bergsdalen Nappes, the Sveconorwegian event is the more important.

The interpretation of the geology of N. Østerøy and consideration of its relationships to adjacent regions allows a new plate tectonic interpretation of Western Norway.
ACKNOWLEDGEMENTS

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VOLUME 2
Plates 1 to 105
Map 1
Figure 158
Microfiche
This thesis is based on a study of the rocks of North Osterøy, near Bergen, West Norway. The general position of the area relative to the main Caledonian Orogen is shown in figure 1.

The work forms part of a project arranged between the University of Bergen and Bedford College (London University), the aims of which are to re-examine the geology of the Bergen Arcs (Kolderup & Kolderup, 1940) and the immediate surrounding region (fig.2). The responsibility of the Bedford College contingent was to examine the geology of the basement gneisses to the east and west of the Arc structure, while the Institute in Bergen concentrated on remapping the lithologies comprising the Arcs. The area mapped (fig.9) was chosen on several counts, both logistic and academic. Less was known about the basement to the east of the Arcs than the Øygarden basement to the west, and it was felt that this should be studied first, with the object of examining the Arc basement contact.

The British contribution is financed by N.E.R.C. and directed by Professor B.C. King.

1.1. TOPOGRAPHY AND GEOLOGY

Osterøy is the only place on the eastern margin of the Arcs where the contact between the basement gneisses and the rocks of the Arcs is well exposed; elsewhere the junction is covered by sea. The rocks of the Arcs comprise the south-west portion of the island to the south and west of the central Storvatnet lake. These rocks form a fertile soil and the area is populous and well served by roads. The gneisses occupy the whole of the north-eastern portion of the island to the north and east of Storvatnet. These rocks form the highest ground of the island as
THE SCANDINAVIAN CALEDONIDES

The Northwestern Bergen Gneisses

The Telemark Gneisses
they resist weathering and are for the most part exposed as barren rock.
The villages are sited in the coastal valleys where glacial drift provides
arable land. The tree line on this ground is at about 150 metres.

The highest point in the area mapped (the outcrop of these north­
eastern gneisses) is Hogafjell, a mountain in the centre of the outcrop
which rises to 869 metres. The hills are dissected by steep-sided
valleys which follow sets of north-east and north-west trending joints
(map 1). Numerous lakes develop at the points where these joints inter­
sect. The tops of the hills are reasonably level but routes up to this
plateau are scarce (map 1). Access from the fjord coast is impossible,
the fjord sides are precipitous except at the mouths of rivers, which
are sites of villages.

Climate

The mapping season in this area is from June to October; at other
times the outcrops may be snow covered and/or the working day is too
short. Rainfall is generally high throughout the year.

Mapping

The segment of basement chosen for mapping forms a triangle of gneisses
between the rocks of the Bergsdalen and Bergen Arc nappes, ideally positioned
to study the relationship between the two, as regards basement structure
and constitution. The area north-east of the southernmost thrust on
figure 9, was mapped at various scales, depending on the availability of
maps. The basic mapping is at 1:5000, with the remainder of the area
covered at 1:10,000, except where no maps were available, when mapping
was done direct onto air photographs at a scale of 1:15,000. These
photographs were later used in the construction of map 1, using the
techniques of radial line plotting to transfer the more important
topographic features and the geology.
Form of the Thesis

The thesis is divided into five main sections. The first places Osterøy in its regional setting, reviewing what is known of the geology of the surrounding areas. The next two sections are descriptive, dealing with rock composition, field and thin section appearance, structure and the elucidation of structural and metamorphic history; any immediate conclusions that may be reached are presented as discussions within these sections. In the process of mapping and in attempting to understand the mineral textures observed, it became clear that a straightforward description of the observed features would not provide a complete understanding of these rocks. The apparent heterogeneity of the observed strains and the palaeometamorphic textures could only be adequately explained in the light of recent advances in the understanding of the nature and generation of mylonites. The fourth section is a description of the rocks in this light and is in part theoretical. The final section deals with the conclusions that may be drawn directly from the area of study, incorporating the theory discussed in the fourth section. Regional issues raised by these observations are discussed at the end of this section.

Techniques, computer programmes and data are included as a set of appendices.

1.2 Regional Setting

Osterøy is an inland island, surrounded by fjords, situated 30 Km to the north-east of the city of Bergen. It straddles the central eastern "front" of the Bergen Arc structure. The south-western portion of the island cuts deep into the centre of the Arcs and comprises most of the usual lithologies encountered within the structure. The north-east portion (the area mapped) extends into a large tract of gneissose
rocks believed to form part of the "North Western Gneisses" (Kvale, 1960), an area of basement gneisses, outcropping in the core of a major geanticlinal structure on a north-easterly trend between Trondheim and Bergen (fig.1), (see also Southern Norway Tectonic Map, 1975). Immediately to the south east of Osterøy are the "Bergsdalen Nappes" (Kvale, 1946), which continue on a north-east trend passing beneath the "Upper Jotun Nappe" (Goldschmidt, 1916), to the north and out to sea to the south (fig.3).

This part of south-western Norway can be divided into six main regions (fig.4):

1) The Øygarden Basement Gneisses Complex
2) The Bergen Arc Nappe Complex
3) The North-Western Basement Gneiss Complex
4) The Bergsdalen Nappe Complex
5) The Jotun Nappe Complex
6) The Telemark Basement Gneiss Complex

The Nappe complexes are preserved in major synformal troughs (Southern Norway Tectonic Map, 1975). It is generally thought that the nappes root to the north-west (Hossack, 1978) in a position presently occupied by the North Sea, and that they once covered all of the North-Western Basement. They have subsequently been eroded, being now only preserved as infolds in basement depressions (Ramberg, 1977).

Figure 5 is a gravity map of the region (after Ramberg, 1977), which shows two large areas of basement gneisses, the North-Western Gneiss Complex and the Telemark Complex form areas of strong negative anomaly. This indicates the crust in these regions is thick and not very dense, comprising mainly acid material. Beneath the outcrop of the Bergen/Jotun Nappes, however, this negative anomaly is not so marked. Indeed
Principal Tectono-stratigraphic Units of the Southern Norwegian Caledonides.

Fig 4

TECTONICS OF THE REGION ROUND OSTERØY.

1. Øygarden Basement Gneiss Complex
2. Bergen Arc Nappe Complex
3. North Western Basement Gneiss Complex
4. Bergsdalen Nappe Complex
5. Jotun Nappe Complex
6. Telemark Basement Gneiss Complex
Fig 5.

BOUGUER ANOMALY MAP OF
SOUTHERN NORWAY

After RAMBERG, 1971.

[MGAL'S]
the area is anomalous in the manner in which the anomaly departs from
the general trend. There are two possible explanations for this anomaly
(Ramberg, 1977; Smithson, Ramberg & Grønlie, 1974), either the crust
is thinner in this region, or there is more basic rock included in the
crust.

Seismic traverses over the region (Sellvoll & Warrick, 1971;
Kanestrom & Haugland, 1971; Kanestrom, 1977) prove the crust in the region
to be 32-38 Km thick, with the thickest part lying under the outcrop of
the Bergsdalen/Jotun Nappes. This indicates that the Bouguer anomaly
must be caused by the presence of denser rocks somewhere within the
crustal section.

The question is then one of the depth and thickness of this body of
dense rock; whether the anomaly is attributable to the rocks seen in
outcrop, or is caused by something at depth.

There is an accompanying magnetic anomaly associated with the
Bergsdalen/Jotun outcrop, which follows the gravity anomaly. The
preferred explanation for this feature is the presence of a large mass
of gabbroic rock, about 3 Km thick (Aalstad et al., 1977). This is at
variance with the preferred interpretation of the gravity data (Smithson
et al., 1974), which requires a thickness of at least 8 Km for the
gabbroic rock and the problem remains to be resolved.

Whatever its cause, the anomaly forms a marked linear feature in
the crust on the Bergsdalen/Jotun axis, passing to the south-east of
Osterøy. Off the west coast of Bergen there is another strong positive
anomaly, a steep gradient on a north-south axis passes right through the
Arcs. The cause of this is unclear, but the presence of the Viking
Grablen off the coast to the west, in the North Sea (Nilsen, 1976), might
possibly offer an explanation for this. This would mean that the gradient
through the Arcs is probably a relatively recent Mesozoic (Nilsen, 1976) modification of the older north-east - south-west trend described above.

Osterøy is thus situated just to the north-west of a major linear anomaly in the crust; it is therefore ideally placed to study how the basement has responded to the emplacement of the Bergen Arcs and Bergsdalen Nappes. If the geophysical feature has any surface expression it is likely that the rocks on Osterøy might shed some light as to its nature.

The Bergen Arcs and the Øygarden Basement

The first account of the geology around Bergen was given by Naumann (1824), who described the arcuate form of the outcrops. He was followed by Meilhau (1850), Hiortdahl and Irgens (1862), Kjerulf (1879), Reusch (1882), Kolderup (1897), Kolderup and Monckton (1912), Kolderup (1931), but it was Kolderup and Kolderup (1940) who appropriately referred to "The Bergen Arc System". This describes the two parallel belts of schists, approximately semicircular in shape and with a centre immediately to the west. The neighbouring formations are also arranged in a similar manner and so, together with the arcs s.s., form a large system of arcs of which there are five major units of differing lithology and of varying metamorphic grade. These units are separated by tectonic junctions (fig.2), which are major planes of transport for the allochthonous units, although within each unit there is much smaller scale imbrication. The five major units are, from the centre outwards:

1) Øygarden Gneiss Complex
2) Minor Bergen Arc
3) Ulriken Gneiss Complex
4) Anorthosite Complex
5) Major Bergen Arc.
The Øygarden Gneiss Complex

This is exposed as the "kernel" of the arcs and is on Sotra; it is represented by a series of granitic, granodioritic and tonalitic gneisses with amphibolites. They are considered to be of Svecofennian age (Sturt, Skarpenes, Ohanian & Pringle, 1973), but to have undergone several phases of reworking under amphibolite facies conditions with at least two stages of anatetic mobilisation, which evidently occurred in both the Grenvillian and Caledonian orogenies (Sturt et al., 1975). There is also some evidence for activity during the Grampian event.

Although no thrusts are seen in this unit and it apparently forms a local basement underlying the Bergen nappes to the east, it may itself be an allochthonous unit.

The Minor Bergen Arc

This overlies the Øygarden gneisses and crops out beneath the city of Bergen. The unit comprises metasediments and metavolcanics in which a number of impersistent horizons of gneiss occur. These usually have strong mylonitic fabrics. The contact with the underlying Øygarden unit is also a zone of intense deformation in which a great variety of mylonitic rocks are developed (Thon, Pers. Comm., 1977). To the east the Minor Arc is bounded by a major thrust underlying the Ulriken Gneiss Unit. These features suggest that the Minor Arc is a flattened zone of imbrication between the basement and a succession of thrust nappes, of which the Ulriken Gneiss Unit is the first. The Minor Arc has generally been regarded as of Lower Paleozoic age (Sturt & Thon, 1978), though there is no proof of this; all that can be said is that it is pre-Ashgillian.
The Ulriken Gneiss Complex

This consists of a complex of thrust slices of gneisses with interlayered zones of metasedimentary rocks. The gneisses are essentially migmatitic with a comparatively late anatectic event at 1488 Ma. (Sturt et al., 1975). The gneisses are traversed by numerous zones of mylonite and are separated from the underlying rocks of the Minor Arc by a thick band of mylonite (Sturt et al., 1975), and at least three thrust slices have been identified in this allochthonous basement nappe.

The Anorthosite Complex

Outcropping to the east of the Ulriken Gneiss Complex, and separated from it by a distinct thrust contact, the Anorthosite unit comprises a complex of imbricated and thrust plutonic igneous rocks, para- and ortho-gneisses. The oldest members (1775 Ma., Sturt et al., 1975) are anorthositic in composition. These rocks have a protracted history of deformation and metamorphic recrystallisation in the granulite facies (Griffin, 1972), before the intrusion of a younger suite of mangeritic rocks (1064 Ma. Sturt et al., 1975; Sturt & Thon, 1978). The mangerites have also been subject to granulitic facies metamorphism. The rocks of the Anorthosite Complex show the effects of reworking under both amphibolite and greenschist facies conditions, which are largely attributable to the Caledonian orogeny (Sturt & Thon, 1978).

The Major Bergen Arc

The Major Arc is subdivided into two sequences separated by a major unconformity (Sturt & Thon, 1976). The older of these sequences, the Samnanger Complex (Faerseth, Thon, Larsen, Sivertsen & Elvestad, 1977), comprises heterogeneous schists, quartz conglomerates, amphibolites and greenstones, together with metaperidotites, gabbros and quartz diorites. The rocks are characterised by highly complicated structures and several
zones of extensively mylonitised and recrystallised Precambrian gneisses within the schists. The age of these rocks is unknown, though in the literature (Faerseth et al., 1977) the sequence is assumed to be of Lower Palaeozoic age.

The Holdhus Group is unconformable on the Samnanger Complex (Faerseth et al., 1977). The Moberg conglomerate represents the oldest formation of the group, passing up into marble which is followed by the Gråsdalen Formation, including a fossiliferous Ashgillian marble.

Figure 6, after Sturt and Thon (1978), is a table of the Bergen Nappe sequence.

The Structure of the Bergen Arcs

Kolderup and Kolderup (1940) regarded the rocks of the Arc system as representing a Lower Palaeozoic metamorphic suite, containing a continuous stratigraphic sequence from Precambrian gneisses through late Precambrian metasediments (the Minor Arc and Ulriken Gneiss), late Precambrian metavolcanics (the Anorthosite Complex), into fossiliferous lower Palaeozoic rocks (the Major Bergen Arc). The only basement in the area is that exposed on Øygarden. Hernes (1967) supported this view, but Kvale (1960), following Reusch (1882), interpreted the area in terms of nappe tectonics. He believed that the anorthositic and associated gneissic rocks represented an allochthonous thrust unit which overlay and separated the Major and Minor Arcs with their sequence of Lower Palaeozoic metamorphic rocks. Recent field and geochronological studies support Kvale's proposition, although the pattern of thrust tectonics is more complex than he envisaged.

Kvale regarded the Major Arc as being overlain by the anorthositic rocks, and suggested that the presence of an outlier of these rocks in the easternmost part of the Major Arc (fig. 2) represented a thrust mass
Fig 6.

### Bergen Nappe Sequence, Southern Norway

<table>
<thead>
<tr>
<th>TECTONO-STRATIGRAPHIC UNITS</th>
<th>MANTOLOGIES</th>
<th>FAULT TRENDS</th>
<th>MINERALOGY</th>
<th>CHARACTERS OF DEFORMATION AND GRADE OF METAMORPHISM</th>
<th>PALEOMAGNETIC ARCS</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>FANAFJELL NAPPE</strong></td>
<td>Orthoclase-porphyroclastic</td>
<td>None</td>
<td>Granites</td>
<td>Polycyclic deformation probably both Pre-orogenic and Caledonian. Caledonian fabrics dominated by Pre-orogenic, Caledonian greenschist facies metamorphism.</td>
<td>453 ± 50 m.y., ( \text{LA} ).</td>
</tr>
<tr>
<td><strong>ANORTHOSITE COMPLEX</strong></td>
<td>Anorthosite, hornblende, magnetite, quartz, plagioclase, orthopyroxene</td>
<td>None</td>
<td>Anorthosite, hornblende, magnetite, quartz, plagioclase, orthopyroxene</td>
<td>Pre- orogenic polyphase deformation, Preorogenic polyphase green schist facies metamorphism. Preorogenic polyphase green schist facies metamorphism.</td>
<td>1347 ± 150 m.y. ( e ).</td>
</tr>
<tr>
<td><strong>ULRIKEN GNEISS NAPPE</strong></td>
<td>Quartz, feldspar, garnet, muscovite, biotite, amphibole</td>
<td>None</td>
<td>Quartz, feldspar, garnet, muscovite, biotite, amphibole</td>
<td>Pre- orogenic polyphase deformation, Preorogenic polyphase green schist facies metamorphism. Preorogenic polyphase green schist facies metamorphism.</td>
<td>1347 ± 150 m.y. ( e ).</td>
</tr>
<tr>
<td><strong>SARMANGIE COMPLEX</strong></td>
<td>Muscovite, quartz, biotite, garnet, amphibole</td>
<td>None</td>
<td>Muscovite, quartz, biotite, garnet, amphibole</td>
<td>Pre- orogenic polyphase deformation, Preorogenic polyphase green schist facies metamorphism. Preorogenic polyphase green schist facies metamorphism.</td>
<td>1347 ± 150 m.y. ( e ).</td>
</tr>
<tr>
<td><strong>DYGARDENS GNEISS COMPLEX</strong></td>
<td>Plagioclase-orthogneiss with amphibole</td>
<td>None</td>
<td>Plagioclase-orthogneiss with amphibole</td>
<td>Preorogenic polyphase deformation, Preorogenic polyphase green schist facies metamorphism. Preorogenic polyphase green schist facies metamorphism.</td>
<td>1347 ± 150 m.y. ( e ).</td>
</tr>
</tbody>
</table>

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subsequently folded into a tight synformal structure. The discovery of the mid-Ordovician unconformity (Sturt & Thon, 1976), shows that the development of the Major Arc was a two stage process, comprising an early deformation before the deposition of the Holdhus Group, which is responsible for major stratigraphic inversions within the Samnanger Complex (Faerseth et al., 1977). Further deformation causing both new folding and secondary imbrication, affects both sequences of rocks (Ryan & Skevington, 1976; Sturt & Thon, 1976; Sinha Roy, 1977c; Faerseth et al., 1977 and 1978).

The youngest major feature and the highest unit within the Arcs is the Fana Nappe (Sharman & Sturt, 1977). This comprises mainly granitic gneisses and minor amounts of metasediments transgressively emplaced across the already folded thrust plane between the Anorthosite Complex and the Major Bergen Arc. This late thrust is itself strongly folded (Sturt & Thon, 1978).

The Bergen Arc System can be regarded as a complexly refolded nappe pile sitting in a synformal basement depression. Each unit is a separate thrust sheet and the simplest explanation would be to correlate the rocks of the Minor Bergen Arc with those of the Samnanger Complex of the Major Arc, joining them up beneath the Anorthosite Complex. The structural succession would then be as shown in figure 6.

The North Western Gneiss Complex

The North-Western Gneiss Complex (area 3 on fig. 4), contains a great variety of rock types, bounded in the west, along the coast, by several synformal structures in which Devonian and what are regarded as Cambro-Silurian rocks are preserved. These later sediments appear to be tectonically emplaced along low lying thrusts (Bryhni, 1977; Skjerlie, 1968). In the east, the region is bounded by a thick sequence of Eocambrian rocks.
of arkosic character, while to the south is the junction between the basement gneisses and the overthrust rocks of the Bergsdalen and Jotun Nappes (Banham, 1968; Bryhni, 1977).

The gneisses include obvious metasedimentary rocks, mica schists and quartz schists, together with sporadic layers of amphibolite, eclogites and anorthosites.

Until 1938, Norwegian geologists believed this area formed a Pre-cambrian substratum to the Eocambrian and Cambro-Silurian rocks which surround it. Barth (1938) and Holtedahl (1938), however, postulated that some of the rocks in the gneisses might well be Eocambrian or Cambro-Silurian in age. Kolderup (1960) suggested that the whole of the gneissose area was produced by the granitisation and metasomatic alteration of what were originally Eocambrian and Cambro-Silurian rocks, having lithologies similar to those which surround the gneiss area. He regarded the junction between the observed Eocambrian and Cambro-Silurian rocks and the gneisses as the limit of the effects of the granitisation. He presumed the eclogites to be produced in the deep-seated heart of this area of granitisation from portions of the rocks which were originally amphibolitic in composition. This view was also supported by Hemes (1970).

More recently Bryhni (1966), subdivided the gneisses into two complexes, one of which, the Jostedal Complex, a sequence of relatively homogeneous two-feldspar gneisses, migmatites and augen gneisses, he believed to be reworked (or Caledonised) Pre-Eocambrian basement. This was considered to be overlain (in the west) by a more heterogeneous series of gneisses containing schists, quartzites, marbles, eclogites, amphibolites, peridotites and anorthosites - the Fjordane Complex. He believed this complex to be derived from rocks of Cambro-Silurian age and older elements corresponding to the Jotun thrust masses.
The two complexes (fig. 7), have been widely recognised throughout the area (Brueckner et al., 1968; Strand, 1969; Bryhni & Grimstad, 1970; Bryhni et al., 1971; Carswell, 1973). Brueckner (1977) called the two units in his area the Fetvatn Gneiss and the Vikvatn Sequence respectively, further subdividing the latter into three separate units (Bryhni, 1977 and Lappin & Smith, 1978).

Current argument concerns the relative ages of the two sequences and the mode of formation of the eclogites and anorthosites. Carswell (1973) believed that the Jostedal Complex was younger than the overlying Fjordane Complex and that the latter was either intruded by, or thrust over the former. The Jostedal Complex gives ages of 1000 Ma. (Bryhni, 1973, 1977) while the ages on the Fjordane Complex are nearer 1800 Ma. (Bryhni, 1973, 1977). Both complexes show a retrogression, identified on the basis of individual mineral ages, between 500-350 Ma., showing they have been subject to a Caledonian reworking. However, the most recent dating seems to indicate that the Jostedal Complex too is older than previously thought, as it has given ages of 1700 Ma. (Krogh, 1977). The younger dates of 1000 Ma. come from a suite of granites, e.g. the Hestbreppigan granite (Banham & Elliot, 1965; Banham, 1968 and Priem et al., 1973), which intrude this older part of the Jostedal basement. The gneisses then seem to be the produce of some ancient (pre-1000 Ma.) orogeny(s), involving a basement with a sedimentary cover sequence. It is now clear that the rocks of the Fjordane Complex are not Eocambrian or Cambro-Silurian as was once thought but represent an older Precambrian metasedimentary sequence (Brueckner, 1979).

The eclogites found in this complex are an early feature in the crust, being involved in the earliest recognised deformations (Brueckner, 1977). There are, however, two schools of thought as to their origin. Griffin and Bryhni (1977) and Krogh (1977), believe the eclogites in the gneisses
Fig 7.

THE UNITS OF THE N.W. GNEISS AREA

FJORDANE COMPLEX
(Fetvatn Gneisses)

Migmatized Metasedimentary Complex
+ Amphibolites + Anorthosites + Eclogites
[pre 1500 Ma]

Unconformity?

JOSTEDAL COMPLEX
(Vikvatn Sequence)

Homogeneous Two Feldspar Gneisses
[1800 - 1700 Ma]
were formed in situ during the regional metamorphism of a variety of basic rocks. A Precambrian continent-continent collision is involved where the environment was typified by high pressure and a very low pH2O. The other school follows Eskola (1921), and is led by Lappin and Smith (1978), who examined the mineral phases using microprobe techniques. They conclude that the eclogite and anorthosite rocks crystallised at temperatures of around 1200-1370°C and pressures of around 30-40 Kb, implying therefore a mantle origin. They propose that these bodies of rock, occurring as aligned strings of boudins in the gneisses, which themselves do not show such high grades of metamorphism, are mantle wedges brought into the crust during an orogenic episode.

The Bergsdalen and Jotun Nappe Complex

This complex (areas 4 & 5 of fig. 4), comprises several nappe units of varying lithology and metamorphic grade. The complex is preserved in a synformal structure which runs from the coast inland on a north-easterly trend following the Hardanger Fjord. The dips along the north-western margin of the structure are steeper than those along the south-eastern border, where rivers have cut long re-entrants into the nappe pile, forming windows through the structure.

In the eastern portion of the complex, three major nappe units are recognised (Sturt & Thon, 1978), see figure 8. The highest unit of these, occupying the central part of the synform, is the Upper Jotun Nappe (Goldschmidt, 1916; Strand, 1972). This consists of a variety of rock types (anorthosites, mangerites, jotunites and pyroxene granulites) (Griffin, 1971 & 1977; Battey & McRitchie, 1975). Because of similarities in lithologies with rocks of the Bergen Arcs these are therefore referred to as the Bergen/Jotun kindred rocks. The rocks of the Upper Jotun Nappe preserve evidence of an early granulite facies metamorphism predating
Nappe Sequence Eastern Jotunheimen, Southern Norway.

<table>
<thead>
<tr>
<th>TECTONOSTRATIGRAPHIC UNIT</th>
<th>MAIN LITHOLOGIES</th>
<th>FOSSIL EVIDENCE</th>
<th>IGNEOUS ROCKS</th>
<th>CHARACTER OF DEFORMATION &amp; METAMORPHIC GRADE</th>
<th>RADIOMETRIC AGES</th>
</tr>
</thead>
<tbody>
<tr>
<td>OLYSVATN UNIT (Formation)</td>
<td>Sandstone, polymict conglomerates</td>
<td>None</td>
<td>None</td>
<td>Polycrystalline deformation greenschist facies metamorphism (probably Caledonian)</td>
<td>None</td>
</tr>
<tr>
<td>ALLOCHTHONOUS NAPPE</td>
<td>None</td>
<td>None</td>
<td>Polycrystalline deformation with strong foliation (Bygde cleavage), Recumbent folds in the Malnes area, Caledonian low grade metamorphism, ( 6 % ) mineral ages in the range 428 - 1260 m.y.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>UPPER JOTUN NAPPE</td>
<td>Anorthosites, Plagioclases, Various garnet-gneiss facies and granites</td>
<td>None</td>
<td>Anorthosites, Plagioclases, gneisses, granites</td>
<td>Polycrystalline granite facies, cores forming in anorthosites. Anorthositic facies produced, Strongly developed crustal rocks during Caledonian thrusting, with greenschist facies metamorphism, ( 6 % ) mineral ages in the range 428 - 1260 m.y.</td>
<td></td>
</tr>
<tr>
<td>VALDRES OG MELLSEN GROUPS</td>
<td>Skiltings, sericite, millilitic breccias and some sandstones</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>ESPEDAL ROSSJOKOLLAN</td>
<td>Varied greenschists and other rocks resembling those of the Upper Jotun nappe.</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>LOWER JOTUN NAPPE</td>
<td>Quartz-sandstone facies</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>AUTHOCTHONOUS NAPPE</td>
<td>Quartz-sandstone facies</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>CAMBRIAN</td>
<td>Thickness ( 130 ) km., Alumina-chlorite, Dark sandy shale</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td></td>
</tr>
<tr>
<td>PRECAMBRIAN BASEMENT</td>
<td>Various greenschist facies in the basement</td>
<td>Polycrystalline deformation, Mostly Precambrian amphibolite facies, Also Polycrystalline greenschist facies in Tremolite amphibolite, Precambrian granite facies in Arenite area, None</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Fig. 8.

emplacement. Whole rock ages of around 1280 Ma. from the granulites date this metamorphism, whilst rocks recording retrogression due to nappe emplacement yield mineral ages of around 428 Ma. (Sturt & Thon, 1978).

Situated tectonically below the Jotun Nappe, the Valdres Nappe consists, in its lower portions, of Precambrian crystalline basement similar to that of the overlying Jotun Nappe (Hossack, 1972 & 1976; Heim et al., 1977). These crystalline portions are succeeded in places by a thick sequence of Eocambrian sparagmites which in turn pass up through a Cambrian sequence. The highest stratigraphical level preserved is middle Ordovician.

The lowest of the three nappes is the Quartz Sandstone Nappe (Heim et al., 1977; Sturt & Thon, 1978), which contains sediments ranging in age from late Precambrian to middle Ordovician.

The rocks of these nappes show strong internal deformation (Bjorlykke, 1978; Hossack, 1978). The nappes were apparently folded before emplacement (Sturt & Thon, 1978), which seems to have been a Caledonian event occurring under greenschist facies conditions. The thrust planes are always marked by extensive zones of cataclastic rocks and it is thought by many workers (Hossack, 1978), that these units are far travelled, the Upper Jotun Nappe having originated some considerable distance to the west of the present Norwegian coast. Such a provenance means that the nappe sequence must have travelled from beyond the west coast of Norway over the gneisses of the North-West Gneiss Complex. Other workers (Banham, Gibbs & Hopper, 1979) hold the view that the nappes originated beneath their present outcrops, and have moved to both the north and south. This model is consistent with the regional gravity anomaly in the area (Smithson et al., 1974).
The western and southern portions of the complex are surprisingly different from the patterns described above (fig. 3). The thick Valdres and Quartz Sandstone Nappes are not represented here and the rocks of the Bergsdalen Nappes (Kvale, 1946 & 1948), underlie the Jotun Nappe. Kvale (1960), subdivided these into two units, the Upper and Lower Bergsdalen nappes, separated by a thick unit of phyllites which are assumed to be of Lower Palaeozoic age.

The nappes are also separated from the basement along the north-west margin by similar phyllites (fig. 3), though these are probably phyllonites.

The Bergsdalen Nappes themselves comprise an assemblage of Precambrian gneisses and supracrustals (quartzites, conglomerates, meta-volcanics and their mylonitic equivalents). They are intruded by a series of granitic plutons of Precambrian age (1274 Ma. Pringle et al., 1975; Brueckner, 1972; Grey, 1978 pers. comm.). As with the northern part of the complex much of the early deformation predates the nappe transport, and took place at a higher metamorphic grade than that coeval with the nappe transport (middle greenschist in the phyllites and zones of cataclasis along thrust planes, Sturt & Thon, 1978).

There is much similarity between the rocks of the Bergsdalen Nappes and those found in the Samnanger Complex of the Major Bergen Arc. Similar rocks are also to be found all along the north-west margin of the outcrop of the Bergsdalen/Jotun Complex (Banham et al., 1979). This imparts an asymmetry to the rock types outcropping on either margin of the complex; the north-west edge is very different in rock types and has undergone much higher strain than the south-eastern margin.

In the south-west, the Bergsdalen Nappes are overlain by the Bergen Nappe Sequence, with a contact complicated by later folding (Sturt & Thon, 1978). This suggests that the Bergen Arc structure continued to develop after the Bergsdalen and Jotun Complex had formed.
The Telemark Basement Gneiss Complex

This area (region 6, fig.4), which underlies the whole of southern Norway to the south of the Bergsdalen/Jotun complex, comprises granites, gneisses, migmatites, metasediments, quartzites, metavolcanics, amphibolites and anorthosites. These rocks form a vast complex wherein the metasedimentary sequences occupy an arc, following the coast, round a central region in which the granites, gneisses and migmatites are to be found. The whole complex is Precambrian, yielding ages ranging from 1730-860 Ma. (O’Nions & Heier, 1972; Versteeve, 1975). In most respects this basement area is very similar to that north of the Bergsdalen/Jotun complex, in that it comprises a Precambrian metasedimentary and metavolcanic sequence of varied lithology overlying a more uniform gneissose basement into which much granitic material has been intruded, the two having been deformed together (Torske, 1977). While eclogites are not found in this southern basement, in the south the metamorphic grade does reach granulite facies and in all other respects the lithologies present in the southern basement are identical to those in the north-western gneisses, even including anorthosites (Carswell, 1973). The area is interpreted as representing an ancient cordilleran-type, orogenic segment that developed over a protracted period in Precambrian times (Torske, 1977).

To the north-east in the Hardangervidda-Ryfylke area the basement is overlain by a younger, Cambro-Ordovician autochthonous cover sequence. This is then overthrust by a sequence of nappes which are subdivided into five distinct tectono-stratigraphic units, two of which are attributable to the Caledonian depositional cycle. The other three nappes are composed of tectonised basement. These nappes continue down the west coast into the Stavanger area. In the east, the Telemark basement is involved in the upper Palaeozoic rifting of the Oslo Graben.
2. STRUCTURE OF THE AREA

The gneisses of north Osterøy (fig.10), are divided into the Northern Gneiss Unit and the Southern Gneiss Unit, along a line passing through Tysse, for the following reasons: the foliation in the Northern Gneiss Unit strikes to the north-east paralleling the Bergsdalen front and the structures seen within it are the oldest in the area. In contrast, within the Southern Gneiss Unit, which structurally overlies the Northern Gneiss Unit, the foliation strikes to the south - south-east paralleling the outcrop of the Major Bergen Arc and the structures seen within it are younger than those seen in the Northern Gneiss Unit.

The structural history of the two areas is described separately; in each case starting with the oldest feature seen. The relationships between the two units are discussed, together with the nature of the junction between them.

References are made to the strain ellipsoid in discussion of the deformations affecting the Osterøy gneisses. The properties of the ellipsoid are inferred from structures present in the rocks, shape fabrics; fold styles; general orientations or reorientations of visible structures; schistosities, veins and igneous intrusions, etc. At no time has it been possible directly to determine the shape of the ellipsoid by measurement and only in one instance (D3 of the Southern Gneiss Unit), has it been possible to define more precisely the nature of the ellipsoid.

2.1 STRUCTURAL HISTORY OF THE NORTHERN GNEISS UNIT

The deformation sequence in the Northern Gneiss Unit is summarised in figure 36.
Fig. 10.

NORTH OSTERØY

MAJOR

ROCK UNITS

NORTHERN GNEISS UNIT

SOUTHERN GNEISS

BERGEN ARC THRUST

TYSSÉ THRUST

GNEISS

Km

0 2.5 5

Strike & dip direction of dominant foliation.
2.1.1 Early Deformations and Production of the Early Foliation "S Early"

The oldest feature seen is a gneissose banding accompanied by a layer of parallel schistocity. Over most of the area this banding is destroyed by subsequent deformations, but it is preserved at (11.37, 34.85) and in exposures along the north and west coasts of Osterøy. The gneissose banding and accompanying fabric may represent primary igneous layering, migmatisation of an older complex, or earlier layering modified by an earlier tectonism.

The third of these possibilities seems the most likely because, while all the old rocks preserve the schistocity, only some of them are banded. The schistocity is best developed (as a penetrative amphibole alignment) in the more mafic components, while the banding (a compositional layering) is found in the more acid rocks. The banding is marked by leucocratic and melanocratic layers about 1-8 cm. thick (pl.1).

These features indicate that the banding and foliation are, in part at least, the product of an earlier deformation or deformations rather than a primary feature. For this reason the foliation is called "S Early" rather than "SO".

2.1.2 D1, The First Deformation Phase

Evidence for D1 is only sporadically preserved (fig.11), and is best seen at (11.37, 34.85). This is the first fold-forming deformation, the folds cropping out as tight intrafolial isoclines (pl.2).

It is possible to trace the developmental history of D1 using structures developed in syntectonically intruded igneous bodies, and their relationship to periodic pegmatite emplacement.
Fig. 11

NORTH OSTERØY

D1
PRINCIPAL D1
EXPOSURES

0  2.5  5
Km
TABLE OF D1 EVENTS

D1a
1) Folding of "S Early" to form F1.
2) Intrusion of basic igneous bodies.
3) Initiation of vein development.

D1b
4) Flattening of F1 into intrafolial isoclines.
5) Flattening of the igneous bodies.
6) The development of a new foliation-S1.
7) The development of an intersection lineation (mullions in places)-L1.
8) Termination of vein emplacement started in D1a, associated with the limb regions of F1 folds.

This chronology is primarily established on evidence using the basic to ultrabasic intrusions at (11.37,34.58 and 14.65,44.63). The "S Early" foliation is isoclinally folded (pl.2), but this folding does not involve the basic bodies themselves which must therefore be intruded along the fold limbs and postdate this folding. After intrusion the basic bodies were veined by pegmatites that also started to develop in association with the limb regions of the folds (pl.3). The first stage of the D1 deformation is thus separated from the second by two intrusive events.

The second stage of the D1 deformation essentially involved flattening the pegmatitic veins in the basic bodies and those branching at an angle to the axial planes of the F1 folds became folded (fig.12), and a penetrative fabric, S1, developed axial planar to the F1 folds. The fabric is defined by the deccusate texture of amphibole crystals. In fold noses its intersection with "S Early" produces a lineation (L1). Occasionally an associated, slight, coaxial mullion structure developed on the outer arc of some of the more leucocratic "S Early" bands (fig.13).
Fig 12.

ILLUSTRATION OF A POST F1 GABBRO AFFECTED BY S1 FLATTENING.

Coarse Grained Centre, With Igneous Texture

Gabbro

S1

F1

S 'Early'

S1 'Flattening' Folds

Fine Grained Amphibolitic Zone, With a Strong

S1 Schistosity

Late Acid Vein

Gneiss

0 M 1
Fig 13.

SIMPLIFIED GEOMETRY OF STRUCTURES PRODUCED IN D1

Plunge of F1 & L1

Limbs of F1 Folds

STRUCTURES ASSOCIATED WITH THE F1 FOLDS

0 M 0.5

Flaggy S1

L1 (Mullion)

L1 (Intersection)

Amphibolite
Intrusion

Penetrative
S1

Limb
Pegmatite

S 'Early'

Folded D1a
Pegmatite
During the folding of the pegmatite veins the basic bodies themselves were flattened and boudinaged in the plane of the Si schistosity. The amount of this flattening is difficult to determine because of post D1 events. The D1 flattening increases in intensity towards the margins of the larger bodies (up to 4 meters thick), and the original coarse-grained igneous texture is replaced in the flattened margins with a finer-grained tectonic fabric (fig.12).

During D1b the limbs of the F1 folds were attenuated and sheared out, leaving isolated fold noses. It is in the sheared limb areas that the new foliation is most strongly developed, along with foliation parallel pegmatitic veins (pl.4). In outcrop, therefore, the rock is divided into layers of relatively unmodified fold noses, separated by thinner layers, developed from the limbs, rich in pegmatites and showing none of the older structures. It is impossible to tell whether these pegmatites are, in Sederholm's (1967) terms, arterites or venites; they possess characteristics of both. Selvedges of mafic constituents (mainly amphibole now retrogressed to biotite) are developed at the pegmatite margins. The pegmatite vein emplacement continued throughout the flattening in the last formed shears (pl.4).

The trend of the F1 folds axes is difficult to determine because of the lack of suitable exposures and modification by later (post D1) deformations. The observed fold axes lie north-east - south-west and plunge in both directions depending on the geometry of the later folding. After removal of the later modification it is thought they were probably originally subhorizontal.
Discussion

Figure 13 shows the main structures developed in DL and their orientation. Exposures preserving evidence for this deformation are rare (fig. 11), being mainly confined to the northern and western coasts of the island. Inland, apart from the problem of weathering, which obscures detail in the exposures, the evidence for the deformation becomes masked by structures developed during post-DL deformation. The spatial occurrence of the unmodified exposures is not haphazard and is controlled by position on later structures.

In view of the relationships described above the DL deformation probably spanned a considerable time period, but grouping into one episode is justified on the grounds that the foliation developed in DLb is axial planar to the folds which began to form in DLa. The possibility that the DLa structures rotated into alignment with the DLb structures is recognised.

The model proposed is folding, followed by intrusion of basic lenses and pegmatites. Considering the progressive nature of the SI development subsequent to the F1 folding, it is likely that the igneous intrusions would be in the XY plane of the developing F1 strain ellipsoid; the plane of the SI schistosity. It also seems that because the axial planes of the F1 folds and the folds in the pegmatitic veins formed in D2b are parallel, the development of the DI strain ellipse during D1 was, after F1, achieved with very little accompanying rotation. "a", "b" and "c" for the F1 folds can, therefore be considered to be coincident with X, Y and Z for the DI strain ellipsoid.

2.1.3 The Period between DI and D2

After the last stage of the DI flattening (see above), and before the onset of the next phase of deformation, D2 (see below), two
important events, vein emplacement and the deposition of a sedimentary complex occurred.

The veining is best seen where sections at right angles to the dominant foliation are exposed. On the northern coastline, a set of veins is seen to cross-cut those developed in D1. They do not preserve the S1 fabric, but are folded by D2 structures and develop the D2 fabric (pls.5 & 6).

Other very coarse-grained, granitic veins (frequently very rich in quartz) are found, which are difficult to place in the structural chronology since they do not preserve fabrics well; even where fabrics occur they can be attributed to either D2 or subsequent deformations. A further complication arises because migmatitic veins with a similar composition were also formed during early D2.

The sedimentary complex

In the period between D1 and D2, a sedimentary complex was either deposited or structurally emplaced in the area (map 1). The complex, described in detail in section 3.2.2, was essentially psammitic in composition with minor amounts of quartzite and mica schist. It is extensively migmatised, and all but the most quartzitic members have been altered. The migmatites are, however, distinguished from the gneisses because they do not show the intrafolial D1 isoclinals and by the presence of quartzite. The basement gneisses appear to have been migmatised with the sediments, as at (11.27, 34.85) ghost D1 structures are seen in the migmatised gneiss.
Discussion

Two problems are raised: firstly, the relationship of the pegmatites to the deposition or emplacement of the sediments. No exposures in which the pegmatites intrude the migmatites were found, suggesting that the sediments postdate the pegmatites; secondly, the relative age of the sediments. If they had been deposited directly on their present basement, a basal unconformity would be expected. As no such unconformity has been identified it follows that either the unconformity is obliterated by the migmatization and subsequent deformations, or the sediment/basement contact is entirely tectonic; the sedimentary complex may have been transported into the area during the early stages of D2.

Thin section study of the quartzite units shows several phases of fabric development. This conclusion is supported by field observations, the rocks are seen to be extensively folded. Minor thrusting accompanying the folding could account for the outcrop pattern seen at (21.00, 39.00, map 1); alternatively, the pattern could represent a refolded major D2 or D3 schuppen zone, in which case the sediments form an allochthonous nappe emplaced before the folding. In the absence of definite evidence, however, it is felt that the sediments are largely autochthonous and that the pattern is produced by original sedimentary variation; lenses of pure quartzite in a less pure sequence.

A remaining problem is the timespan indicated by the veining episode(s) and deposition of the sediments. At the close of D1 the rocks were undergoing amphibolite facies metamorphism (section 3.5.2), and since the sediments are a surface phenomenon, a considerable period of uplift and erosion is indicated. This was punctuated in the early stages by phases of vein emplacement which represent the only feature attributable to this period. The gap between D1 and D2 events is therefore probably considerable and could span the period between one orogenic cycle and the next.
2.1.4 D2, The Production of F2, S2, S3 and L2

D2 is the second deformation event to produce folds. It can be subdivided into two phases (table below). The effects of the second episode are inhomogeneously developed across the area, increasing in intensity from west to east.

In the western half of Osterøy, F2 folds are the dominant structures and unlike the intrafolial F1 folds, deform the dominant foliation. They are the obvious folds seen in outcrop.

**TABLE OF D2 EVENTS**

**D2a**
1) Migmatisation
2) Production of a metasedimentary and migmatitic layering, S2
3) Minor pre-fold thrusting ?
4) Folding
5) Production of S3a
6) Production of L2
7) Pegmatitic vein emplacement

**D2b**
8) Intrusion of basic igneous bodies
9) Pegmatitic vein emplacement
10) Flattening and enhancement of the S3a fabric to form S3b. (This event was heterogeneously developed across the area, being strongest on the eastern side of the island.)

The migmatite phase affecting the sedimentary complex has been referred to in the previous section. Metamorphic segregation and the development of kyanite, almandine and hornblende in the metasediments suggests reworking at about 600°C and 20 Km burial (Miyashiro, 1973). During this metamorphism a gneissose banding develops sub-parallel to the original
colour banding seen in the more quartz-rich sedimentary units; these resisting migmatisation to a greater degree than the less pure components. The colour banding appears to represent sedimentary layering, the gneissose banding therefore probably developed tangential to the Earth's surface rather than normal to it, and is therefore the product of hydrostatic burial during metamorphism rather than lateral compression. This foliation is called S2 and is everywhere parallel to S1.

Except for a few arteritic (Sederholm, 1967) pegmatites derived from the migmatites, some of which are garnetiferous, the migmatite event in the metasediments is only poorly seen in the older gneisses, which already possessed a mineralogy stable at the almandine amphibolite facies of metamorphism. Within the sediments there are numerous pegmatitic veins produced during migmatisation; these are coarse-grained and very rich in potash feldspar. Examples of these are seen at (21.50,39.00), where dioritic granites intrude the migmatites. These diorites bear the migmatitic fabric S2, and are now augen gneisses. They are arteritic in origin, not a product of the migmatisation.

At locality (22.00,38.00 - map 1), schlieren of quartzite are present in the migmatites. As has been discussed, these could point to the presence of a hidden thrust. The picture is also confused because their contacts are sharp and apparently tectonic. However, the schlieren are not thought to be a schuppen zone but sedimentary/metamorphic in origin and the tectonic contacts are thought to be produced during F2, as the result of the first movements of folding.

**F2 Folding**

The F2 folds fold the S2 migmatite foliation and S1 (pls. 7 & 8), which as a composite structure form the dominant foliation of the Nord Nordland gneiss region. The F2 folds are asymmetrical and overturned,
verging north-west and trending north-east - south-west. The long limbs of the folds dip south-east forming the east slopes of the local mountains, while the short, steeply dipping north-west limbs give the topography a sawtooth appearance. The major folds have a wave-length of about 6 Km. Parasitic folds are very common in the nose regions of the main folds (e.g. the western coastline, 11.5°, 36.0°), and range from 1 to 30 meters in wavelength.

The plunge of the F2 folds is variable for the following reasons:

a) They are affected by later deformations, which modify the F2 geometry.

b) In the nose regions of the major folds the parasitic folds become non-cylindrical, the pitch of the axis, measured in the axial plane, the general dip of which is 110°/50°, varies by as much as 50°.

c) There is a variation in style of fold related to the rock type involved in the folding.

Figure 14 shows the variation in plunge for the F1 folds; the spread in the data is a function of the non-cylindricity, while the south-west plunges are produced by later folding. Removing later deformation, the general plunge of the F2 folds is to the north-east, at 045°/27°. On a large scale these folds are cylindrical. On the small scale however, especially where the rocks involved have high ductility contrasts (such as between quartzite and migmatite; and amphibolite and migmatite) and in the nose regions of the major folds (where the strain is constrictional) the parasitic folds become non-cylindrical.

On Osterøy two large scale F2 fold closures are seen, one synformal and one antiformal (fig. 15). The geometry of the synform is altered by later thrusting, as is seen from the outcrop pattern of the quartzites (fig. 26). The thrusting is also responsible for repetition of the antiform closure (fig. 15).
Fig 14.

**Plunge of F2 Folds**

The contours represent the following percentages:

- 0.9 ≤ 1-2% ≤ 2.9
- 3-4% ≤ 5-6%
- 7-8% ≤ 9-10%
- 11-12%
- 13-14%
- 15-16%
- 17-18%
- 19-20%
- 21-22%
- 23-24%
- 25-26%
- 27-28%
- 29-30%
- 31-32%
- 33-34%
- 35-36%
- 37-38%
- 39-40%
- 41-42%
- 43-44%
- 45-46%
- 47-48%
- 49-50%
- 51%

Note: A value of 0.5-0.9 is left and right 0.0.
Fig. 15

NORTH OSTERØY

D2

F2 FOLD TRACES AND QUARTZITE OUTCROPS

0 2.5 5 Km

N
A new foliation, S^a, developed with the folding. It is first apparent in the quartzites where a muscovite and potash feldspar schistosity, best seen in thin section, develops parallel to the axial plane of the F2 folds, (the biotite of S2 tends to recrystallise) (fig.16). In the gneisses the S^a schistosity is marked by small flattened pods of amphibole, now mainly replaced by biotite, which crosscut the earlier S1 and S2 foliations in the nose regions of the F2 folds (pl.11).

An intersection lineation, L2, forms between the dominant S1-S2 foliations and S^a (pl.9). The plunge of the lineation is parallel to the F2 folds and may show a mullion-like character in the nose regions, especially in the quartzites. Figure 17 is a plot of all the lineations found in the Northern Gneiss Unit; the majority of the readings plotted are L2, and it is clear that the orientation is similar to that for the F2 plunges (fig.14). The spread of the readings is not a reflection of the non-cylindricity, because the readings were mostly made on the limbs rather than hinges of the F2 folds. The spread is attributed to the effects of later folding (F3, see below), refolding this D2 lineation so that it now plots on two small circles round the hinge of the later (F3) folds (fig.25).

Pegmatite Relationships

The relationship between the F2 folds and the various pegmatitic veins allows a detailed subdivision of the D2 event. Four groups of veins can be differentiated:

a) The D1 limb region pegmatitic veins, now seen to be folded round the F2 noses along with refolded F1 folds (pls.10 & 11).

b) The pre-migmatite pegmatites, which are very difficult to identify (see p.34), (pl.5).
Fig 16.

**DEVELOPMENT OF S3a SCHISTOSITY**

**SKETCH OF HAND SPECIMEN (FH B39)**

- S3a
- L2 Intersection Lineation
- F2 Folds
- Quartz Matrix
- Biotite Foliation S2
- S3a (Feint)

**SKETCH OF THIN SECTION**

- F2 Microfold
- Late Quartz Lath Cleavage S5 (D4)
- Folded Biotite Foliation S2
- Muscovite Books Aligned With S3a
- Quartz Matrix
- K-Feldspar S3a
Fig 17.

Plot of all N. Gneiss Lineations

The contours represent the following percentages:

- 0.9% ≤ 1.2%
- 1.2% ≤ 2.5%
- 2.5% ≤ 5%
- 5% ≤ 6.5%
- 6.5% ≤ 8.5%
- 8.5% ≤ 9.9%
- 9.9% ≤ 11%
- 11% ≤ 12%
- 12% ≤ 13.5%
- 13.5% ≤ 15%
- 15% ≤ 16.5%
- 16.5% ≤ 18%
- 18% ≤ 20%
- 20% ≤ 22.5%
- 22.5% ≤ 25%
- 25% ≤ 27.5%
- 27.5% ≤ 30%
- 30% ≤ 32.5%
- 32.5% ≤ 35%
- 35% ≤ 37.5%
- 37.5% ≤ 40%
- 40% ≤ 42.5%
- 42.5% ≤ 45%
- 45% ≤ 47.5%
- 47.5% ≤ 50%
- 50% ≤ 52.5%
- 52.5% ≤ 55%
- 55% ≤ 57.5%
- 57.5% ≤ 60%
- 60% ≤ 62.5%
- 62.5% ≤ 65%
- 65% ≤ 67.5%
- 67.5% ≤ 70%
- 70% ≤ 72.5%
- 72.5% ≤ 75%
- 75% ≤ 77.5%
- 77.5% ≤ 80%
- 80% ≤ 82.5%
- 82.5% ≤ 85%
- 85% ≤ 87.5%
- 87.5% ≤ 90%
- 90% ≤ 92.5%
- 92.5% ≤ 95%
- 95% ≤ 97.5%
- 97.5% ≤ 100%

Note: A value of ≥ 0.5% LT 12% and ≥ 0.2%
c) The migmatite pegmatites and granites (pl. 12)

d) A suite of pegmatitic veins developed during the F2 folding, which lie along the axial planes of the folds (pls. 10 & 11).

All these pegmatitic veins show the S3a schistosity, making detailed subdivision of the older veins difficult. Accompanying the axial planar pegmatites, in the long limbs of the F2 folds, a suite of gabbroic igneous bodies was intruded (see 22.00, 34.00). After intrusion further pegmatitic veins were emplaced. These coarse-grained, quartz and potash feldspar rich pegmatites cut the gabbros, F2 folds and S3a schistosity at moderately high angles, but are deformed in the D2b events (fig. 21).

The D2b Flattening

The final phase of the D2 deformation is a flattening event. Unlike the D1b flattening, which is homogeneous across the island, it increases from west to east of Osterøy (cf. pls. 7, 8 & 10 from the west coast, centre and east of the island). As the F2 folding becomes tighter, the S3a schistosity is more strongly developed. Further east the new schistosity becomes more intense, replacing S1 and S2, it forms the dominant foliation. It imparts a flaggy nature to the outcrops on the east of Osterøy, as the F2 folds become intrafolial to S3b (intensified S3a). On Osterøy the flattening increases with altitude and eastwards (fig. 18).

It is apparent from this that the deformation increases in intensity along a vector normal to the F2 axial planes, towards the east. The build-up to this flattening is gradual, the first effect seen is folding of the late pegmatites discussed above (pl. 13). Higher up the structural section as seen in the mountains, the flattening affects the igneous bodies and their cross-cutting pegmatites, which are boudined within
Fig 18.

DEVELOPMENT OF D2 STRUTURES ACROSS N. OSTERØY

Fig 19.

DEVELOPMENT of the FLAGGY S3b SCHISTOSITY
the gabbros (pl.14). Higher still, and further east the S3b foliation becomes dominant (pl.10), the foliation appearing first in the middle limbs of the F2 folds (fig.19). On the east slopes of Hogafjell (fig.9), the D1 structures start to be overprinted.

Discussion

The problems which arise when considering the D2 deformation are similar to those for D1. The grounds for grouping two phases of deformation, D2a and D2b, are that the S3b foliation is parallel to that of S3a and the development of S3b is associated with the tightening of the F2 folds, which formed in D2a.

The heterogeneity of the S3b development across the area must be of fundamental significance, it increases in intensity towards the Bergsdalen front. This observation can be interpreted in two ways:

1) The D2 deformation was associated with emplacement of the Bergsdalen Nappes and is therefore the first indication of their emplacement (see fig.20a).

2) That the site now occupied by the Bergsdalen Nappes formed a fundamental feature in the local basement, which preceded emplacement of the Bergsdalen Nappes (see fig.20b). Deformation on this feature would be responsible for the alignment of the D1 structures and the F2 folds. The intensity of the D2b flattening may be a function of absolute distance from this feature.

The sense of translation on the shears in the middle limbs of the F2 folds and their vergence is to the north-west. This, and the fact that the D1 folds axes parallel those of D2 and the presence of a further phase of folding (F3 below), indicate Figure 20b (OR) probably best explains the observations.
Fig 20a

MODEL 1 EXPLAINING THE SIGNIFICANCE OF F2 & S3b

Emplacement of the BERGSDALEN at the end of D2

Fig 20b

MODEL 2 EXPLAINING THE SIGNIFICANCE OF F2 & S3b

First in D2

Major Crustal Feature Here in D2, an Orogen?

Then Later, D3? D4?

EITHER

BERGSDALEN Emplaced From The NW

OR

BERGSDALEN Emplaced From The SE
Like the D1 deformation, D2 obviously spans a considerable period. The strain ellipsoid for the event is complex; development of D2 structures is tabulated below:

a) Initial extension parallel to the layering (D1) accompanying migmatisation, during which pegmatites were intruded.
b) Folding of the S1/S2 layering to produce F2.
c) Formation of a new schistosity in the X/Y plane of the F2 strain ellipsoid.
d) Injection of gabbro and pegmatite emplacement parallel to the X/Y plane of the F2 strain ellipsoid.
e) Pegmatite emplacement in the Y/Z plane of the F2 strain ellipsoid (see fig. 21).
f) Further modification of the F2 strain ellipsoid during D2b.

This sequence of development is very similar to that seen in D1 and is summarised in figure 22. (Again a, b and c for the folds are assumed to be coincident with X, Y and Z for the D2 ellipsoid).

2.1.5 D3, The Production of F3, L3 and S4

D3 is the third deformation seen in the area, it is subdivided into several episodes.

TABLE OF D3 EVENTS

1) Thrusting
2) Production of the F3 folds
3) Development of S4
4) Development of the L3 lineation
5) Pegmatite emplacement
D2 PEGMATITES AND THEIR RELATIONSHIP TO THE F2 FOLDS

Fig 21

Pegmatites in b c Plane of the F2 Folds, Folded in D2, During the D2b Flattening

(D3 Pegmatite)

(D3 Fold)

D1 Pegmatite

S3b in a b Plane of the F2 Folds

0 M 1
SYNOPTIC DIAGRAM OF STRUCTURES ASSOCIATED WITH THE F2 FOLDS
While the deformation produces some of the most obvious folds in the Northern Gneiss Unit, F3 folds only become dominant structures further east on the mainland. Like the D2b flattening, the structures produced in D3 are heterogeneously developed across the island; they are absent in the west, first seen in the centre and are moderately developed in the east.

Figure 23 shows the dominant trend of the D3 structures on north Osterøy, where they are confined to the areas where the earlier D2 deformation and S3 foliation are best developed. The F3 folds fold the D2 foliations and the D2 amphibolites (pls. 14 & 15).

The first features associated with D3, the thrusts seen on figure 23, could in fact more properly belong at the end of D2. Dipping steeply to the south-east these thrusts cross-cut the F2 folds (figs. 15, 23 & 26), displacing their axial planes. The effects of the thrusting on an F2 synformal closure are seen in the quartzite outcrop at (17.00, 33.00 & 19.20, 25.50); the fold nose is cut out leaving a "pipe" of quartzite running through the mountain (fig. 26). Although the thrusts are sub-axial planar to the F3 folds they are in fact slightly folded by F3; it is this which places them at the beginning of D3 or at the end of D2. The sense of translation on the thrusts, inferred from the displacement of the F2 axial planes, is from the south-east to the north-west. Typical displacements measured on the map are of the order of 100s of meters.

The F3 folds are asymmetrical and open, verging slightly to the north-west. The degree of asymmetry and amount of overturning increases eastwards. The folds are developed on a large scale, between 20 meters and 500 meters in wavelength. The open geometry and large scale of the folds makes it almost impossible to measure plunge and axial plane orientations
Late D2 or Early D3 THRUSTS

THRUSTS AND F3 FOLD TRACES

TYSSE THRUST

F3 Axial Traces

BERGEN ARC THRUST

Fig. 23

NORTH OSTERØY
directly, these are therefore determined from readings made on the S3 schistosity, where this is strongly affected by the F3 folding.

Figure 24 indicates that the folds plunge to 080/11° and the axial planes dip at 170/85°. In the east of the island some of the folds are much tighter and develop shears on their middle limbs. In places the folds are non-cylindrical, as is seen from the outcrop pattern at (21.50,28.50), but this seems to be atypical. Elsewhere the folds have a more cylindrical geometry (pls.16 & 17). The F3 folds are responsible for the outcrop pattern in the amphibolites at (20.00,32.00), (fig.112).

In the east the middle limbs of the sheared folds begin to develop a new axial plane schistosity, S4 shown by hornblende. This is very weak on Osterøy but strengthens eastwards, where, with the F3 axial plane, it rotates to give progressively shallower dips to the east, as the F3 folds become more overturned.

A strong mullion rodding, L3 tends to develop with the F3 folds even where they are open and S4 is weak (pl.16), this becomes more pronounced as the folds tighten. The mullion development enhances the older lineations (L1 and L2), over most of the area, and in the east of the island it is L3 itself which forms the strongest feature of the topography. It is very difficult to distinguish between L2 and L3 in outcrop and figure 17 is therefore an amalgamation of both taken from the area as a whole. However, the two fold phases (F2 and F3) are not strictly parallel; F3 plunges to 080° while F2 plunges to 045°, thus enabling separation of the L3 component on figure 17; the cluster seen at 075° is attributable to D3, while the general skewed spread of the points is caused by the D2 lineations. They are folded round the F3 axis on a small circle (fig.25), (also allowing for variation caused by the non-cylindrical nature of some of the F2 and F3 folds). The refolding also explains some of the spread in F2 plunges in figure 14.
POLES TO S3 DEFINING THE F3 FOLDS

The contours represent the following percentages:

0.0 - 0.9 = 0.5
1.0 - 1.9 = 1.0
2.0 - 2.9 = 1.5
3.0 - 3.9 = 2.0
4.0 - 4.9 = 2.5
5.0 - 5.9 = 3.0
6.0 - 6.9 = 3.5
7.0 - 7.9 = 4.0
8.0 - 8.9 = 4.5
9.0 - 9.9 = 5.0
10.0 - 10.9 = 5.5

NOTE A VALUE OF > 0.5 - 0.9 = 0.5 AND 1.0 - 1.9 = 1.0

This plot has undergone -0- rotations using coordinate card 1.
THE RELATIONSHIP BETWEEN L2 AND L3

F3

L3 Mullion Rodding

S3 Foliation Plane

L2 Refolded Round F3

Intrafolial F2

Small Circle
Spread of the L2 Lineation

INTERPRETATION OF Fig. 17.

L3 and Plunge of F3
SCHEMATIC DIAGRAM TO SHOW HOW THE LATE D2 OR EARLY D3 THRUSTS CROSS-CUT THE D2 STRUCTURES.
Following the F3 folding there is evidence in a few localities for a further phase of pegmatite emplacement in the Northern gneisses (fig. 21). The pegmatites are granitic in composition and intruded along the axial planes of the F3 folds at a high angle to the foliation, e.g. (11.37, 34.85).

The last structural feature attributable to D3 is further thrusting. The thrusts are only seen on the small (outcrop size) scale and are the result of further development of the F3 folds, shearing out their middle limbs. The effects of this shearing increase south-eastwards where the folds are at their best developed and tightest. The translation (from the direction of overturning and displacement) is again from the south-east to the north-west.

Discussion

The most significant feature of this deformation is the orientation of the structures and their sense of translation. The repetitive history of folding along a north-east axis, observed through D1 to D3; the consistent north-westerly direction of thrusting and overturning of the fold axial surfaces, points to transport from the south-east, and is consistent with model 2 (fig. 20) proposed for D2.

The proposed development of the D3 structures was:

a) Thrusting ? shortening of the layering to produce F3.

b) A new schistosity and foliation developing in the east where the F3 folds are tightest, the schistosity forming in the X/Y plane of the F3 strain ellipsoid.

c) A new lineation L3 developing parallel to the Y axis and enhancing that previously developed in D2 (L2) because of the nature of the interference pattern between F2 and F3, (a Ramsay (1967) type 3 pattern) as proposed by Watterson (pers. comm. 1977), (fig. 27).
Fig 27.

PRODUCTION OF A PENETRATIVE TECTONITE FABRIC BY SUPERIMPOSED STRAINS. [After WATTERSON 1977]

Principal strain axes for

\[ \text{Da} \quad \text{and} \quad \text{Db} \]

which, when superimposed produce finite strain axes:

Finite Situation Rodding Structure
Aligned Parallel to the 'Y' Axis of the Folds
d) Pegmatite emplacement along the X/Y plane of the F3 strain ellipsoid (fig. 21).

e) Minor ductile thrusting in the form of shearing along the X/Y planes of the F3 strain ellipsoid (the axial planes of the folds - the middle limbs were sheared out).

This sequence of development is summarised in figure 28.

2.1.6 D4, The Production of F4 and S5

The D4 deformation was the last affecting the Northern Gneiss Unit. It is distinguishable from the preceding deformations primarily because D4 structures are orientated at right angles to the D1-D3 trends.

TABLE OF D4 EVENTS:

1) Development of F4 folds
2) Local development of S5 schistosity and foliation
3) Development of quartz veins.

In general the trend of the D4 structures in the Northern gneisses is around 135° (fig. 29), but the folds are highly non-cylindrical and the pitch of the axis measured in the axial plane varies by as much as 90° over distances of 10 meters. The folds are asymmetric and overturned, verging to the south-west.

Except where the foliation has a shallow dip, the F4 folds fold rocks in which the dominant foliation was aligned close to the X/Z plane of the D4 strain ellipsoid. Well developed F4 folds are therefore very rare in the Northern Gneiss Unit. D4 folds are better developed in the Southern Gneiss Unit, and it is clear that F4 folds in the Northern gneisses are not only restricted to areas where the foliation was almost flat, but also to bands across the Northern gneisses. These bands are seen
Fig 28.

**SIMPLIFIED GEOMETRY OF THE STRUCTURES PRODUCED IN D3**

- Steep Limb
- L3 and Plunge of F3
- F3 Axial Plane, S4 and Late D3 Thrusts
- Early D3 or Late D2 Thrusts
- Shallow Limb

**STRUCTURES ASSOCIATED WITH THE F3 FOLDS**

- Early D3 or Late D2 Thrust Plane
- Intrafolial F2
- Open F3
- Incipient Thrusting in Middle Limb Late in D3

X,Y,Z = F3 Strain Axes

- F3 Axial Planes
- Tight Non-cylindrical F3
- Late Axial Planar D3 Pegmatite
Fig. 29.

NORTH OSTERØY

D4

STRUCTURAL TRENDS

Near Vertical Shear Zones and Trend of F4 Fold Axes

0 2.5 5 Km
in figure 30, which is a computer generated surface representing a
gliation surface across north Osterfj, constructed to enhance east-
west trending structures and suppress the effects of structures in
other orientations. (a) shows the resultant surface as a block diagram,
(b) as a contour map. The bands represent the periodicity (wavelength)
the folding would tend to assume on the larger scale. D4 structures
are therefore best developed along these bands. The folds are impersistent
and die out over short distances, though great numbers of parasitic folds
generally occur together (e.g. 11.90, 35.50). (A further account of
these folds will be given in section 2.2.4 on the Southern Gneiss Unit).

Shear zones on the same general trend are also to be found throughout
the Northern Gneiss Unit (pl.18). These zones are generally between
0.05 and 0.5 meters wide and 10 meters long, though much larger examples
(5-20 meters wide) were found in the north of the island (pl.19).
The pre-D4 foliation is deflected sharply on entering these bands. In
the shear zones the texture and mineralogy is altered to produce a new
schistosity and foliation: S5. Figure 31 shows examples of these
shear zones from which it is clear that they are associated with the F4
folds (pl.20). The shear zones form either on the middle limbs of
partially developed F4 folds or replace them. The displacement across
the shears indicates a translation from the north-east to the south-west,
consistent with the sense of translation on the F4 folds.

Within the quartzites (fig.15), there are thin bands of mica schist.
The schistosity developed during D2 and D3, and is parallel to S3. In
D4 a strong crenulation cleavage develops in these, although few F4 folds
are seen. Microscopically the geometry of this crenulation cleavage
is identical to that of the D4 shear zones (fig.32). These can therefore
be regarded as analogous to the crenulation cleavage, which is orientated
035/17° in the mica schist.
Fig 30.

PERIODICITY OF THE MAJOR F4 FOLDS ACROSS THE N. GNEISSES

a

b

PLOT OF THE FOLIATION IN THE NORTHERN GNEISSES USING WDPTM.
AZIMUTH = 45°   ALTITUDE = 65°
SMOOTHING = -1.00   #HEIGHT = 6.00

= BEFORE FOreshortening 05/03/79
EXAMPLES OF D4 SHEAR ZONES AND S5

F4 Middle Limb Shear

Three Dimensional View

Profile View of an F4 Fold
Showing Shear & Quartz Vein

Oblique View of D4 Quartz Vein
HAND SPECIMEN [FH B77] OF MICA SCHIST SHOWING THE
F4, S5 CRENULATIONS

SKETCH OF THIN SECTION [FH B54] SHOWING AN OBLIQUE VIEW
OF AN F2 FOLD REFOLDED ROUND AN F4 FOLD AND FORMING
THE S5 CRENULATION CLEAVAGE
Along the axial planes of the F4 folds and within the shear zones it is common to find large lensoid quartz veins, ranging from 0.01 meters wide, 0.3 meters long, to 1 meter wide, 30 meters long (pl.21 and fig. 31c & d). These veins are very common and form a characteristic feature of the Northern gneisses. Like the shear zones they are related to the F4 folding. Figure 33 shows that the majority of the veins are vertical and strike at 145°, the spread is the result of the non-cylindricity of the F4 folding (see below). The plots near the centre of the net are of veins with a morphology similar to that seen in figure 31c.

Discussion

Several relationships were observed between the F4 folds, shears and quartz veins:

1) Folding without associated shears or quartz veins
2) Folds with shears in their middle limbs
3) Shearing with little preceding or accompanying folding (abrupt shears) (fig.31b)
4) Folds with sheared middle limbs, cut by quartz veins
5) Shears cut by quartz veins
6) Quartz veining with little or no preceding deformation.

These observations can be explained by assuming a model for D4 where the deformation occurred in a serial manner (fig.34). The folds were the first structure to develop, after this shearing became the dominant deformation mechanism. Shears formed in the middle limbs of the F4 folds, while elsewhere (where the S3 and S4 foliation is more orientated in the X/Z plane of the D4 strain ellipsoid), they affected undeformed rock. Next the quartz veining modified the earlier F4 folds and shears, but also occurred in new localities in the rock.
Fig 33.

POLES TO D4 QUARTZ VEINS

PLOT OF QUARTS VEIN ORIENTATIONS

THIS PLOT HAS UNDERGONE -0- ROTATIONS
USING COORDINATE CARD 0

SAMPLING 2ACROSS/-1DOWN POLES PER 1PERCENT AREA 97 DATA EQ AR

THE CONTURS REPRESENT THE FOLLOWING PERCENTAGES

0.9<=1-2%>2.9<=3-4%>4.9<=5-6%>6.9<=7-8%>8.9<=9-10%>10.9<=11-12%>12.9
12.9<=13-14%>14.9<=15-16%>16.9<=17-18%>18.9<=19-20%>20.9<=21-22%>22.9
22.9<=23-24%>24.9<=25-26%>26.9<=27-28%>28.9<=29-30%>30.9<=31-32%>32.9
32.9<=33-34%>34.9<=35-36%>36.9<=37-38%>38.9<=39-40%>40.9<=41-42%>42.9
42.9<=43-44%>44.9<=45-46%>46.9<=47-48%>48.9<=49-50%>50.9<=51%

NOTE A VALUE OF:0.5-0.9<=LT.1%AND.GT.0%
OBSERVED RELATIONSHIPS BETWEEN THE F4 FOLDS, S5 AND THE D4 QUARTZ VEINS


2. Older Foliations

S5 Developed in the D4 Shears

3. No Accompanying S5
The D4 deformation also affects the Southern Gneiss Unit rocks (see below). Here the folds are better developed and shears and quartz veins are uncommon. The reason for this is that the pre-D4 foliation in the Northern gneisses strikes at right angles to the F4 axial planes and the pitch of the intersection of the foliation on the axial plane generally exceeds 45°. The higher this angle the more difficult the foliation is to fold, and the more restricted the style of structure that can develop (Ramsay, 1967; Cosgrove, 1976): (By analogy it is like trying to fold the spine of a telephone directory, edge on, over the edge of a table). The deformation started as a flexural slip buckling of suitably orientated layers, producing class 1a-c folds (Ramsay, 1967). Once these layers had folded, because the unsuitably orientated layers could not fold, shearing took over and the shear zones were formed. The quartz veins formed at a late stage in the process during the decay of the stress field, the pore fluid pressures having been "pumped up" by the early high stresses. The rock hydraulically fractures along the planes of weakness established during the earlier part of D4; these were the only planes with the right orientation in the waning D4 stress field.

The development of the structures seen in D4 are (fig. 35):

1) Compression of the layering to produce F4.

2) Further shortening, producing shear zones, a crenulation cleavage and the S5 schistosity, parallel to the X Y plane of the F4 strain ellipsoid.

3) Hydraulic fracture and the generation of quartz veins in the X Y plane of the F4 strain ellipsoid along zones of incipient F4 and S5 deformation.
SIMPLIFIED GEOMETRY OF STRUCTURES PRODUCED IN D4

Variation in F4 Axial Plane
Attitude
Zone of Possible F4 Plunges
Crenulation Cleavage S5
Average Shallow Limb
D4 Shears and Quartz Veins
F4 Plunge
Average Shallow Limb

STRUCTURES ASSOCIATED WITH THE F4 FOLDS

Non-cylindrical F4
Axial Plane of F4 and S5 Crenulation Cleavage
Composite Older Foliations [F2 or F3] (steep)
D4 Quartz Vein
S5 in a D4 Shear Zone
2.1.7 Discussion on the Deformation of the Northern Gneiss Unit

Figure 36 summarises the deformation history of this unit, in which several distinct trends can be seen. Firstly, deformation was cyclic. The basic cycle seems to be:

A) Undeformed foliation.
B) Pegmatite emplacement at a high angle to the foliation.
C) Folding (open folds which tend to have a parallel geometry).
D) Pegmatite emplacement and igneous injection parallel to the axial planes.
E) Pegmatite emplacement at a high angle to the axial planes.
F) Flattening of all the above, (folding of (E) pegmatites), ( (C) folds become isoclinal with a similar geometry).

This sequence of igneous and tectonic activity characterises one "deformation" cycle. Occasionally, while all the structural characteristics of a cycle may be present, an igneous phase may be absent, presumably because of an independent non-tectonic control. It is important to note that when present the igneous bodies are usually intruded in or close to the plane containing the maximum and intermediate stress axes, as defined by the geometry of local macroscopic or mesoscopic structural features.

Many authors (Price, 1970, 1975; Price & Hancock, 1972; Blay et al., 1977; Dubey & Cobbold, 1977), believe that folding is a sequential diachronous phenomenon. Folds may develop one at a time in a fold train, so that of any two adjacent folds one is older than the other, even though they belong to the same deformational event. Using this theory it is possible to explain the repetition of the deformational sequence (A-F above) observed in the Osterfjy basement.
### Fig 36. Deformation in the Northern Gneisses

<table>
<thead>
<tr>
<th>D “Early”</th>
<th>D 1</th>
<th>D 2</th>
<th>D 3</th>
<th>D 4</th>
<th>Stage in Fig 37</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td><strong>F 1</strong></td>
<td><strong>F 2</strong></td>
<td><strong>F 3</strong></td>
<td><strong>F 4</strong></td>
<td><strong>C</strong></td>
</tr>
<tr>
<td><strong>S 2</strong></td>
<td><strong>S 3a</strong></td>
<td><strong>S 4</strong></td>
<td><strong>S 5</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Igneous</td>
<td><strong>L 2</strong></td>
<td><strong>L 3</strong></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>activity</td>
<td></td>
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<td></td>
<td></td>
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</tr>
<tr>
<td>and</td>
<td></td>
<td></td>
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<tr>
<td>Pegmatites</td>
<td></td>
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<td></td>
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<td></td>
</tr>
<tr>
<td>Flattening</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>D</strong></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>E</strong></td>
</tr>
<tr>
<td>S “Early”</td>
<td><strong>S 1</strong></td>
<td><strong>S 3b</strong></td>
<td><strong>Thrust</strong></td>
<td></td>
<td><strong>F</strong></td>
</tr>
<tr>
<td><strong>L 1</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>A</strong></td>
</tr>
<tr>
<td>Pegmatites</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td><strong>B</strong></td>
</tr>
</tbody>
</table>

**S 2** indicates migmatisation of sediments.
The model proposed makes some assumptions; that the stress strain compatibility equation \( \sum_i C^2_i = 0 \) (Jaeger, 1956; Ramsay, 1967), holds during the deformation under the crustal conditions pertinent to the north Osterfy gneisses; that the stress ellipsoid is the reciprocal of the strain ellipsoid (Ramsay, 1967); that on the broad scale the maximum principal stress axis will be normal to the Earth's surface for the majority of geological time (Price, 1966; Dietrich & Carter, 1969).

Evidently during periods of crustal shortening, it will be orientated sub-parallel to the Earth's surface.

The model relates the deformation seen in the Northern gneisses, (D1-D4), to the deformation sequence (A-F) (fig. 36), by considering how the regional stress ellipsoid may develop compared to the local stress ellipsoid during an orogeny. Figure 37 summarises the model and shows that while it is possible to erect a chronology of events for the deformation and intrusion relating to one fold, this chronology is not absolute because the deformation and intrusion are diachronous; (while the axial plane pegmatites are emplaced at one locality, folds are just forming in another). The model also describes a single tectonic event, providing justification (from theory as well as observation) for the grouping of deformational features into D1-D4.

Clearly the Northern gneisses have undergone at least three of these complete tectonic cycles up to D4, at which point the record of deformation ceases. Of the three earlier cycles D2 occurred at the deepest crustal level and D4 at the shallowest (see section 3.5 below).

Another characteristic of the deformation in the Northern gneisses is the flattening that precedes the folding; this is explained in the model proposed, but in the case of D2 and D3 there is a further complication, the flattening is heterogeneous. Plate 22 shows leucocratic layers in the
FOUR STAGE STRESS MODEL TO EXPLAIN THE RELATIONSHIP BETWEEN BASEMENT STRUCTURAL FEATURES

A

Usual Situation

Earth's Surface

B

Buckling Situation

C

New Cleavage

D

Local Stress Field

Regional Stress Field During Deformation

E

Flattening situation

F

The STRESS Ellipse

--- = Igneous Intrusions
gneisses from Ostersøy; some of the layers are tightly folded, yet the layers immediately above and below are less intensely deformed. This observation accords with theoretical and scale model concepts for folding (Ramberg, 1959, 1963; Biot, 1961, 1964; Donath, 1962; Chapple, 1968; Cobbold et al., 1971; Cosgrove, 1976), which predict that the strain round a folded layer dies out within approximately one fold wavelength from the layer in flexural slip buckling deformation. This concentration of strain round the "site of tectonic activity" is also postulated on the larger scale for continental plate boundaries (Bott & Dean, 1973). This would mean that intense deformation in the orogen is not likely to extend very far into the basement and it is therefore to be expected that the intensity of deformation will decrease sharply with distance from a "site of crustal activity". We might then expect to find that the deformation in a "basement" to such a "site" would increase sharply towards the paleo-surface (contact with the overlying folded rocks) and towards the "site of activity" horizontally (fig. 38). Such a model could be used to explain the observed sequence in tectonic style and intensity away from the Berghsålen front (figs. 18, 20 & 23).

A plot of the entire foliations measured on the Northern gneisses shows that the dominant structure is that of D2 (fig. 39, cf. fig. 24). The large (yet statistically insignificant) scatter is caused by the D4 cross folds. Figure 40 is a computer generated block diagram of the foliation surface, generated from the same data as that used in figure 39. It shows the major D2 antiformal structure with the effects of D4 suppressed, while figure 30 shows the effect of D4 with D2 structures suppressed.
PERSISTENCE OF DEFORMATION AWAY FROM DEFORMATION CENTRES

(Start Condition)

Cover

Basement

(Ramberg 1959 Model)

Basement Deformation Dies Out Within One Fold Wavelength From The Folded Cover

(Bott & Dean 1973 model)

"Site of Crustal Activity"

Untoformd "Foreland"
Fig 39.

PLOT OF ALL THE FOLIATIONS MEASURED IN THE NORTHERN GNEISSES

THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES
00.9c1-2%>0.9c3-4%>0.9c5-6%>0.9c7-8%>0.9c9-10%>0.9c11-12%>0.9c13-14%>0.9c15-16%>0.9c17-18%>0.9c19-20%>0.9c21-22%>0.9c23-24%>0.9c25-26%>0.9c27-28%>0.9c29-30%>0.9c31-32%>0.9c33-34%>0.9c35-36%>0.9c37-38%>0.9c39-40%>0.9c41-42%>0.9c43-44%>0.9c45-46%>0.9c47-48%>0.9c49-50%>0.9c51%

NOTE A VALUE OF>0.5-0.9c,LT.12%AND.GT.0%
2.2 STRUCTURAL HISTORY OF THE SOUTHERN GNEISS UNIT

The principal difference between the rocks comprising the Southern and Northern units is not one of petrology, but deformation. The early deformation in the Southern Gneiss Unit, unlike the Northern gneisses, proceeds to the point where the rocks develop mylonitic characteristics.

The chronology of deformation presented in this chapter is specific to the Southern gneisses, and cannot be directly equated with that presented for the Northern gneisses. Correlation between the two units will be made in the next chapter, as part of the discussion of the nature of the contact between the two units.

The deformation sequence in the Southern Gneiss Unit is summarised in figure 95.

2.2.1 The Early Deformation and the Production of the Early Foliation, S"Early".

Like the rocks of the Northern Unit the earliest deformation feature seen in the Southern gneisses is a gneissose banding and accompanying layer parallel schistosity. For similar reasons to those given for the S "Early" of the Northern gneisses, this banding and schistosity are interpreted as being of tectonic rather than igneous or sedimentary origin. The schistosity is seen as a penetrative amphibole (now much retrogressed to biotite) alignment. This is much better developed than its correlative seen in the Northern gneisses, such that the foliation overprints all evidence for older deformations.

S "Early" is very difficult to recognise in the field, because of the strong development of later fabrics which overprint it. It is most clearly established in the nose regions of F1 folds, which are best seen
at grid reference 17.50,17.00 (pl. 23, 24 & 25). The gneissose foliation is cut by early pegmatites (pl. 24 & 25), but the exact relationship between the pegmatites and the foliation is obscure; they appear as sheets sub-parallel to the schistosity, yet contain the early schistosity themselves. This is interpreted as meaning that the pegmatites were intruded into well foliated rocks, which were subsequently further flattened, such that all the features in the rock rotated into the plane of flattening. This flattening caused the development of a new fabric in the pegmatites and enhanced that fabric already present in the host rock, thereby obscuring all evidence for prior deformations.

There is also some evidence for pegmatites in the Southern gneisses that postdate D "Early" yet predate D1. These crosscut the S "Early" foliation at a shallow angle (pl. 24) and do not contain the S "Early" schistosity.

Discussion

The flattening and textures seen in the Southern gneisses differ from those seen in the Northern gneisses. The gneissose banding is much finer; the leucocratic segregations are only millimeters thick rather than centimeters (cf. pls. 1 & 25). The schistosity, too, is more obvious, being present and well developed in all the rock types, rather than just the melanocratic layers. The rock seems to be generally in a much higher state of strain than for the "Early" deformation in the Northern gneisses. The presence of detectable "Early" pegmatites in the Southern gneisses also distinguishes them from the Northern rocks. It is therefore clear that correlation between S"Early" (Northern) and S "Early" (Southern) is not possible. Thus it is likely that a direct correlation of the ensuing Southern Gneiss Unit deformations with those affecting the Northern Gneiss Unit would be incorrect, despite the fact
that the rock types involved in the D "Early" deformation (discussed in section 3 below), are similar to those found in the Northern gneisses. They include metasediments, gabbros and amphibolites, set in a matrix of granitic and granodioritic gneiss.

2.2.2 D1, The Production of the First Mylonitic Textures, F1, S1 and L1

This deformation produced the first folds seen in the Southern gneisses and is the predominant deformation; the general trend of the fold axes is shown in figure 41. The deformation occurred at a metamorphic grade of almandine amphibolite facies (see petrological section), and imparted a strong mylonitic fabric to the rocks. This texture generally overprints all evidence for D "Early". A table of D1 deformational events is given below:

TABLE OF FEATURES PRODUCED IN D1

1) F1 folds
2) Rotated garnet fabrics
3) A strong "L" tectonite rodding, L1
4) An S1 schistosity and foliation
5) Mylonitization
6) Axial planar pegmatites
7) Further pegmatites emplaced at a high angle to S1
8) Gabbroic sheets intruded parallel to foliation.

In D1 the S "Early" foliation was folded by the F1 folds, and F1 folds are seen throughout the Southern Gneiss Unit. They are also seen, equally developed, in the majority of the rocks to the south-west in the Bergen Arcs themselves.
Fig. 41.

NORTH
OSTEROY

GENERAL TREND
OF
STRUCTURES.

--- F1 Fold Axes

0 2.5 5
Km
The folds are usually isoclinal and tend to be intrafolial (pl. 26), however, where the D1 strain is lower, more open examples of these folds are seen (pls. 23, 24 & 25).

In less flattened examples, the F1 folds are asymmetric and recumbent, frequently with sheared middle limbs (fig. 42). The sense of overturning of these folds is ambiguous, as it is obscured by later deformations (fig. 43), the fact that F1 folds are isoclinal and the limited nature of the exposures. After removal of the majority of the effects of the later deformation, by rotations on a stereonet, it is possible to determine that the F1 folds originally trended north-east - south-west, and had flat-lying axial planes, which must also have had a north-east - south-west strike. The original direction of plunge cannot however be determined, but it is likely that the plunge was very gentle.

During the metamorphism accompanying the folding, garnets grew in some rock types, which show rotation fabrics (fig. 44, pl. 56).

A strong mullion rodding is associated with the development of the F1 folds (pls. 27 & 29). It is much more strongly developed than similar structures seen in the Northern gneisses, and is accompanied by a strong prolate ellipsoidal shape fabric (pl. 28), with the longest axis parallel to the "b" direction of the F1 folds (pl. 29, fig. 45). This rodding, L1, is penetrative and developed on all scales. As with the F1 folds, the lineation is also developed in the majority of the pre-Holdhus (Faerseth et al., 1977) rocks of the Arcs and is traceable into them from the Southern Gneiss Unit. (Elongate rodded boulders are frequently used as fence posts in the area). The exact nature of L1 is complex; in part it is a true mullion structure (pl. 29); in part flattened fold closures (fig. 43), leucocratic segregations (pl. 28), have their longest axis lying in the L1 direction and it is also defined by the penetrative
Fig 42. GENERAL SHAPE OF F1 FOLDS

Fig 43. GENERAL APPEARANCE AND MAIN FEATURES OF F1 FOLDS

Detached F1 Fragments Dispersed Into S2 By The Flattening Accompanying D2
ROTATION FABRICS ASSOCIATED WITH THE F1 FOLDS

SNOWBALL GARNET

GARNET WITH ROTATIONAL INCLUSION TRAILS OF QUARTZ

Note - Garnets now corroded and enveloped by a new younger schistosity
Fig 45.

PLOT OF THE GEOMETRY OF THE F1 FOLD AT [12.50,18.70],
SHOWING L1 LYING IN TECTONIC Y' FOR THE FOLD

- Plunge of Fold
- L1
- Poles to S 'Early'
- Axial Plane
- Fold Profile Plane
- S 'Early'

*, X, Y, Z = Tectonic Axes for The Fold
alignment of inequant mineral grains (pl.27). All these structures serve to define a prolate finite strain ellipsoid and such strain geometry has developed throughout the body of the rock, irrespective of the relative position on the meso- and macroscopic deformation structures. A model for the development of this strain geometry is proposed in figure 27. The fact that this requires two phases of deformation is further indication that S "Early" is the product of deformation rather than a primary sedimentary or igneous layering.

A new schistosity, SI, developed axial planar to F1 folds (fig.45) and this, although penetrative, cuts but does not overprint S "Early" in the F1 nose regions (pl.24); in the limb regions S "Early" and SI are coplanar. The extent of SI development is variable; in certain rock types (pl.25), the S "Early" foliation is overprinted to a greater extent. The variation is governed primarily by rock type; in the thicker least ductile units (acidic layers more than 10cm thick) the SI schistosity is generally less well developed. This variation in the D1 strain, allows examination of the textural development accompanying D1. During D1 folding there was a marked grain size reduction through the Southern gneisses, which coincided with a thermal peak in metamorphic grade; hornblende and almandine garnet developed during (and for some time after) the deformation (fig.44). This imposition of high strains on what were already highly strained rocks, together with the metamorphism, caused the development of a new "tectonic" rock type which is characteristically fine to medium grained, has a marked colour banding (1mm to 30mm thick, pls.24, 25 & 26), lying parallel to the SI schistosity (fig.46), and a flinty appearance. The textures developed and the general appearance of the rocks, closely resembles those of the "Augen Schists" from Eribol, Scotland (Lapworth,1885); the rocks are mylonitic.
The mylonitisation affects all the Southern Gneiss Unit, except some rare exposures of F1 fold noses, where pre-existant textures and foliations are seen. The process accompanied the formation of the F1 isoclines and the L1 rodding and was not confined to localised zones. (The textures and structures developed as a result of this mylonitisation, and an explanation as to the processes involved are given in section 4.)

The F1 folds were intruded by two suites of pegmatites. The earliest of these (emplaced during or soon after the F1 folding), are parallel to the axial planes of the F1 folds (pl. 25). They are granitic in composition and uncommon.

Subsequently the Southern gneisses were veined by a suite of much thicker (up to 1m thick) granitic pegmatites that cross-cut the new S1 foliation at a high angle (pl. 23). Some portions of the Southern gneisses are very rich in these pegmatites (10.80, 26.60), and can be mapped on this basis; in other parts, however, there are no pegmatites.

The final phase of igneous activity in D1 postdated the majority of the deformation, and comprises a suite of gabbroic sills emplaced parallel to the S1 foliation, the largest of which is seen at (05.20, 32.70). This particular body has a continuous outcrop for several kilometers along strike. There are also numerous smaller bodies associated with this main intrusion, which cut across F1 folds.

These bodies are confined to the south-west margin of the Southern gneiss outcrop. They do not contain L1, and are not cut by the late D1 pegmatites; this distinguishes them from the older D "Early" amphibolites and gabbros, which are cut by these pegmatites.

Discussion

The D1 deformation is seen to affect all the pre-pegmatite rocks
comprising the Southern Gneiss Unit. These include, in the south-west region, quartz and mica schists, which were of sedimentary origin, and a suite of granodioritic granites (which developed into augen gneisses during D1, and therefore contain Li). They are also cut by post-intrusional but pre-deformational quartz veins which became isoclinally folded in D1. An age of 794.08 ± 18.82 Ma. (see appendix 1) has been obtained on these granites. The granites contain xenoliths of the sediments, e.g. at grid ref. (06.26, 32, 30), and must therefore postdate them. The F1 folds are the earliest structures seen in the sediments and granites. S "Early" is not present in the granites; their foliation is entirely attributable to D1 and subsequent deformations. The identification of the structural sequence in the sediments is however difficult because the exposures are very limited and there is no immediate evidence as to whether the foliation defining F1 in them is of sedimentary or tectonic origin.

If it is tectonic, this implies that the sediments were complexly folded with the underlying gneisses prior to the granite intrusion. This would necessitate the introduction of a new fold phase for which there is no other evidence; there is petrological evidence that these sediments cannot be correlated with those seen in the Northern Gneiss Unit (see section 3.3.1) and the sediments may be traced in outcrop into the sediments of the Major Bergen Arc to the south-east of Osterfj. The foliation is therefore regarded as bedding.

However, this interpretation raises problems; xenoliths of the sediments are found in the granite, yet the outcrops of the two are everywhere separated by at least 200 meters of gneisses. This problem is resolved by assuming that the relationship between the sediments and the granites has been modified by secondary imbrication of the succession
during the formation of the Bergen Arc Thrust (fig. 10, section 2.3.2) either in late D1 or D2.

On this evidence, therefore, it is possible to propose a sequence of events preceding D1 and postdating D "Early". This was a period of deposition of a sedimentary cover complex, followed at the start of D1 by a phase of granite intrusion. The granites were sill-like intrusions, sub-parallel to the developing S1 schistosity.

The presence of the mica and quartz schists within the gneiss terrain raises the problem of whether they are autochthonous or allochthonous. If allochthonous, they must have been emplaced prior to D1. There is however no evidence for such an additional phase of deformation. The main translation event emplacing the Bergen Arc and Bergsdalen Nappes is interpreted as occurring at the end of D1, with perhaps a reactivation in D2 as discussed above. Therefore, as has been suggested, the sediments in the gneiss terrain are correlated with those in the nappe units; they are para-autochthonous or autochthonous. This means that as with the time gap between D1 and D2 in the Northern gneisses, the D "Early" deformation was followed by a period of erosion and deposition. The time gap between D "Early" and D1 is then probably of major importance.

The most significant feature of the D1 deformation is the mylonitisation, which imparts a "false simplicity" to the Unit. The mylonitic fabric is used as a "marker" in the analysis of later deformations.

The D1 strain history started with the production of F1 folds. The L1 lineation developed parallel to the Y axis of the strain ellipsoid as defined by F1 and the new S1 schistosity developed in the X/Y plane. Compression continued normal to the X/Y plane and the F1 folds tightened up, becoming isoclinal. In this process the mylonites developed.
At some stage during the compression, a suite of pegmatites and gabbros was emplaced into the gneisses, parallel to the S1 schistosity, while older pegmatites belonging to D "Early", were boudinaged in the plane of the new schistosity. Finally a new suite of pegmatites was emplaced across S1. These remain undeformed in D1 and therefore mark the close of this deformation.

The most important structural element developed in Bl is the Li fabric, this is very marked and persists through later deformations. (It is correlated with L3 of the Northern gneisses, see correlation section 2.3.1 below).

The sequence of deformation - folding followed by flattening with igneous activity - is similar to that seen in the Northern gneisses, but the intensity of the D1 flattening is far greater than in the Northern Unit.

The features developed in D1 are summarised in figure 46.

2.2.3 D2, The Production of the Second Mylonitic Tectures, F2 and S2

The structures produced during D2 are as follows:

1) F2 folds
2) S2 schistosity
3) Mylonitic fabrics
4) Some pegmatite emplacement.

It is noteworthy that the S1 foliation is not folded by F2; only the late cross cutting D1 pegmatites are affected (pl.23, fig.47). On the basis of these immediate observations, it could be argued that D2 does not exist. However, the possibility that the folding seen in the pegmatites is attributable to the D1 flattening event rather than F2,
SIMPLIFIED GEOMETRY OF THE STRUCTURES PRODUCED IN D1

Fig 46.

STRUCTURES ASSOCIATED WITH THE F1 FOLDS

L1 Mullions

L1 Tectonite Rodding L1

L1 Inter - section Lineation

Axial Plane

Axial Plane

Pegmatite

S1 Cleavage

Deformed Porphyroclasts in the Mylonite

S1 Mylonite

Boudin

0 M 0.5
Fig 47.

F2 AFFECTING THE LATE D1 PEGMATITES

Late D1 Pegmatite

S'Early

F1

F2

D1 Shear in a Middle
Limb of an F1 Fold

S1/S2 Composite

0  M  1
and that the D1 pegmatites were intruded earlier than indicated above, is disproved since a new schistosity, S2, accompanies this folding and crosscuts S1 at a shallow angle. This is fundamentally different from the relationships between the schistosities produced in the "double flattenings" (stage F (fig. 37)) in the Northern gneisses; these are parallel. While this difference could be explained by non-coaxial strain late in D1, F2 folds are observed within the Arcs (Faerseth et al., 1977). D2 and F2 in the basement must therefore be considered as a separate deformation.

The D2 deformation is heterogeneous on two counts (fig. 45); its over-all intensity increases from north-east to south-west across the Southern Gneiss Unit outcrop, and locally it is frequently confined to narrow bands of intense shearing (pl. 23). These shear zones are usually localised on certain portions of the earlier D1 structures; they frequently (but not always, they can cut "fresh" rock) form on the middle limbs of the F1 folds and where the D1 mylonite development is strongest. The intensity of the D2 fabric also tends to increase towards contacts between different units in the gneisses. The contact between the augen granite and the gneisses at (18.80, 09.50), provides a good example of this and plates 42 to 45 show the textural changes moving from the centre of the granite to the contact. In general, the D1 and D2 deformations seem to follow each other, where D1-S1 is strongest, then D2-S2 is strongest. For example, the areas in which the unflattened F1 folds are seen are also less affected by D2. In the old amphibolites, which are cut by the younger pegmatites that were folded in D1, the effect of D2 was to disrupt the folding, producing a chaotic structure in which the fragments of the folds are rotated in all directions (pl. 30).

The spatial relationship between S1 and S2 is not immediately apparent, on superficial examination the rocks do not appear to have two schistosities.
Fig. 48.

NORTH OSTEROBY

D2

STRENGTH OF

DEVELOPMENT

(Based on the dominance of the D1 and D2 schistosities)
This is because S2 is both variably developed and, where present, lies at only 2-5° oblique to S1. The schistosity is seen in isolation only in the post-D1-flattening pegmatites, and once its presence is suspected, structures attributable to it become apparent. These include:

A) A spindle-like shape to quartz grains in the gneisses

B) The boudinage and shearing of the L1 fabric-bearing rocks to form porphyroclastic remnants in a new S2 mylonitic banding (e.g. at 07°.90, 29°.80)

C) Boudinée of leucocratic layers.

The D1 fabric is dominantly an (l) tectonic fabric. Where this is intersected by S2 which, from the shape of porphyroclasts in areas of strong S2, appears to relate to an oblate ellipsoid, the L1 fabric is chopped up as shown in figure 49, producing spindle shaped relics. The individual rods are also separated laterally and wrapped round by S2 (fig.49), so instead of remaining as true prolate ellipsoids, they become flattened in the plane of S2. Thus the bulk of the gneisses have a good foliation on which the rock will split into flat sheets, yet on any one foliation surface there is exposed a strong linear fabric confined to that surface (pl.27). The immediate effect of D2 was to emphasise the D1 fabric, but as S2 gets stronger L1 is progressively destroyed. This effect is best seen in an along strike traverse through the middle of the Southern gneiss within one lithological unit. There is an increase in the intensity of D2 and consequent progressive destruction of L1 across strike to the south-west, and along strike to the north-west (fig.48).

During D2 there was extensive boudinage within the Southern gneisses. This not only affected the late D1 pegmatites (fig.50), but also involved leucocratic members of the gneissose banding when one of these was sandwiched between melocratic portions (fig.51). The late stage D1 boudins were also enhanced in this process.
DEFORMATION OF L1 IN S2

L1 Exposed on S2

Plane of S2

Quartz Spindle

Detailed Nature of L1 After Modification By D2

Shape of Cross-section of an L1 'Rod', in the S2 Schistosity

DETAIL OF A QUARTZ SPINDLE

Macroscopic Direction of L1

'Spindle' of Mono or Polycrystalline Quartz, the Remnant of a D1 Rodded Structure
TWO EXAMPLES OF D2 BOUDINAGE

**Fig 50.**

**Fig 51.**
100.

The amount of strain involved in the D2 boudinage, gauged from the spacing between "pips", increases moving up section towards the south-west. The D2 boudinage is also seen to extend some way into the Northern gneisses.

The visible foliation in the Southern gneisses is therefore a complex structure formed by the interplay of two separate foliations of slightly different character. As far as can be ascertained in view of the subsequent reorientation of S2 in D3 (2.2.4 below), throughout the Southern gneisses, S2 cross cuts S1 more steeply (fig.52a).

Accompanying the formation of S2, in the regions of strong D2 development, there was grain size reduction and the growth of a new minerals of upper greenschist facies (see 3.2, petrological section); the textures formed are mylonitic. The principal difference between the D1 and D2 mylonites is that the D2 generation is less penetrative and confined to zones which isolate patches where the D1 fabrics remain dominant. The D2 mylonites are best developed in a thick zone paralleling the junction with the Major Bergen Arc; here virtually all textures are attributable to it.

The D2 structures are cut by two sets of pegmatites, both very rare. The first of these, thin discordant sheets of fine grained grey trondhjemite, must have been emplaced during the folding because they contain a weak S2 and are only seen where they cut the folded D1 pegmatites. They are axial planar to the F2 folds (pl.23). The second set are coarse grained quartz, potash feldspar-rich pegmatites, only two examples of which were found. These trend across strike and are lenticular, being about 0.5m thick and 30m long. Their margins are diffuse; they seem to merge with the country gneisses and are interpreted as being of local metasomatic origin (veinites).
RELATIONSHIP BETWEEN S1 AND S2

a) 

DIAGRAM SHOWING HOW THE D2 STRUCTURES IN THE BASEMENT ARE THOUGHT TO RELATE TO THOSE IN THE ARCS

b) 

D2 Thrusts
(Allochthon)

F2 Folds

Sheared F2 Folds

Flattened, Reworked Basement
(Para-autochthon)

ARCS
Discussion

It is apparent that in D2 the Southern gneisses were, though not folded, highly strained. A new penetrative fabric was acquired and secondary mylonitic textures pervasively developed. The D2 deformation is therefore as important as D1.

D2 displays a similar heterogeneity to that seen in D2 and D3 of the Northern gneisses (see 2.1.4 above), which increases in intensity approaching the Bergsdalen Nappes, only here there is an increase towards the Bergen Arcs as well (fig. 48).

It can be proved (see 2.3.2 below), that the D2 flattening episode in the Southern gneisses must have been synchronous with a fold-producing deformation identified by Faerseth et al. (1977), within the Arcs. D2 is therefore interpreted as a basement response to deformation which at a higher level in the tectonic pile gave rise to folding. Because of the similarity in nature of the deformatinal effects produced in the Southern gneisses and the semi-parallelism of S1 and S2, the D2 deformations must have been very similar to those in D1. A model envisaged is given in figure 52b.

Determination of the D2 direction of tectonic transport is critical but presents a problem, because no major F2 folds are exposed in the Osterøy basement. There are four possibilities; like D1 the structures could have been aligned on a north-east (Bergsdalen) axis with transport from either the north-west or south east; or the structures could have had a north-west (Bergen Arc) axial trend, with transport from the north-east or south-west.

It is important in regional terms to decide at which stage in the deformation history of the area structures with a Bergen Arc trend first appear. The thinning and cutting out of lithologies moving along strike
in the Southern gneisses away from the Bergsdalen front (map 1), a D2 phenomenon, seems to indicate a component of transport along a north-west vector, though as in the Arcs (Faerseth et al., 1977) this is inconclusive because of the effects of later folding. However, the distribution of S2 (fig.48), which strengthens towards the Bergsdalen Nappes, suggests D2 should be included with the Bergsdalen affiliated deformations.

The D2 strain ellipsoid is probably very similar to that formed in D1 with some rotation, the X Y plane of the D2 ellipsoid is slightly oblique to that of D1. S2 is therefore a schistosity which formed in the bc plane of a major fold phase. The folds seen to affect the late D1 pegmatites are the expression of these folds in the basement, and formed during the development of D2; S2 is axial planar to these folds.

There are several important deductions that may be made based on observations of the D2 boudinage. Firstly, it is the leucocratic portions that form the boudins, and the melanocratic portions that form the "ductile" matrix. This is the reverse of the situation usually seen in gneiss terrains, where the basic portions form the boudins (as with D1 in the gabbros of the Northern gneisses). In the Northern gneisses the ductility contrasts controlling boudinage were the result of elevated temperatures (the leucocratic portion was of high ductility because it was near its melting point). While in the Southern gneisses the controls were mechanical, there is a grain size difference between the two units (the finer grained melanocratic rock was the weaker (fig.144) ) and slip on the cleavage planes of the mafic minerals (biotite and amphibole) would have taken place. The D2 deformation occurred at relatively lower temperatures than D1 (Northern Unit). This is borne out by the fact that during D2 the rocks were at a much lower grade of metamorphism than previously seen on Osterfy (see section 3.5).
The observation that this deformation crosses into the Northern gneisses indicates conclusively that the two units were adjacent at the time of the D2 deformation.

The dominant foliation mapped in the Southern Gneiss Unit is then a complex foliation, composed in some parts mainly of S2 while in others it is dominantly an S1 surface. Separation of the two is generally impossible in outcrop because of their near parallelism. This problem also affects the lithological units mapped; in map 1 these units are not strictly lithological divisions defined on a mineralogical basis, they are a reflection of both mineralogical and textural variation. This latter variation was produced during D1 and D2 (mainly D2 because of its heterogeneous nature). The mapped units are then strictly lithotectonic units, except for the igneous members, and all contacts are tectonic.

The structures associated with D2 are summarised in figure 53.

2.2.4 \( \text{D4, The Production of F}_3, \text{S}_3, \text{L}_2 \text{ and L}_3 \)

The \( F_3 \) folds are the most obvious folds seen in the Southern gneisses, and within the Bergen Arcs generally. Their development will be considered in detail. It is D3 that is correlated with \( D_4 \) of the Northern gneisses, for reasons which will become apparent during this description and which are discussed in the correlation section (2.3.1 below).

The structures produced in D3 are as follows:

1) \( F_3 \) folds, folding the \( S_1/S_2 \) composite foliation
2) \( S_3 \) schistosity
3) \( L_2 \), mineral growth lineation
4) Folding of \( L_1 \)
5) An \( L_3 \), \( S_1/S_2 \), \( S_3 \) intersection lineation
6) \( D_3 \) quartz veins, and hydraulic fracturing.
SIMPLIFIED GEOMETRY OF THE STRUCTURES PRODUCED IN D2

STRUCTURES ASSOCIATED WITH THE F2 FOLDS

Only S2 in the Vein

Late D1 Granitic Vein

Weak S Early

Strong S2 in a D2 Mylonite Band

S1/S2 Composite

F2

F1
The F3 folds fold the Sl/S2 composite foliation and refold F1 and F2 folds (pl.31 and figs.43 & 54). The folds are asymmetrical, non-planar, non-cylindrical, recumbent folds overturned to the south-west (towards the Bergen Arcs). The wavelength of these folds is, on the largest scale, about 10 Km. The fold style is summarised in figure 55. The whole attitude of the Southern Gneiss Unit and a large portion of the adjacent Arc units is governed by one such fold, the short steep limb occupying the region of the large lake in the centre of Osterøy (fig.56). On the smallest scale in thin section, these folds are seen to affect the Sl/S2 biotite schistosity (fig.57). Between these extremes the F3 folds are pervasively developed throughout the Southern Gneiss Unit and have been traced across south Osterøy over as far as Bergen, where they have been identified in the Ulriken Gneiss Complex (Thon, pers.comm. and Faerseth et al., 1977). They are also equally well developed in the south of the Arcs at Os and in the north on Lindas. In short they form the most obvious folds in the whole Bergen area. Plates 32 and 33 and figure 58 show examples of these folds.

Figure 59 shows the variation in plunge of F3 folds. This variation is the product of a primary non-cylindricity and is clearly not the result of cross folding. The pitch of the fold axis, measured in the axial plane, may vary by up to 90° over short (0.3m) lengths of hinge, giving rise to concentric outcrop patterns, similar to a Type 1 interference pattern (Ramsay, 1967), (pl.34, fig.60). Furthermore the axial planes of the F3 folds are primarily non-planar (fig.61), and these are subsequently folded by D4 (see below). Thus while the over-all geometry of the F3 folds defines a recumbent structure (fig.55), in individual outcrops the folds may be upright. Figure 62 shows how this affects the plunge direction and therefore the variation present in figure 59.
Fig 54.

F1 FOLDS REFOLDED BY F3

D3 Slide

F1 Fold Closures

5 cm. in Diameter
Fig 55.

DIAGRAM SHOWING THE GENERAL STYLE OF THE F3 FOLDS
Fig. 56
NORTH OSTERBY

D3
S2 Defining the Major F3 Fold

[shallow]
[steep]

0 2.5 5
Km
Fig 57.

**SMALL SCALE F3 CRENULATIONS AFFECTING AN OLDER (S2) Biotite Schistosity**

![Diagram of small scale F3 crenulations affecting an older (S2) biotite schistosity.]

Fig 58.

**SMALL F3 FOLD WITH SHEARED MIDDLE LIMB**

![Diagram of a small F3 fold with sheared middle limb.]

Old S1 & S2 Foliations

The Middle is Thinned and Then Translation Occurs on it

New S3 Schistosity Develops in the Shear, but is not Present Outside it
Fig 59.

PLOT OF PLUNGE OF THE F3 FOLDS

THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES

0.0, 9 = 0.2, 9 = 0.4, 9 = 5-6%, 9 = 7-8%, 9 = 9-10%, 9 = 11-12%, 9 = 12.9
12.9 = 13-14%, 9 = 15-16%, 9 = 17-18%, 9 = 19-20%, 9 = 21-22%, 9
22.9 = 23-24%, 9 = 25-26%, 9 = 27-28%, 9 = 29-30%, 9 = 31-32%, 9
32.9 = 33-34%, 9 = 35-36%, 9 = 37-38%, 9 = 39-40%, 9 = 41-42%, 9
42.9 = 43-44%, 9 = 45-46%, 9 = 47-48%, 9 = 49-50%, 9 = 51%+

NOTE A VALUE OF > 0.5-G, 9 <= L.T. 1% AND GT. 0%
OBSERVED NON-CYLINDRICITY OF F3 FOLDS

SIDE VIEW OF AN F3 FOLD HINGE

240°
Non Plane Non-Cylindrical Folds
With Cylindrical, Curvilinear Axial Surfaces

Axial Surface

Non Plane Non-Cylindrical Folds
With Non-Cylindrical Axial Surfaces

Axial Surface

Persistence of the Folds in Profile,
Along the Axial Surface.
Fig 62.

EXPLANATION FOR THE VARIATION IN F3 PLUNGE SEEN IN

Fig 59

Plunges on a Non-Cylindrical Fold With The Axial Plane Horizontal

Plunges on a Non-Cylindrical Fold With The Axial Plane Vertical

Axial Plane

Axial Plane
The mean plunge of F3 is 135/10° (defined by the longest portion of the more cylindrical members) and there is a slight (10°) rotation of this across Osterfy accompanying the "swing" of the Arcs. As the folds become more non-cylindrical the axial curvature becomes asymmetric, with the "culminations" pointing to 240 or 060° (fig. 60). This situation is similar to that found on Sørfj (Ramsay & Sturt, 1973), but in contrast the "synformal" culminations are no tighter than the "antiformal".

The non-cylindricity of F3 folds varies over the Southern gneiss outcrop and is controlled by two factors:

1) The position of the fold in question relative to the fold on which it is parasitic.

2) The "competence contrast" between the folded layers involved and their absolute and relative thicknesses.

The first of these controls is well known and described by Ramsay and Sturt (1973). They measured the angle between the axial plane and the mean attitude of the layering and plotted this against the axial deviation (fig. 63). It is impossible to produce such a plot for F3, because, as described above, the axial plane does not provide a constant reference. Also the Osterfy gneisses do not provide adequate exposures of complete surfaces, in a limited area needed for such a diagram to be reliably constructed. It is possible, however, knowing the morphology and position of the largest scale F3 fold which affects the whole Southern gneiss outcrop, to comment objectively on the observed axial deviation in outcrop-sized minor folds in various situations on the larger structure. On the gently dipping long limb, these minor folds are markedly non-cylindrical; an example is seen in the road section at (09.50,34.60), this is typical of the style of F3 along strike. The nose of the main fold crops out in the lake at (11.50,24.00); on its north shores the
NON-CYLINDRICITY AND THE ANGLE BETWEEN THE AXIAL PLANE AND THE MEAN ATTITUDE OF THE LAYERING

AP = Axial Plane
S = Mean Layering
foliation is steeply dipping and the minor folds show less axial deviation, the deviation also occurs over much longer lengths of axis. The F3 folds conform then to the pattern of non-cylindricity described by Ramsay and Sturt, despite being generally more complex in development than the Sχρψy examples.

A second factor is that under the prevailing metamorphic conditions (lower greenschist facies), it was the melanocratic portions of the multilayers that were the most ductile (p.103, figs.50 & 51). Where a sequence comprising a leucocratic layer set in a melanocratic matrix was involved in F3 folding, the folds developed show extreme axial deviation regardless of position on the mother fold (pl.32). The effect becomes less marked as the relative thickness of the leucocratic layer increases and as the absolute thickness of the leucocratic layer increases while the relative thickness of the multilayer components remains constant.

Discussion on F3

The above observations provide an insight as to how non-cylindricity develops. Biot (1961 & 1964), Ramberg (1959) and Ramsay (1967), show that the wavelength of folds is dependent on the thickness of the buckling layer and the viscosity contrast between the folding layer and the matrix (virtually constant for the Southern gneisses in D3). Further, Cosgrove (1976), shows that in a multilayer in which individual layers are of varying thickness, the thinner layers fold in advance of the rest. It is therefore reasonable to assume that the extreme non-cylindricity seen in thin layers is controlled by the time of fold formation. This observation is borne out by the fact that changes in the relative and absolute thicknesses of the layers is accompanied by changes in geometry.
The obvious explanation as to why early formed folds are the most non-cylindrical is that they have been more flattened as the deformation progressed. A high degree of flattening would also be true for the folds in the major-fold limbs (fig.63).

It follows from the non-cylindricity of folds observed in the field that it should be possible to observe a change in fold style as one moves from place to place along the axis. Unfortunately, it was not possible to do this on a single fold in the field or the laboratory, but many profile view photographs were taken of the F3 folds, then analysed by dipisogons and t'\(^2\) \(\theta\) plots (Ramsay, 1967). Some of these analyses are presented (figs.64, 65, 66 & 67). From these it is seen that the basic style of the F3 folds is class 1C, with class 1A and 1B elements appearing in some layers. This probably means that when first formed the folds were class 1A or 1B and the 1C geometry developed subsequently as they flattened. Where the middle limb of the fold begins to shear, the geometry changes from class 1C to class 2 in the shearing limb. Where the non-cylindricity begins to develop, seen as a progressive thickening of the layers in the nose region, the modification of the 1C geometry is to class 3.

This describes the over-all trend, but as is seen from figures 66 and 67, the fold geometry tends to become complex because the folding layers are not perfect and shear on the middle limbs occurs simultaneously. The situation seen in the F3 folds is summarised in figure 68, and it is important to note that any one fold dies out as one moves in any direction in the plane of the axial surface, away from the point of maximum axial deviation (fig.61). In the third dimension, the situation is as shown in figure 69; in a profile through D where the folds are "dying out" a class 1A geometry is inevitable. It is therefore clear that the production
Fig 64.

PROFILE SECTION THROUGH AN F3 FOLD, WITH ISOGONS

Dip Isogons

[Fold Class Determined From Isogons and t'α Plots]

Melanocratic

Class 1C

Leucocratic

Class 1C-1B

Class 1C

Melanocratic
Fig 65.

PROFILE SECTION OF AN F3 FOLD WITH DIP ISOGONS

[Dip Isogons]

Class 1C - Leucocratic

Class 3 - Leucocratic

Class 1C - Melanocratic

Unclassifiable

Class 2 Limbs

Unclassifiable

Leucocratic

Melanocratic

[Fold Class Determined From Isogons and f₀ Plots]
PROFILE OF AN F3 FOLD SHOWING MODIFICATIONS CAUSED BY EXTREME CONTRAST IN COMPETENCE BETWEEN THE LAYERS
PROFILE OF AN F3 FOLD WITH DIP ISOGONS CONSTRUCTED

[ Classes Determined from the Isogons and t_d Plots ]

0 M 1
Fig 68.

**GEOMETRY OF A LAYER IN SUCCESSIVE PROFILE SECTIONS ACROSS A NON-CYLINDRICAL [F3] FOLD**

**SECTION**

- A
- B
- C

**CLASS**

- 1A / 1B
- 1C
- 3
DIAGRAM EXPLAINING WHY 1A GEOMETRIES ARE FOUND IN
THE F3 FOLDS

[Class 1A profiles must occur in places for the non-cylindrical folds to fit together across the layering]
of non-cylindrical folds induces the production of class 1A and class 3 geometries. A single layer traced out in three dimensions is folded in all classes (except class 2) in different parts, (the class 2 geometries being interpreted as the product of a different deformation process which acts after the folds initiate). The fold pattern produced interlocks, such that the total shortening across any one layer is the same in any section cut normal to the average profile, provided a number of whole wavelengths are involved. The layer when unfolded would be rectangular (fig. 70, pl. 35).

This style of development requires the movement of material, not only in a direction parallel to the axial plane, from the limbs to the crest, but also in the plane of the axial plane, parallel to the crest (from areas (a) to areas (b) fig. 70). We can therefore consider the fold geometry in a section parallel to the axial plane cut through the hinge. Figure 71 summarises the possible geometries. The thickening in the hinge region where the geometry parallel to the axial plane is class 1A/3, is accommodated more by movement of material from the limbs than along the crest. It thus accompanies the development of a class 3 profile geometry (fig. 72). This relationship is only developed if the deformation process is one of heterogeneous simple shear.

A lineation, L2 develops with the F3 folds. This has two components; on the layering surfaces it is defined by inequidimensional mica flakes or mica flakes included at differing attitudes to S1/S2 and therefore presenting elongate sections on the foliation planes. Within and on the more leucocratic layers, the lineation is further seen as elongate, spindleform grains of quartz or feldspar lying at an angle to the S1/S2 foliation surface. This may develop into a ridge and groove structure closely resembling slickenside lineation where the plane of the foliation
Fig 70.

**DIAGRAM OF A NON-CYLINDRICALLY FOLDED LAYER**

One 'Set' of Folds

Two 'Sets' of Folds

Showing the Inevitable Geometry of the Interlocking

NOTE - While the Deformation is Only Buckling the Layer Remains RECTANGULAR.

[ Fold Culminations Must Therefore Form an En-échelon Pattern ]
All Classes of Fold

Non-Cylindrical Folds Looked at in Axial Planar Section

Profile Geometry —— Corresponding Axial Geometries

Class 1A

Class 1B

Class 1C

Dip Isogons

Class 1

Class 2

Class 3

Note: 1B or (3/1A)
MOVEMENT OF MATERIAL DURING HETEROGENEOUS SIMPLE SHEAR

Movement Parallel to the Axial Plane

Transport of Material From the Limb Regions to the Nose Region
lies close to the long axis of the "spindles". This latter component is only strongly visible in thinner leucocratic layers showing extreme axial deviation.

As described by Ramsay and Sturt (1973), similar lineations can be interpreted as the product of layer parallel extension in the long axis of the sectional strain ellipse \( (x') \) for any particular foliation attitude.

The lineation then shows the direction of mass transport on any one plane, and when the geometry of the lineation is considered over-all it defines the direction of tectonic transport during the shearing \( (a) \) (fig. 73). The geometry of this lineation on a fold is described by Ramsay and Sturt (1973). In their examples the lineation is not present in regions of the fold where the foliation is at a high angle to the \( (X) \) direction of the strain ellipsoid, (the regions of axial culmination). Ramsay and Sturt suggest that the lineation developed in a constrictional regime \( (K>1) \). As is seen from figure 73, the \( L_2 \) lineation is present all round the \( F_3 \) folds and forms a hinge parallel lineation along the entire length of the hinge. This suggests that during the production of \( L_2 \) the deformation plane strain \( (K=1) \) was operative. The situation is, however, more complex than this since examples of \( F_3 \) folds were found at \( (06.70,31.00) \) where \( L_2 \) was refolded round the nose of minor \( F_3 \) folds lying on the long limbs of larger \( F_3 \) folds (fig. 74). This implies that the development of \( F_3 \) structures was long lived and multi-staged, once a fold was formed by flexural slip/tangential longitudinal strain (Ramsay, 1967), the deformation switched to a process of heterogeneous simple shear, while on the long limbs new folds were initiated. Similar observations have been made on the folds from Sprey (Ramsay & Sturt, 1973).

The development of non-cylindrical folds has also been described by Knill and Knill (1958), Voll (1960), Brynhi (1962) and Ramsay (1962, 1967).
Fig 73.

RELATIONSHIP OF L2 TO THE F3 FOLDS

Non-cylindrical F3

Tectonic 'a' Direction

L2

L3 Turns Parallel to The Hinge at the Crest of the Fold

F3

S1/S2

Weak S3 in Nose Region

0 M 0.3
Fig 74.

**DIAGRAMS SHOWING HOW L2 IS FOLDED BY SOME LATE DEVELOPING F3 PARASITIC FOLD**

**Simplified Geometry of Modified L2**

*Note - Plunge of F3 Removed*
There are four explanations as to their formation, summarised by Voll (1960):

1) Unequal stretching along the strike of the axial surface (internal rotation).

2) Passive rotation of detached hinges between discrete schistosity surfaces (external rotation).

3) Buckling of the layer prior to the development of the schistosity, which thus produces a curvilinear intersection with the layering.

4) Layering planes intersected by schistosity with different attitudes.

Ramsay and Sturt (1973) suggest that the range of complexity of structure observed can only be explained by non-uniform flow parallel to the axial surface between adjacent folds; they suggest that the non-uniform flow is the result of flattening ($K < 1$) superimposed on folds initially formed in the constrictional field ($K > 1$). Rhodes and Gayer (1977), suggest they are produced by layer parallel shearing during nappé translation. Fold hinges develop in the Y direction and rotate into X as the result of large strains at points of high cohesion.

It is considered that methods (3) and (4) are special situations and should be disregarded as not pertinent to the observed situation on Osterfj. Methods (1) and (2) are both possible but on the evidence from Osterfj, internal rotation (1) seems more likely because of the range of fold styles produced, the nature of L2 and the general geological context in which the folds form. The mechanism postulated, unlike that of Ramsay and Sturt, starts with a flexural flow tangential longitudinal strain deformation ($K < 1$ or $K = 1$), which, after the folds "locked up" changes to a process of
heterogeneous simple shear ($K=1$), the greatest axial deviation occurring in the regions of highest strain, the "less competent" melanocratic layers deforming ahead of those which are leucocratic.

The hypothesis is further supported by the behaviour of LI deformed round F3 folds. Figure 75 is a plot of LI across the Southern gneisses, the pattern seen is interpreted as in figures 76 and 77. Stage 1 shows the original geometry of LI (slightly disturbed by D2), Stage 2 shows LI folded by F3, where the folds have a 1A-1C geometry. At this point the deformation could no longer proceed by flexural slip (flow)/tangential longitudinal strain, because the folds "locked up". The deformation process then switched to one of heterogeneous simple shear and the folds flattened. The Y axis at this stage had shifted so that the shear was oblique to the original fold axis (thus explaining the asymmetry in axial curvature), Stage 3. This produced a final distribution of LI plunges as shown in Stage 4 (compare this with figs.75 & 78).

There is no evidence in the Southern gneisses to support Rhodes and Gayer's (1977) model of layer parallel shearing, indeed it is clear from the broad scale development of the major F3 folds that this model is untenable. The process must be fundamental to the deformation regime rather than a "special situation".

It is thus postulated that the non-cylindricity is the result of the mechanical state of the rock at the time of deformation and the mechanism by which it starts to deform. The rocks were at a metamorphic grade where there were marked mechanical (ductility) contrasts between layers of differing composition; the layering was very irregular and the rock locally extremely heterogeneous. The deformation started with folds initiating at points of heterogeneity, forming discrete buckles of the layering (pl.35), at numerous sites in the rock at one time. These folds
Fig 75.

PLOT OF L1 FROM THE SOUTHERN GNEISSES

THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES

0.9<1-2%:02,9<5-6%:04,9<5-6%:06,9<7-8%:08,9<9-10%:10,9<11-12%:12.9
12.9<13-14%:14,9<15-16%:16,9<17-18%:18,9<19-20%:20,9<21-22%:22.9
22.9<23-24%:24,9<25-26%:26,9<27-28%:28,9<29-30%:30,9<31-32%:32.9
32.9<33-34%:34,9<35-36%:36,9<37-38%:38,9<39-40%:40,9<41-42%:42.9
42.9<43-44%:44,9<45-46%:46,9<47-48%:48,9<49-50%:50,9<51%

Note a value of >0.5-0.9<LT.12AND.GT.0%
PROPOSED PATH FOR DEFORMATION OF L1, PART 1

STAGE 1

Original L1 Geometry

STAGE 2

Area Occupied by L1
After - Flexural Slip /
Tangential Longitudinal
Strain Deformation

F3 Steep Limbs, some Overturned

Shallow Limb

Axial Plane [AP]

TT Girdle of F3

(a)

(b)

a < b
PROPOSED PATH FOR DEFORMATION OF L1, PART 2

STAGE 3

Lines of Movement Traced by L1 During the Second Stage of Deformation by Simple Shear Processes

L1 as it was at the End of Stage 2

STAGE 4

THE SCATTER OF L1 WHICH WOULD RESULT FROM THESE PROCESSES [Compare this with the actual scatter plotted in Fig 75]
Fig 78.

SKETCH SHOWING HOW F3 WAS OBSERVED TO AFFECT L1 IN THE SOUTHERN GNEISSES

F3 Fold Culmination

L1

a > b > c

[This Geometry Supports the Deformation Path Proposed in Figs 76 & 77]
amplified and extended axially from their points of formation until they met another fold and interlocked. Because the folds did not initiate regularly, it is unlikely that every crest would merge with a crest (Dubey & Cobbold, 19??), and certain folds developed in advance of the rest to form an en-echelon pattern (pl.35). Smaller folds became parasitic on the larger folds and the buckling surface was first affected by a harmonic spectrum of developing folds with differing wavelengths and amplitudes, yet with constant wavelength/axial length/amplitude ratios. Thus preserving an identical "initial buckle" geometry throughout the layer. The layer, therefore remains rectangular at this stage (fig.70). Amplification continued until the folds "locked up"; this occurred first at the points of initiation as, at any one time these would be the most developed. The folds then had a primary non-cylindricity. At the point of "lock up" the deformation mechanism changed to one of heterogeneous simple shear, and the nose region extended even more, thus enhancing the primary non-cylindricity and causing the extreme axial deviations observed, L2, and the asymmetry along the fold axis. The lack of a well developed schistosity while there is a strong L2 lineation is taken to indicate that the shearing process occurs in a near plane strain regime. Because the process involves two mechanisms, there will be a relationship between "tightness" of the fold and the axial deviation (fig.63).

In the more micaceous lithologies, a weak new schistosity, S3, developed parallel to the axial plane of the F3 folds. It is marked by a crenulation cleavage and a parallel alignment of quartz deformation lamellae (pls.36 & 46). S3 forms a weak intersection lineation, L3, with S1/S2 in the noses of some F3 folds, which is parallel to L2. It is parallel to the average plunge of the F3 folds (fig.79).
Fig 79.

PLOT OF THE L3 INTERSECTION LINEATION

PLOT OF BERGEN ARC FOLD PLUNGE PARALLEL LINEATION

THIS PLOT HAS UNDERGONE -0- ROTATIONS
USING CORRINATE CARD 0

SAMPLING 2ACROSS/- 10DOWN
POLES PER 1PERCENT AREA 31 DATA EQ AR

THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES
00.9<=1-2%>02.9<=3-4%>04.9<=5-6%>06.9<=7-8%>08.9<=9-10%>10.9<=11-12%>12.9
12.9<=13-14%>14.9<=15-16%>16.9<=17-18%>18.9<=19-20%>20.9<=21-22%>22.9
22.9<=23-24%>24.9<=25-26%>26.9<=27-28%>28.9<=29-30%>30.9<=31-32%>32.9
32.9<=33-34%>34.9<=35-36%>36.9<=37-38%>38.9<=39-40%>40.9<=41-42%>42.9
42.9<=43-44%>44.9<=45-46%>46.9<=47-48%>48.9<=49-50%>50.9<=51%
NOTE A VALUE OF<0.5-0.9<=LT.12AND.GT.02
Also associated with the D3 deformation are two varieties of fracture and quartz veining. The first is associated with the F3 folds, the middle limbs of which develop shears (figs. 58 & 80). In some cases the shearing has proceeded to the point of rupture, where two adjacent fold noses are separated along a local thrust. In the thrust plane quartz and quartz-microcline veins up to 30 m long and 0.5 m wide frequently develop; they are usually at a low angle to the S1/S2 foliation and axial planar to the F3 folds. The quartz veins are also seen to intrude the axial planes of folds formed by the reclosure of the D2 boudins (in both the Southern and Northern gneisses) during the D3 deformation (figs. 81 & 82). The displacement across the veins varies from 0.2 m to 20 m. These veins are the same as those developed in D4 in the Northern gneisses (cf. section 2.1.6).

The second variety of fracture occurs late in D3, and is seen at (17.70, 11.50). Here there is a strike parallel band of rock bearing the S1/S2 foliation, folded by F3 and net veined with quartz and calcite. The veining is unaffected by F3 and the rock is completely shattered with rotated brecciated fragments set in a matrix of rock flour (pls. 47 & 50). There is no evidence of movement across the zone and it is interpreted as indicating the sight of rapid dewatering as the D3 stresses relaxed (pl. 37).

Discussion

F3 is responsible for the outcrop patterns seen in the Southern gneisses (map 1), the less resistant lithologies are eroded from the long limb regions back to the hinge region where the foliation is vertical. Thus, it is in the hinge regions that the variation in lithology is seen and because of the steep inclination of the foliation there is little expression of topographic effect on the outcrop patterns. The F3 hinges
Fig 80.

GENERAL F3 STRUCTURES VIEWED FROM THE SOUTHEAST

S1/S2

F3

D3 Thrusts or Shears

F3

S1/S2

S1/S2

F3 Shear with reverse transport

[From a Major Steep Limb (F3)]

Small Band of Mylonite

Quartz Veins
**Fig 81.**
RECLOSED D2 VEIN BOUDIN, — SITE OF AN F3 FOLD

**Fig 82.**
F3 FOLD NUCLEATED ON A D2 BOUDIN
follow the trend of the Bergen Arcs, but it must be stressed that the "arc" of the Bergen Arcs is not responsible for the non-cylindricity of the F3 folds, it is a younger structure that reorientates the F3 hinges (F4 see below).

The sense of translation deduced from overturning of F3 folds (fig. 80) is from the north-east to the south-west. This is the first clear indication of folding producing folds with north-easterly trending axial traces and translation at right angles to the general trend of the Arcs across Osterfjy, and probably represents the first phase in the formation of the Bergen Arc Structure.

On the basis of orientation and style of development this phase of deformation is correlated with D4 in the Northern gneisses; the contact between the two units is seen to be deformed by F3 folds. The difference in the effect of the deformation in the two units is that in the Northern gneisses the folds tend to have axial planes which are much steeper than in the Southern gneisses. In both units the structures developed progressively (figs. 34 & 74); this is interpreted as indicating that the deformation was long lived.

The development of the D3 structures is summarised in figure 83.

2.2.5 D4, The Production of F4 and S4

The D4 deformation is not readily apparent in the Southern Gneiss Unit; its presence is mainly inferred from the progressive reorientation of older structures. The features produced in this deformation are:

1) F4 folds
2) S4 fracture cleavage
3) Thrusts.
Fig 83.

SIMPLIFIED GEOMETRY OF D3 STRUCTURES

F3 Axial Plane, Quartz Veins & S3
Plunge of F3 & L3

Shallow Limb
Overturned Steep Limb

STRUCTURES ASSOCIATED WITH F3 FOLDS

Axial Plane
L2
L3
F3
S1/S2
Quartz Vein & Shear
Weak S3
Regionally, that is in the Southern Gneiss Unit, the Northern Gneiss Unit and the Bergen Arcs, D4 has a profound effect on all earlier structures though in any one area its effects are cryptic. A very large scale synformal fold structure is present and best appreciated by examining its regional effects. Plate 38, taken from (19.00,23.00) in the Southern gneisses, looking across the Arcs to the north, clearly shows this large scale synform.

The axis of the structure passes through south Osterøy (fig.84), and it is this structure which is responsible for the preservation of the tectonically higher units seen in the Arcs; they have been eroded from the corresponding antiforms on either side. The axial trace of the fold is arcuate but whether this is a primary feature or the product of some later deformation cannot be ascertained from evidence on Osterøy. It is this arcuate form which is responsible for the "arc" of the Arcs.

In the Southern gneisses the F4 folding is not directly observable because the S1/S2 foliation is too highly disturbed by the F3 folding. If, however, the attitude of the F3 axial planes are examined, F3 folds become apparent. The F3 axial planes, as described above, were not originally planar, but described a non-cylindrical surface. In part this folding of the axial surface is primary but a large portion is due to superimposition of D4. Figure 85 is a plot of the F3 axial planes measured across the Southern gneisses. There is obviously a large scatter, reflecting the primary non-cylindricity of the surface. Two clusters which lie on a great circle defined by smaller concentrations are however the product of D4 and represent the limb regions of F4.

The general southerly dip of the dominant foliation in the Southern gneisses (the long limbs of the F3 folds), towards the axis of the F4 synform in the south-west is appreciated from figure 86, a plot of the
Fig. 84
NORTH OSTERØY
D4
POSITION OF THE MAJOR F4 AXIAL TRACE

North Osterøy

Position of the major axial trace

Km

N

F4

0 2.5 5
Fig 85.

PLOT OF F3 AXIAL PLAINS WHICH DEFINES THE F4 FOLDS

THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES

00, 9 = 1-2% ≤ 0.9 = 3-4% ≤ 0.9 = 5-6% ≤ 0.9 = 7-8% ≤ 0.9 = 9-10% ≤ 0.9 = 11-12% ≤ 0.9
12, 9 = 13-14% ≤ 0.9 = 15-16% ≤ 0.9 = 17-18% ≤ 0.9 = 19-20% ≤ 0.9 = 21-22% ≤ 0.9
22, 9 = 23-24% ≤ 0.9 = 25-26% ≤ 0.9 = 27-28% ≤ 0.9 = 29-30% ≤ 0.9 = 31-32% ≤ 0.9
32, 9 = 33-34% ≤ 0.9 = 35-36% ≤ 0.9 = 37-38% ≤ 0.9 = 39-40% ≤ 0.9 = 41-42% ≤ 0.9
42, 9 = 43-44% ≤ 0.9 = 45-46% ≤ 0.9 = 47-48% ≤ 0.9 = 49-50% ≤ 0.9 = 51%

NOTE A VALUE OF 0.9 < 0.5 ≤ 0.9 < 0.6 ≤ 0.9 < 0.7 ≤ 0.9 < 0.8 ≤ 0.9 < 0.9 < 1.0 ≤ 0.9 < 0.0
PLOT OF S1/S2 FOLIATIONS FROM THE SOUTHERN GNISSES

This plot has undergone 0 rotations using coordinate card 0.

Sampling 2 across/1 down, poles per 1 percent area 17/99 data eq ar.

The contours represent the following percentages:

- 0.9% <= 1.2% <= 2.9% <= 3.4% <= 4.9% <= 5.6% <= 6.9% <= 7.8% <= 8.9% <= 9.10% <= 10.9% <= 11.12% <= 12.9%
- 12.9% <= 13.14% <= 14.9% <= 15.16% <= 16.9% <= 17.18% <= 18.9% <= 19.20% <= 20.9% <= 21.22% <= 22.9%
- 22.9% <= 23.24% <= 24.9% <= 25.26% <= 26.9% <= 27.28% <= 28.9% <= 29.30% <= 30.9% <= 31.32% <= 32.9%
- 32.9% <= 33.34% <= 34.9% <= 35.36% <= 36.9% <= 37.38% <= 38.9% <= 39.40% <= 40.9% <= 41.42% <= 42.9%
- 42.9% <= 43.44% <= 44.9% <= 45.46% <= 46.9% <= 47.48% <= 48.9% <= 49.50% <= 50.9% <= 51%

Note a value of >0.5-0.9 <= .LT. 1 AND .GT. 0%.
composite foliation (S1/S2) across the Southern gneisses (the weak cluster at the other end of the great circle reflects the steep limbs of F3. The information used in plotting figure 86 is also plotted as for figure 40 (fig. 87).

The great circle described by the axial planes of the F3 folds, figure 85, defines a plunge for the F4 structure at 140/15°.

At some localities within the Southern gneisses, (e.g. 05.50, 33.00) a vertical close-spaced fracture cleavage, S4, is seen to cross cut the F3 folds (fig. 88); it strikes at 140°. Despite the absence of any visible folds to which this cleavage might relate, consideration of its attitude suggests that it is axial planar to F4.

Late thrusts, which cut the F3 folds, are to be found within the Southern gneisses (18.00, 15.00). These thrusts or reverse faults form marked gullies in the Southern gneiss terrain, running parallel to the strike of the S1/S2 foliation (fig. 89). These thrusts persist into the Northern gneisses (14.00, 35.20). They are interpreted as having formed during the F4 folding and representing conjugate shear planes on the inner arc of the F4 synform (fig. 90). The thrust planes contain rock flour.

Discussion

The effects of the D4 deformation are not apparent in the Osterdy gneisses. Within the area generally the structure is best traced by observing the attitude of the long limbs of the F3 folds which F4 refolds. It is this effect which is visible in plate 38.

The D4 strain first developed with the generation of F4 folds. The deformation occurred under lowest greenschist facies metamorphic conditions (see section 3.5.4). Associated complementary shears in the limb regions were then developed on favourably orientated long limbs of F3 structures (fig. 90). A new weak fracture cleavage developed in association with
Fig 87.

MAJOR [F3] FOLDS AFFECTING THE SOUTHERN GNEISS

PLOT OF THE FOLIATION IN THE SOUTHERN GNEISS UNIT USING INCL. - DEPTH:

AZIMUTH = 45
ALTITUDE = 55
SLOPE H/UNIT = 1-00
WEIGHT = 0.50
* BEFORE FOreshortening 22/11/78
THE S4 FRACTURE CLEAVAGE CROSS-CUTTING F3 FOLDS

Fig 88.

Fig 89.

A. D4. THRUST

Scarp

F3 Fold Nose

Steep Thrust Plane

Power Pylon

Lake
Fig 90.

EFFECTS OF THE D4 THRUSTING

[The Thrusts Affect the Long Limbs of the F3 Folds]

D4 Thrusts

S1 / S2

F3

D4 Thrusts Develop in the Inner Arc

Median Surface
the F4 folds, orientated parallel to the a/b plane of the F4 folds, as established from figure 85.

These relationships are summarised in figure 91.

2.2.6 D5, The Production of Late Joints

The topographic effects of the joints on Osterøy are seen in plate 39. The joints are important economically as they form the valleys in which agriculture is located and also control the directions of the fjords (Nilsen, 1973). There is very little displacement associated with any of these features, though they are frequently zones of intensive brecciation, hydraulic fracture and mineralisation. The mineralisation is typically epidote, chlorite and calcite, which replaces and corrodes fragments of country rock found in the joint zones (pls. 48 & 49), which frequently have plane boundaries (pl. 40).

No detailed analysis of the joints on Osterøy has been attempted but a general analysis of lineaments on a macroscopic scale, based on air photographs and structures seen on map 1, is presented.

On Osterøy the joints seem to form three distinct sets (fig. 92):

a) Those which are radial to the Bergen Arc structure, orientated 020 to 050° with a concentration maxima at 045°.

b) Those normal to the Bergsdalen front, orientated 320 to 335°.

c) Those which parallel the fjord forming the east coast of Osterøy, orientated 355 to 005°.

All the joints are more or less vertical and the (a) and (b) sets have some small shear component (e.g. 10.00, 31.00 and 20.90, 33.70).

It is only in the Northern Gneiss Unit that all three sets are developed equally (fig. 92). To the south-west in the Southern Gneiss Unit and
Fig 91.

**SIMPLIFIED GEOMETRY OF D4 STRUCTURES**

![Diagram showing simplified geometry of D4 structures]

**EFFECTS OF F4 FOLDING**

![Diagram illustrating effects of F4 folding]

- Axial Plane of F4 and S4
- Limb
- Local Plunge of F4
- TT Girdle
- F4 Axial Plane
- North Osterøy
- Representative F3 Axial Plane
- F3 Long Limbs
- Define F4
Fig. 92

NORTH
OSTERØY

MAJOR JOINTS

320–335°
Joint

045° Joint

355–005°
Joint

0 2.5 5
Km
Bergen Arc, where the dominant foliation strikes $320^\circ$, the ($b$) set is not seen; and to the south-east in the Bergsdalen, where the dominant foliation strikes $045^\circ$, the ($a$) set is not seen. The absence of these two sets in these areas may be either because they are lost in the foliation or not developed there.

Figure 93 illustrates the distributions and this analysis of clustering for the joints on Osterøy is also applicable to adjacent regions and conforms well with the pattern found by Ramberg et al. (1977).

It is difficult to attempt meaningful analysis of jointing in gneissose basement, as in gneissose terrains with complex deformation histories it is thought that such analyses are unfruitful. At first sight this is the case on Osterøy; the joint sets are not clearly relatable to either the Bergen Arc or Bergsdalen structures, and do not fit patterns described for joints situated at orogenic fronts. For example, Muecke and Charlsworth (1966), who found that within the fold belt the joints form wrench sets and away from it become normal to the fold axes.

A sensible analysis of patterns seen on Osterøy can however be made by making two assumptions: firstly, that the joints are a phenomena of stress relief and dewatering as the rocks are uplifted by erosion and isostatic adjustment - dewatering is supported by the mineralisation and the stress relief hypothesis is well established (Price, 1966); secondly, that elastic strain remains stored in the rocks and the orientation of some of the joints probably reflects fossilised paleo-stress systems induced in the rocks during earlier deformations - the fact that elastic strain is stored in rocks is documented by Price (1966).

Examination of the structural development of the Osterøy gneisses has established that during their early history the maximum principal stress was probably horizontal, aligned at right angles to the Bergsdalen
ORIENTATION OF JOINTS

Joint Orientation Expected From Parallelogram Solution

(a) = 020-050° Joints
(b) = 320-335° Joints
(c) = 355-005° Joints

For the Bergen Arc Orogen

O°1 resolved for North Osterøy using the parallelogram of forces

THE PARALLELOGRAM OF FORCES

225°, O°1 for Bergen Arc

O°1, 135° for Bergsdalen

180° = Resolved O°1 (a change in magnitude of either vector will alter the position of resolved O°1, this will produce a scatter of joint orientations)
front (135°). Later, in D3 and D4, this switched to become orthogonal to the Bergen Arcs (locally 225°). The stresses stored as elastic strain in the gneisses must be the product of both of these, and assuming that one stress axis is vertical (gravity), the orientation of the maximum principal stress axis can be resolved using the parallelogram rule (fig. 93). The resolved maximum principal stress lies near 180°, given that the exact orientation depends on the relative magnitude of the two unresolved stresses at any one location; these will change the shape of the parallelogram, imparting a slight scatter to the resulting structures. Thus the expected orientations for shear joints would be 035 and 335° (normal to the Bergen Arcs and normal to the Bergsdalen (a) and (b) respectively).

A third set of joints (c) could develop along the acute bisectrix of conjugate set (a) and (b), parallel to the resolved maximum principal stress. This set could be interpreted as the result of progressive stress relief; after the initial formation of the shear joints, the maximum principal stress switched with the intermediate, and vertical dewatering joints developed in the 0l/02 plane (fig. 94). These joints would be orientated 000° and in the field these are the most mineralised, reflecting a high Pp.H₂O involvement.

Discussion

The interpretation presented here fits very well with the pattern seen on Osterøy; it is however at variance with that proposed on a regional basis by Hobbs (1911).

2.2.7 Discussion on Deformation of the Southern Gneiss Unit

Figure 95 is a summary table of the observed deformation features affecting in this unit. It is not possible to detect a pattern in the deformation as it was for the Northern gneisses. There are several possible explanations for this:
**RELATIONSHIP OF JOINTS TO RESOLVED STRESS FIELD**

**FIRST**—Shear Joint Formation

**SECOND**—The 355° Joint Formation

[the second joints are extensively mineralized with calcite & epidote]
**DEFORMATION IN THE SOUTHERN GNEISSES**

<table>
<thead>
<tr>
<th>D'Early'</th>
<th>D1</th>
<th>D2</th>
<th>D3</th>
<th>D4</th>
<th>D5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>F1</td>
<td>F2</td>
<td>F3</td>
<td>Thrusting</td>
<td>Jointing</td>
</tr>
<tr>
<td></td>
<td>S2</td>
<td>L2</td>
<td>S3</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Acid Igneous Activity And Pegmatites**

<table>
<thead>
<tr>
<th>Minor Sediments</th>
<th>Pegmatites</th>
<th>Qtz. Veining</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pegmatites</td>
<td>Flattening</td>
<td>L1 Thrusting</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>S'Early'</th>
<th>Basic Igneous Activity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
</tr>
<tr>
<td>Pegmatites</td>
<td>Pegmatites</td>
</tr>
</tbody>
</table>
a) The mechanisms operating in the Southern gneisses were fundamentally different.

b) Any pattern is obscured by the deformation as it proceeds.

c) Only part of a "cycle" is developed.

It is likely that all three factors have been operative. In D1 the deformation took place at the highest metamorphic grade and was superplastic (see section 4). Under such conditions folds would probably not develop serially, because layering would exert very little influence and thus the "cyclic" model postulated above could not operate. There is therefore no clear "flattening" (fig. 37(P)) as a second stage to the D1 deformation as was so typical of the Northern gneisses. The pegmatite relationships then become much more confused; D1 involves axial planar parallel pegmatites but these seem to have developed contemporaneously with the formation of S1. The amount of deformation which D1 represents is, for any given portion of rock, much higher than that seen in the Northern gneisses. It is postulated that D1 represents deformation at a "higher level" in the basement, a para-autochthonous nappe formed further in towards the centre of tectonic activity than anything seen in the Northern gneisses (fig. 38).

D2 on the other hand probably represents part of a "cycle". It is known, that at a higher level folding accompanied this deformation (see p. 173), and these folds fold the layering and are isoclinal (Faerseth et al., 1977). Such folds would be formed in the first stages of a "cycle". In the second flattening stage, the deformation would progress to deeper and deeper levels of the crust, forming a new foliation (S2). It is interesting to note that while the D1 mylonitisation is homogeneous throughout, the D2 mylonitisation is heterogeneous, weakening as it descends the tectonic pile, eventually to die out just across the contact between the Northern
and Southern gneisses. This suggests that these two units were "in place" prior to D2.

D3, on the evidence presented, is clearly the same deformation as D4 of the Northern gneisses. It occurs at a much higher crustal level, (structurally and metamorphically) than the preceding deformations, and clearly shows how at higher levels the layering determines the geometry of the developing structures. The trend of the D3 structures and the south-westerly transport is important as it is the first evidence seen in the basement for the development of the Bergen Arc structure.

The effect of the D3 folding across all the north Osterøy gneisses is seen on a computer generated block diagram of the foliation surface across them (fig. 96).

The D4 structures show a continued development of the Arc structure, folding the allochthon into a broad synform.

The question that thus arises is whether the Øygarden basement to the west of the Arcs is the same as that on Osterøy, and is the arcuate form of the synform primary or secondary? More work is needed on the western margin of the Arcs before these questions can be answered.

The D4 thrusts develop a different type of mylonite, more akin to fault gouge (Gibson, 1977), indicating that D4 operated at lower temperatures and thus at a higher level in the crust then the preceding deformations (see section 4).

D5 probably considerably postdates the earlier deformations, being the product of uplift and stress relief during erosion, the stresses responsible however seem to belong to the "earlier deformations", and as such it must be grouped with them.
COMBINED EFFECTS OF F4 (N. Gneisses) & F3 (S. Gneisses) ACROSS NORTH OSTEROY. (These Folds Belong to the Same Deformation)
The deformation sequence in the Southern gneisses is interpreted as representing tectonic events occurring near a "centre of orogenic activity" (the allochthonous rocks of the Arcs are in contact with the Southern Gneiss Unit and both contain similar structures) at progressively lower and lower metamorphic conditions. The Southern gneisses were, therefore, originally the same as the Northern gneisses, but have suffered higher (structural) level deformation related to the emplacement of the higher nappes, and must therefore be grouped with these nappes on a structural basis.

2.3 CORRELATION OF DEFORMATIONAL EVENTS

2.3.1 The Northern and Southern Gneiss Units

The deformation phases in the Northern and Southern Gneiss Units are summarised in figures 36 and 95 respectively. All the deformations cannot be directly equated; the geometry of the folds and structures accompanying D1 (Northern gneisses) and D1 (Southern gneisses) are different.

In the following discussion deformations referring to the Northern gneisses will be denoted Dn, those of the Southern gneisses Ds and the suggested correlated events Ds (e.g. D1n refers to D1 of the Northern gneisses).

In any discussion of correlation of the deformation events in the units, the nature of the contact between them, the Tysse Thrust (fig.10), is of vital importance. It could be argued that this contact is a major plane of transport and that the Southern gneisses are entirely allochthonous. In this case no correlation can be made between the two units prior to emplacement. In the field the thrust is cryptic; it is not at all obvious as a definite boundary (pl.41). Its presence only begins to be suspected when foliation orientations are examined (figs.39 & 86) and on subdivision
of individual lithological units (map 1). It separates two regions in which the dominant foliations strike at right angles to each other, and its position becomes very clear on a computer generated block diagram showing the variation of the northerly component of dip across the Osterøy gneisses (fig.99). Figure 99 and map 1 show that the 'strength' of the feature increases towards its western end. This represents an increased displacement on the thrust, in this region, with regard to the geometry of the structures it is juxtaposing. These features indicate that the Southern Gneiss Unit cannot be allochthonous or the displacement would be equal across the entire outcrop; it must therefore represent a para-autochthonous basement nappe derived by metamorphic and deformational reworking from the Northern Gneiss Unit, a view supported by examination of the petrology and metamorphic history of the two units.

To understand when the reworking and transport occurred, the deformations in the two units will be successively removed starting with the youngest structures to affect both units (D⁶), which is the late jointing. The point at which the deformational histories divide will evidently give the age of the Tysse Thrust.

The correlations between the two units are given in figure 97 and a discussion of the basis for these correlations follows:

Figure 97.

THE NORTHERN AND SOUTHERN GNEISS DEFORMATION CORRELATION

<table>
<thead>
<tr>
<th>NORTHERN GNEISS</th>
<th>SOUTHERN GNEISS</th>
<th>COMPOSITE</th>
</tr>
</thead>
<tbody>
<tr>
<td>D &quot;Early&quot;</td>
<td>Not seen</td>
<td>D &quot;Early&quot;</td>
</tr>
<tr>
<td>D¹</td>
<td>Not seen</td>
<td>D¹</td>
</tr>
<tr>
<td>D²</td>
<td>D &quot;Early&quot;</td>
<td>D²</td>
</tr>
<tr>
<td>D³</td>
<td>D¹</td>
<td>D³ Tysse Thrust</td>
</tr>
<tr>
<td>Not present</td>
<td>D²</td>
<td>D⁴</td>
</tr>
<tr>
<td>D⁴</td>
<td>D³</td>
<td>D⁵</td>
</tr>
<tr>
<td>Not seen</td>
<td>D⁴</td>
<td>D⁶</td>
</tr>
<tr>
<td>(D⁵)</td>
<td>D⁵</td>
<td>D⁷</td>
</tr>
</tbody>
</table>

PRECAMBRIAN

CALEDONIAN

AND/OR

LATER
Fig 99.

THE TYSSE THRUST IS APPARENT ON A DIAGRAM GENERATED ENHANCING THE NORHERLY COMPONENT OF DIP.
\( D^G \): \( F_4^S \) is not visible in the Northern gneisses because the strike of these rocks is orthogonal to the \( F_4^S \) axial direction. However, because there is evidence that the two units were in place prior to \( D^G \), \( F_4^S \) must in fact affect them.

\( D^G \): it has been proved (p. 143) that \( F_3^S \) and \( F_4^N \) are the same folds; this is the first point of obvious structural similarity between the two units (fig. 96).

\( D^S \): this correlation presents more of a problem. However, two points must be considered: a) \( S_2^S \) affects the Northern gneisses in a zone close to the Tysse Thrust. b) Moving up section in the Southern gneisses the effects of \( D_2^S \) increases and at the contact with the Bergen Arcs \( S_2^S \) is the dominant schistosity. It therefore finds a place between \( D_4^N \) and \( D_3^N \).

From this point back there is no immediate similarity in deformational events between the two units. The emplacement of the Southern Gneiss Unit must therefore be a \( D_3^G \) event and this therefore gives the relative age of the Tysse Thrust.

It is however proposed that \( D_1^S \) is the same event as \( D_3^N \). The bases for this proposition are seen in \( D_2^N \); development of \( F_2^N \) and \( S_3^B^N \) is heterogeneous, as shown in figure 18. If \( D_3^N \) were similar, then the whole Southern Gneiss Unit could be explained as a slice of highly re-worked basement from the root of the Bergsdalen Nappes (fig. 38), thrust over adjacent basement from the south-east, at the end of \( D^G \). \( F_3^N \) would then be the same as \( F_1^S \), \( S \) "Early" would be \( S_3^B^N \) and \( L_3^N \) would equate with \( L_1^S \). Such an origin for the Southern gneisses would also explain the north-easterly alignment of the \( F_1^S \) fold axes and the way the Southern Gneiss Unit thins slightly to the north-west.
The structures seen in the Southern Unit are therefore interpreted as more developed equivalents of those in the Northern Unit. Because the structures seen in the Northern Unit young south-eastwards (fig. 23), the Tysse Thrust brings together structures belonging to D3G and D4G in the west and D3G and D4G in the east, thereby explaining why its effects increase north-westwards. The thrust developed in two stages, first in D3G the Southern gneiss mylonites were formed. These mylonites represent "ductile thrusting" in the roots of an orogen (see section 4) and reflect the primary emplacement of the Southern Gneiss Unit. There is no sharp contact, rather the deformation is accommodated over a broad zone occupying the whole of the Southern Gneiss Unit.

Following this the D3G mylonites were reworked in D4G, a second thrusting event which, while still ductile is more brittle than D3G. For reasons given in section 4.2.3, the D3G mylonites absorb all the D4G strain so that little is transmitted to the underlying gneisses, explaining the absence of D4G in the Northern Gneiss Unit.

The Tysse Thrust is then a plane, separating rocks above which are strongly affected by two deformations that are only weakly represented in the rocks beneath. It is also a plane of transport (map 1) and is therefore best described as a plane of decollement, in some places "stuck" while in others "detached". D3G therefore is the point at which developmental histories of the two units converge.

2.3.2 The Osterfjord Gneisses and the Bergen Arcs

The deformation sequence postulated for the Arcs by Faerseth et al. (1977) (fig. 100), is based on observations of the rocks immediately to the west of the Southern Gneiss Unit, south-east of the area mapped. Extensive traverses were made in this area in an attempt to correlate the two. The following shortcomings in the sequence as presented by Faerseth et al.
SUMMARY OF STRUCTURAL, METAMORPHIC AND INTRUSIVE EVENTS IN THE BERGEN ARC NAPPEs, BASED ON OBSERVATIONS FROM THE SAMNANGER-OSTEROY AREA.

<table>
<thead>
<tr>
<th>DEFORMATION PHASE</th>
<th>STRUCTURES</th>
<th>METAMORPHISM</th>
<th>INTRUSIVES</th>
</tr>
</thead>
<tbody>
<tr>
<td>POST-HOLDHUS GROUP</td>
<td>Conjugate wrench faults. Dextral movements along N60°E faults. Sinistral movements along N120°E faults. Tight to open F3 folds. Mainly located in the eastern part of the Samnanger area, close to the Borgsdalen nappes. Close to tight major and minor F3 folds. Development of S3 crenulation cleavage. M4 Tight to isoclinal major and minor F2 folds. Development of S2 foliation. Thrusting and imbrication accompanied by extensive cataclasis. Emplacement of serpentinite bodies along thrust planes. Large-scale recumbent folding giving an S1 foliation.</td>
<td>Local thermal metamorphism</td>
<td>Intrusion of quartz diorite dykes Prograde in the Holdhus Group, varying from B1.2. in the south to B1.3. in the north. Retrograde in the Samnanger Complex B1.2.-B1.3. Intrusion of quartz diorites</td>
</tr>
<tr>
<td>PRE-HOLDHUS GROUP</td>
<td></td>
<td></td>
<td>Intrusion of ultramafic, mafic and quartz dioritic material</td>
</tr>
</tbody>
</table>

AFTER:--
Faerseh et al. 1977
were found. (References to Faerseth et al.'s sequence will be denoted $D^f$ while the proposed correlation with the Osterøy gneiss sequence will be denoted $D^g$ as before).

1) Their attempt to incorporate $D_5^g$ from the Tysse area with $D_4^f$ is rejected, as the style of the $F_5^g$ folds clearly does not agree with their description of $F_4^f$. It will be shown that the correct correlation is between $D_5^g$ and $D_3^f$.

2) There is no appreciation of the non-cylindricty of the $F_3^f$, these folds are highly non-cylindrical over the whole Osterøy and Samnanger areas.

3) Their interpretation of the structures developed in $D_4^f$ is questioned. Figures 9i and 10 of Faerseth et al. (1977) are supposed to show a dyke intruded along an $F_3^f$ axial plane then folded by $F_4^f$. In both cases the $F_3^f$ axial planes are demonstrably not folded. Emplacement of the dyke across the $S_2^f$ layering (in which $F_2^f$ isoclines can be seen, p.44, fig.10) prior to $D_3^f$ could equally explain these structures and would be in keeping with observations made in the gneisses, i.e. that the last stage of quartz diorite dyke intrusion was a late stage $D_2^f$ or $D_4^g$ event. Many of the other structures they attribute to $D_4^f$ could equally be attributed to $D_3^f$.

4) They interpret the "Osterøy Synform" (p.44, plate 1, AA, Faerseth et al.) as an $F_3^f$ fold refolding an $F_2^f$ fold. This interpretation is held to be incorrect partly because of the misinterpretations (1, 2 & 3) described above and for the following reasons:

a) Along strike from this synform, at Holdhus Fjell (plate 1, Faerseth et al.), similar structures are described as $F_2^f$ folds (p.40).

b) The type 3 (Ramsay, 1967) interference pattern (p.42 and plate 1, Faerseth et al.) could be caused by original sedimentary variation, as shown in
171.

Figure 101, and is therefore, on the basis of the evidence presented, inconclusive.

c) The structural position of the "Osterøy Synform" relative to the major D₅⁶ fold is shown in figure 102. It is therefore clear that the "Synform" is a D₄⁶ structure.

On the basis of the points made above it is possible to make the following correlations between the deformations seen in the Arcs and the Osterøy basement.

Figure 98.

**THE OSTERøy GNEISS AND BERGEN ARC DEFORMATION CORRELATION**

<table>
<thead>
<tr>
<th>NORTH OSTERøy GNEISSES</th>
<th>BERGEN ARCS</th>
<th>FINAL CORRELATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>D &quot;Early&quot;</td>
<td>Not present</td>
<td>D &quot;Early&quot;</td>
</tr>
<tr>
<td>D₁</td>
<td>Not present</td>
<td>D₁</td>
</tr>
<tr>
<td>D₂</td>
<td>Not present</td>
<td>D₂</td>
</tr>
<tr>
<td>D₃</td>
<td>D₁</td>
<td>D₃</td>
</tr>
<tr>
<td>D₄</td>
<td>D₂</td>
<td>D₄</td>
</tr>
<tr>
<td>D₅</td>
<td>D₃</td>
<td>D₅</td>
</tr>
<tr>
<td>D₆</td>
<td>D₄</td>
<td>D₆</td>
</tr>
<tr>
<td>Not present</td>
<td>D₅</td>
<td>D₇</td>
</tr>
<tr>
<td>D₇</td>
<td>(D₆)</td>
<td>D₈</td>
</tr>
</tbody>
</table>

Faerseth et al. report only one deformation common to all the pre-Holdhus rocks of the Arcs D₁. This deformation is also seen to affect the basement as it is possible to trace L₁ into the rocks of the Arcs, across the Bergen Arc Thrust (fig.10). Thus this thrust, like the Tysse Thrust, must be primarily a D₅⁶ structure, which means that the emplacement of the Bergen Arc Nappes is a D₅⁶ event. This D₅⁶ emplacement is supported by Faerseth et al., who describe D₁ as "inverting major parts of the succession before the development of the later D₂-D₅ structures" (p.37). Within the Arcs D₅⁶ will be imposed on rocks with various earlier histories because they are allochthonous nappes, and this is when the greater part of the tectonic transport occurred.
The "Osterøy Synform" (F2)?

The Position of the "Osterøy Synform" Relative to D5 of the Osterøy Gneisses.
$D_4^g$ thrusting and flattening on the basement is represented as the $D_2^f$ folding, thrusting and imbrication in the Arcs, (4c above). This is the deformation in which $S_2^g$ is imparted to the Osterøy Gneisses. As its effects increase up succession into the Arcs, and because the Bergen Arc Thrust is seen to cross-cut the lithologies of the Southern Gneiss Unit (map 1), it is suggested that the Bergen Arc Thrust is re-activated at this stage while the Tysse Thrust is not. This then is the second thrusting episode, in which the Holdhus rocks are now involved.

The evidence for the presence of $D_5^g$ is conclusive (4c above), and thus $D_5^g$ is correlated with $D_3^f$; the small scale structures of $D_5^g$ and $D_3^f$ are identical as suggested (2 above).

The Bergen Arc Synform is clearly a $D_6^g$ structure (pl.38). Some of the deformation structures described by Faerseth et al. as $D_4^f$ (p.45, fig.11 c & g) could be attributed to this structure.

The $D_5^f$ of Faerseth et al. comprises wrench faulting but such structures are not seen in the basement gneisses of Osterøy. The faulting probably occurred before the late jointing $D_7^g$.

Thus at least eight deformations are required to explain the structural development of the area as a whole and these are summarised in figure 103.
Fig 103.

**Deformation In The North Osterøy Gneisses**

<table>
<thead>
<tr>
<th>Jointing</th>
<th>Wrench</th>
<th>Faults</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>F6</td>
<td>S7</td>
</tr>
<tr>
<td>D6</td>
<td>F5</td>
<td>S6</td>
</tr>
<tr>
<td>D5</td>
<td>F5</td>
<td>S5</td>
</tr>
<tr>
<td>D4</td>
<td>F4</td>
<td>S4</td>
</tr>
<tr>
<td>D3</td>
<td>F3</td>
<td>S4</td>
</tr>
<tr>
<td>D2</td>
<td>F2</td>
<td>S3</td>
</tr>
<tr>
<td>D1</td>
<td>F1</td>
<td>S1</td>
</tr>
</tbody>
</table>

(Sediments)

<table>
<thead>
<tr>
<th>NOT Important</th>
<th>Sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>S2</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Features Developed</th>
</tr>
</thead>
<tbody>
<tr>
<td>Folding</td>
</tr>
<tr>
<td>Sigmoid</td>
</tr>
</tbody>
</table>

Deformation Features Developed
This section is divided into two parts; firstly, the description of lithologies as seen on map 1, based on field and thin section observations; secondly, a description of the metamorphic history of the rocks. The metamorphism is related to the structural elements as given in figure 103.

3.1. COMPOSITION OF THE OSTERØY GNEISSES

The area is divided into Northern and Southern Gneiss Units, A and B (fig.104), on the basis of tectonic history. The units are very similar in general composition, comprising gneissose acidic rocks with mappable minor metasediments and intrusions of both acid and basic plutonics. The minor lithological types provide an insight into the structural geometry of the region. The detailed variations in the background gneisses, however, were not mapped as they could not be traced over any distance at the scale of the map. The Southern Unit is subdivided in more detail than the Northern Unit and this is a reflection of tectonic condition. In the Southern Unit the gneisses form traceable bands distinguishable on the basis of outcrop morphology and colour, while remaining petrologically very similar in hand specimen or thin section. As the differences are more of grain size and texture rather than of chemistry, these units (map 1) must be regarded as litho/tectonic rather than lithological.

Quartz, albite, orthoclase plots (figs.105, 106 & 107), for the acid gneisses collected from north Osterøy (Units A & B), show that the rocks cannot be subdivided or grouped on a compositional basis. There is no clustering and the rock types identified in the field occupy overlapping areas centred, with a wide scatter of points, around the minimum melting eutectic composition for granite, at intermediate-high pressure (Winkler, 1974). The scatter could be caused by the presence of metasedimentary
## Lithological Diagram

### (A) Northern Gneiss Unit

**Flattened Gneiss and Migmatite Group**
(contains flattened representatives of all rock types present below)

**Gneiss and Migmatite Complex**

- **Sediments**
  - Quartzites
  - Mica Schists
  - Quartzose Migmatites

- **Migmatites**

- **Igneous Rocks**
  - Augen Gneiss
  - Amphibolites

**Unflattened Gneiss**

- **Pink**
- **Grey**
- **Intermediate**
- **Basic**

**Acid Veins**

### (B) Southern Gneiss Unit

**Metasedimentary Rocks**

- **Micaeous Members**
  - Mica Schist (1)
  - Mica Schist (2)
  - Mica Schist (3)
  - Phyllonites
  - Quartzites
  - Quartz Mylonites

**Igneous Rocks**

- Augen Gneiss
- **Amphibolites**
  - Type 1
  - Type 2
- Epidotite

**Gneissose Rocks**

- Heddalsfjell Gneiss
- Pegmatite Rich Gneiss
- Rispingen and Granitic Gneiss
- Drangskollen Gneiss
- Nonkletten Gneiss
- Undifferentiated Gneiss (varieties)

**Acid Veins**
(Modal % estimated, with point counting checking)

- Monkeitten granite
- Flatfaced biotite and migmatised
- Migmatised quartz-rich sediments

OSTERøy
SOME ACID GNEISE FROM NORTH
QUARTZ-ALBITE-ORTHoclASE PLAT FOR

Fig 105.
Some normative values from XRF analyses

Modal% estimated with point counted checking 2

Augen gneiss
Micaschists
Unmigmatized gneiss
Undifferentiated gneiss

Quartz-Abte-Albite-Orthoclase Plots

Ostervy

40 100%
Fig 107.
QUARTZ-ALBITE-ORTHoclase PLOT FOR SOME ACID GNEISSES FROM NORTH OSTERøy

Gneiss and Migmatite
△ Drengskollen Gneiss
● Pegmatite Rich Gneiss
■ Quartz Rich Gneiss
□ Granitic Gneiss

(Modal % estimated, with Point Counted checking)

100% AB

100% QTZ

OR 100%
material, deformation and K-metasomatism during migmatisation. The augen gneisses, however, define a tight cluster round the granite melt eutectic region, supporting field evidence which indicates an igneous origin for these rocks.

The similarity in composition of these gneisses means that litho-structural mapping is the only means of classifying most of the rocks in the field.

The basic rocks in the gneisses (Units A & B) are divided on structural bases into three chronological groups. The structural observations point to an intrusive origin. A major element AFM plot (fig.108), of some of the basic rocks from all units suggests that they are basaltic in composition. This together with their structural field relationships and the presence of relict igneous textures in a clump of pigeonite crystals from one of the bodies, confirms an intrusive igneous origin for the basic rocks.

Trace element analyses of specimens collected from each of the three groups are plotted, following Pearce and Cann (1973), (fig.109) and Floyd and Winchester (1975), (fig.110). The scatter observed in figure 110 indicates that both "Tholeitic" and "Andesitic" rock types are present, but detailed plotting shows that rocks belonging to any one 'time unit' fall into both fields. A similar picture is presented in figure 109, where the points scatter across all the fields. This could indicate that the chronological subdivision is wrong or could be explained by a) the intrusive episodes being complex involving more than one suite in each phase, or b) deformation and protracted metamorphism at amphibolite and greenschist facies conditions having affected the trace element distribution in the rocks to the point where they can no longer be used, as Pearce and Cann (1973) suggest.
Fig 10.9

Ti-Zr-Y plot of basic rocks from North Osteroy.

(After: Pearce & Cann, 1973)

(XRF Geochronological Analyses)
Fig 110.

$TiO_2-Zr/P_2O_5$ PLOT OF BASIC ROCKS FROM NORTH OSTERØY

(X.R.F. Geochemical Analyses)

[AFTE:- Floyd & Winchester, 1975]
These observations are similar to those made for the Moine amphibolite suite of central and northern Sutherland, Scotland (Moorhouse & Moorhouse, 1979).

3.2 **LITHOLOGIES OF THE NORTHERN GNEISS UNIT**

The rocks comprising the Northern Unit (A) are subdivided into three groups (fig.104), using both petrological and tectonic characteristics. Within each group there is a wide variation in rock type and petrology.

**TABLE GROUPING LITHOLOGIES OF THE NORTHERN GNEISS UNIT**

3) The Flattened Gneiss and Migmatite Group:

These are similar types of rocks to those of groups 1 and 2 below, but they are more deformed. This group is distinguished on structural grounds only, and is structurally the highest member.

2) The Gneiss and Migmatite Complex:

A suite of migmatised metasedimentary rocks, granites, granodiorites and amphibolites, tectonically involved with a suite of older (group 1) gneisses preserving structures of older deformations than those seen in the migmatised metasediments. This unit, which is structurally at an intermediate level is therefore distinguished on both a structural and compositional basis.

1) The Unflattened Gneisses:

These comprise a suite of granitic to granodioritic gneisses intruded by amphibolites. They form the lowest structural unit and are distinguished on a structural basis, in that they contain a fold phase not seen in any other rocks.

3.2.1 **The Unflattened Gneisses**

These gneisses outcrop on the west coast of the island, are mainly granitic-granodioritic in composition, with small amounts of basic rocks.
They are usually very well compositionally banded, with a banding parallel schistosity. The acidic members can, on the basis of their colour in outcrop, be further subdivided into three main types: Pink Gneisses, Intermediate Gneisses and Grey Gneisses.

The Pink Gneisses are granitic in composition (fig.111, and analyses 1, 3 & 33 in appendix 2), coarse-to medium-grained (1-2mm) and well foliated. Augen of porphyroblastic potash feldspar up to 2cm across are common, together with segregational veins and lenses of quartzo-feldspathic material up to 10cm thick, bordered by selvedges of biotite and mafic constituents.

The biotite in the rocks defines a schistosity which parallels the banding in all areas except in the nose regions of intrafolial F1 folds of the banding (sections 2.1.1 & 2.1.2).

The Grey Gneisses are fine- to medium-grained (0.1-1mm) and granodioritic in composition (fig.111), richer in biotite and plagioclase and poorer in quartz and potash feldspar than the Pink Gneisses. Compositional banding is less evident and where segregational lenses are seen, they are thin, 1-3mm thick. In thin section the schistosity in these gneisses appears much more regular than in the Pink Gneisses and the mafic constituents are more evenly distributed throughout the rock.

The Intermediate Gneisses are more adamellitic in composition (fig.111), and consist of the Grey and Pink varieties interlayered in bands on the scale of 30cm or so in thickness.

The Basic Bodies

The mineralogy seen in these bodies depends on the size of the body sampled and its position on the D2 structures. In the largest (over 5m thick) examples, in the noses of F2 folds, relics of an original igneous mineralogy and texture are preserved. Towards the margins of
## TABLE OF MINERAL CONTENT, MODAL RANGES.

<table>
<thead>
<tr>
<th>ROCK TYPE / RANGE of MODAL MINERAL (%)</th>
<th>UNFLATTENED GNEISS</th>
<th>GNEISS AND MIGMATITE COMPLEX</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Pink Gneisses</td>
<td>Grey Gneisses</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>5-20</td>
<td>30-50</td>
</tr>
<tr>
<td>K-Feldspar</td>
<td>30-40</td>
<td>5-25</td>
</tr>
<tr>
<td>Biotite</td>
<td>5-15</td>
<td>15-25</td>
</tr>
<tr>
<td>Muscovite</td>
<td>0-10</td>
<td>-</td>
</tr>
<tr>
<td>Hornblende</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Tremolite/Act.</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Epidote</td>
<td>T</td>
<td>0-5</td>
</tr>
<tr>
<td>Sphene</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Chlorite</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Garnet</td>
<td>T</td>
<td>-</td>
</tr>
<tr>
<td>Apatite</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Iron Ore</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Zircon</td>
<td>T</td>
<td>-</td>
</tr>
<tr>
<td>Sericite</td>
<td>T</td>
<td>T</td>
</tr>
<tr>
<td>Tourmaline</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Kyanite</td>
<td>-</td>
<td>-</td>
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<tr>
<td>Staurolite</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Graphite</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Calcite</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
such bodies and in the F2 limb regions this igneous character is completely overprinted and the rock is altered from a gabbro to an amphibolite.

In the smaller bodies the amphiboles are partly to completely replaced by biotite and epidote. The petrology of these bodies can therefore be comprehensively described in terms of three end members.

Gabbroic  
Amphibolitic  
Biotitic

<table>
<thead>
<tr>
<th></th>
<th>Gabbroic</th>
<th>Amphibolitic</th>
<th>Biotitic</th>
</tr>
</thead>
<tbody>
<tr>
<td>Diopside</td>
<td>5 per cent</td>
<td>Hornblende</td>
<td>35-99 per cent</td>
</tr>
<tr>
<td>Hornblende</td>
<td>30 &quot;</td>
<td>Flag An 20</td>
<td>1-30 &quot;</td>
</tr>
<tr>
<td>Flag An25</td>
<td>50 &quot;</td>
<td>Biotite</td>
<td>0-20 &quot;</td>
</tr>
<tr>
<td>Quartz</td>
<td>10 &quot;</td>
<td>Quartz</td>
<td>0-10 &quot;</td>
</tr>
<tr>
<td>Epidote</td>
<td>5 &quot;</td>
<td>Iron ore</td>
<td>0-10 &quot;</td>
</tr>
<tr>
<td>Sphene</td>
<td>trace</td>
<td>Epidote</td>
<td>trace</td>
</tr>
<tr>
<td>Iron ore</td>
<td>&quot;</td>
<td>Apatite</td>
<td>&quot;</td>
</tr>
<tr>
<td>Apatite</td>
<td>&quot;</td>
<td>Sphene</td>
<td>&quot;</td>
</tr>
<tr>
<td>Chlorite</td>
<td>&quot;</td>
<td>Chlorite</td>
<td>&quot;</td>
</tr>
</tbody>
</table>

* estimates in thin section.

In the least metamorphosed state, traces of what could be an original igneous compositional layering are visible. It is probably this layering which is responsible, in the metamorphosed state, for the more ultrabasic compositions seen in some places. The plagioclase in these rocks is frequently zoned, becoming more sodic in the outer regions. It is also more sodic than the rest of the mineralogy would indicate, even where pyroxene is still preserved. This is taken to be a feature of metamorphic retrogression. The bodies are frequently cut by younger pegmatites; close to these progress towards the retrograde biotite mineralogy is accelerated. In outcrop the rocks are well foliated though in the thicker, coarse-grained (2-20mm) bodies this foliation is less apparent. Analyses 2, 27, 28, 29, 30, 31, 32, 72, 73, 75, 76, 77 and 78 in appendix 2 show a range of compositions from these bodies.

Discussion

The Unflattened Gneisses are the oldest rocks in the area. Their
composition suggests a correlation with the Jostedal Complex of Bryhni (1966) and Carswell (1973).

The banding in the Intermediate gneisses consists of Pink and Grey gneiss interlayered. This banding at all scales is typical of the Unflattened gneisses regionally.

The origin of the Pink and Grey gneisses is obscure. The earliest textures seen are metamorphic but this does not preclude either a sedimentary or igneous origin. It is clear that some of the basic bodies are igneous intrusions into these gneisses and not metasediments or metavolcanics.

The basic intrusives show very clearly the extent of metamorphic retrogression these rocks have suffered (see section 3.5).

3.2.2 The Gneiss and Migmatite Complex

This group of rocks comprises portions of reworked Unflattened gneiss and a suite of migmatised metasedimentary rocks, which include quartzites, psammites and mica schists. Minor intrusive granitic and pegmatitic bodies and amphibolites are also found.

The amphibolites are basic in composition (fig.108) and field relationships show that they are intrusive into the sedimentary complex.

Sediments

Quartzites. Although similar to the gneisses (both weather grey), the quartzites have a restricted mineral assemblage (fig.111), are more massive and have a waxy appearance. A compositional layering paralleling the foliation in the surrounding gneisses is visible due to an increase in mafic content at certain horizons in the quartzites. Frequently these compositional variations appear gradational on one side and sharp on the other and this is tentatively interpreted as graded
bedding (Myres, 1978). No consistent way up was determined from this because of the complexity of later folding. (The north-west "arm" of the major Quartzite outcrop at (22.00,38.00) appears to be right way up). These rocks preserve the most interesting metamorphic assemblages. The oldest are at the highest grade with the development of kyanite and garnet; younger assemblages show progressive retrogression to chlorite, accompanied by the development of new cross-cutting fabrics in the rocks (see section 3.5 below).

**Mica Schists.** These rocks are found as foliation-parallel lenses within the quartzites. The lenses are up to 1m thick and vary in quartz content (fig.111) from being a micaceous quartzite to a quartz mica schist. Characteristically the more micaceous varieties weather brick red due to secondary iron oxide, whereas the quartz-rich varieties green/yellow. The schists preserve more than one schistosity, one axial planar to tight folds paralleling the regional foliation and a crenulation cleavage at a high angle to this (fig.32). The schists contain staurolite and are interpreted as pelitic beds in the quartzite sequence.

**Quartzose Migmatites.** In outcrop these rocks are distinguished from the migmatites and gneisses by their lighter colour, (light grey with a pinkish hue) and their restricted mineralogy (fig.111). In hand specimen the rocks resemble the quartzites except that they do not develop the translucent waxy appearance of the true quartzites. Feldspars and micas are more abundant than in the quartzites and are found in foliation parallel layers or lenses about 1-4 crystals thick, which are interpreted as being the product of the first stages of migmatisation. The rocks are frequently associated with the quartzites, but also occur separately, e.g. (14.00,45.00). They possibly had their origins as impure sandstones.
Unlike the quartzites no primary layering is detected in these rocks, though, in common with the country gneisses they are well foliated.

**Migmatites**

These rocks comprise the bulk of the Gneiss and Migmatite Complex. They are very similar in composition to the Unflattened gneisses (fig. 111) and include reworked portions of these; the exact amount of Unflattened gneiss incorporated cannot be determined because the re-working ultimately destroys early structures, and without this control compositional differences cannot form a basis for subdivision of the migmatites. However, the metasedimentary portions of the migmatites occasionally contain kyanite and more hornblende is present in some of them than is seen in the Unflattened gneisses. There is also a further segregation of quartz from the feldspars in the metasedimentary leucocratic layers. The quartz forms anastomising, foliation parallel, polycrystalline ribbons of one crystal thickness. These three features, attributed to those portions of the migmatite suite that are of sedimentary origin, together with structural observations, do then provide some basis for separation of the older and younger portions of the Migmatite Complex.

No attempt was made to delimit the exact extent of the metasediments on map 1; they are included in the migmatites and treated as one unit. Analyses 70 and 40 in appendix 2 are of two examples of the more mafic rock types in the Migmatite Complex.

**Igneous rocks**

**Augen Gneisses.** These are intrusive into the older Migmatite Complex, although they form bodies (not more than 30m thick) that are sub-parallel to the regional foliation; low angle cross-cutting relationships are seen and in some localities (12.50, 35.70), the Augen gneisses are seen to surround and include portions of metasedimentary rocks. These
Sediments were subsequently folded with the Augen gneisses in F2, and for this reason they are thought to have been intruded at a late stage in the migmatisation of the sedimentary complex, this event occurring prior to F2. Compositionally, the Augen gneisses are granitic (fig. 111, analysis 60, appendix 2), medium-grained (1-2 mm) and dark grey in outcrop with lensoidal segregation bands of white or pinkish feldspar and quartz, frequently in the form of augen. The augen may be made of single microcline crystals or an aggregate of large feldspar crystals. Myrmekite texture is common. The plagioclase in the rock is albitic (An 6-8), the iron ore is ilmenite and is frequently rimmed by sphene.

**Amphibolites.** Within the Migmatite Complex there are substantial bodies of amphibolite (fig. 112). The origin of these rocks is unclear, but the fact that they are found within the sedimentary complex, are never deformed by F2 folds which fold the sediments and occur as lenses (up to 8 m thick) within the S3a foliation produced by these folds, suggests they are of intrusive rather than sedimentary origin. This view is supported by geochemical analyses of these rocks (figs. 108, 109 & 110) and analyses 41, 43, 62, 63, 64, 65, 66, 67, 68 and 71 in appendix 2, but attempts to distinguish them from their older counterparts in the Unflattened gneisses proved to be unsuccessful. In regions of high strain and retrogression biotite replaces hornblende. Mineralogically then, there are two rock types representative of the basic bodies in the Gneiss and Migmatite Complex:

<table>
<thead>
<tr>
<th>Biotitic Amphibolite</th>
<th>and</th>
<th>Amphibolite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hornblende</td>
<td>0-20 per cent.*</td>
<td>Hornblende</td>
</tr>
<tr>
<td>Plagioclase An 20</td>
<td>3-55 &quot; &quot;</td>
<td>Plagioclase An 30</td>
</tr>
<tr>
<td>Biotite (brown)</td>
<td>30-40 &quot; &quot;</td>
<td>Biotite (brown)</td>
</tr>
<tr>
<td>Quartz</td>
<td>3-8 &quot; &quot;</td>
<td>Quartz</td>
</tr>
<tr>
<td>Epidote</td>
<td>0-25 &quot; &quot;</td>
<td>Chlorite</td>
</tr>
<tr>
<td>Iron ore</td>
<td>2-4 &quot; &quot;</td>
<td>Iron ore</td>
</tr>
<tr>
<td>Sphene</td>
<td>1-5 &quot; &quot;</td>
<td>Sphene</td>
</tr>
<tr>
<td>Chlorite</td>
<td>trace</td>
<td>Apatite</td>
</tr>
<tr>
<td>Apatite</td>
<td>&quot;</td>
<td>Garnet</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Epidote</td>
</tr>
</tbody>
</table>

* thin section estimates.
Fig. 112.

NORTH OSTERØY

OUTCROPS OF
AMPHIBOLITES

\ = Amphibolites from the Migmatite Complex

Late D2 Amphibolites

Km
The rocks are well foliated and while in the biotitic members the grain size is small (0.1-5mm), in the amphibolitic varieties the rocks are spotty in hand specimen; clumps of mafic minerals up to 10mm across made of crystals up to 2mm across are set as augen in a paler plagioclase-rich matrix. Frequently these knots of mafic minerals are centred round cores of allotriomorphic iron ore. As with the older amphibolites, the zoned plagioclase is sodic and is considered to reflect metamorphic retrogression.

Discussion

The rocks of the Gneiss and Migmatite Complex form a cover to the Unflattened gneisses. They appear to consist of more or less migmatised sediments and basic intrusions, and are tentatively correlated with the Fjordane Complex of Bryhni (1966) and Carswell (1973).

Early metamorphism at middle to upper amphibolite facies conditions was followed by metamorphism under progressively lower facies conditions. This is particularly well shown by the amphibolites, which developed chlorite in the final stage.

The early metamorphism represents the highest grade attained by any of the rocks of north Osterøy.

3.2.3 The Flattened Gneiss and Migmatite Group

There are two varieties included in this group of rocks, both are formed from the Unflattened gneisses and the gneiss and migmatites and are therefore compositionally very similar to them (fig.113), (spanning the range granite, adamellite, granodiorite) but in higher states of finite strain. The first of these varieties is found on the high ground in the Northern gneiss outcrop (map 1) and is recognised in the field by a strengthening of the (S4) foliation in the rock; it
### Fig 113.

<table>
<thead>
<tr>
<th>ROCK TYPE / GROUP</th>
<th>FLATTENED GNEISS AND MIGMATITE</th>
<th>Micaceous Members</th>
<th>METASEDIMENTARY ROCKS</th>
<th>Quartzites</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Type 1</td>
<td>Type 2</td>
<td>Type 3</td>
</tr>
<tr>
<td>Quartz</td>
<td>25-30</td>
<td>17-45</td>
<td>20-25</td>
<td>38-43</td>
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<td>Plagioclase</td>
<td>15-35</td>
<td>0-8</td>
<td>10-15</td>
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</tr>
<tr>
<td>K-Feldspar</td>
<td>20-37</td>
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<td>-</td>
<td>10-20</td>
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<tr>
<td>Biotite</td>
<td>10-20</td>
<td>5-27</td>
<td>20-25</td>
<td>0-5</td>
</tr>
<tr>
<td>Muscovite</td>
<td>0-2</td>
<td>45-55</td>
<td>15-20</td>
<td>25-40</td>
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<tr>
<td>Hornblende</td>
<td>0-5</td>
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</tr>
<tr>
<td>Tremolite/Act.</td>
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<td>-</td>
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<td>Epidote</td>
<td>0-3</td>
<td>-</td>
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<tr>
<td>Sphene</td>
<td>0-1</td>
<td>-</td>
<td>-</td>
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</tr>
<tr>
<td>Chlorite</td>
<td>0-3</td>
<td>0-2</td>
<td>-</td>
<td>-</td>
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<tr>
<td>Garnet</td>
<td>T</td>
<td>0-10</td>
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<td>1-2</td>
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<tr>
<td>Zircon</td>
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<tr>
<td>Sericite</td>
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<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Tourmaline</td>
<td>-</td>
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</tr>
<tr>
<td>Pyroxene</td>
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<tr>
<td>Kyanite</td>
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<tr>
<td>Staurolite</td>
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<tr>
<td>Graphite</td>
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</tr>
<tr>
<td>Calcite</td>
<td>-</td>
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</tr>
</tbody>
</table>
becomes apparent in outcrop (fig.19). There is also a general reduction in grain size when compared to the Gneiss and Migmatite equivalents. The latter are medium- to coarse-grained, while the flattened rocks are fine- to medium-grained (0.01-0.5mm). The more quartzose and basic units in the Flattened gneisses are correlated with the Quartzose Migmatites and amphibolites of the Gneiss and Migmatite Complex. Mineralogically the characteristic feature of these rocks is that biotite is the dominant mafic mineral with very little hornblende, and small euhedral crystals of inclusion-free garnet are common, dispersed in the rocks overgrowing the S4 foliation.

The second variety is found locally developed everywhere in the Northern Gneiss Unit, within zones of D5 shear. Entering these zones, there is a marked reduction in grain size and a strengthening in the schistosity (S6), which is rotated parallel to the walls of the zone.

3.3 LITHOLOGIES OF THE SOUTHERN GNEISS UNIT

The rocks comprising the Southern Unit (B) are broadly subdivided into three groups (fig.104). The rocks in the last two groups are found more to the south-western margin of the Southern gneiss outcrop, along the tectonic junction with the Arcs, while the rocks of the first group form the bulk of the outcrop. The rocks comprising the first group, while very well foliated, are, for the most part, extremely heterogeneous and have been lumped together as "Undifferentiated gneiss". However, traceable units are identified within these gneisses. These are not petrologically distinct and are distinguished more on the basis of their relative homogeneity, which shows them up from the surrounding gneiss, or by the fact that they are unusually rich in vein material. Such gneisses are given local names.
TABLE GROUPING THE LITHOLOGIES OF THE SOUTHERN GNEISS UNIT

3) Metasedimentary Rocks:

These rocks form a distinct suite comprising quartzites and mica schists cropping out in a band along the contact with the Bergen Arcs. Phyllonitic and mylonitic rocks found in this band are also included here.

2) Igneous Rocks:

The rocks included in this group, the majority of the amphibolites and some of the augen gneisses, form a younger igneous suite. Most of the rocks involved are found in the south-western half of the Southern Gneiss outcrop.

1) Gneissose Rocks:

This group of rocks comprises lithologies very similar to those found in the Northern Gneiss Unit. The group includes the Nonkletten Gneiss, the Drangskollen Gneiss, the Heldalfjell Gneiss, the Risplingen Gneiss, Granitic Gneiss, Pegmatite-Rich Gneiss and Undifferentiated Gneisses. The igneous rocks included here are part of the country rock gneisses into which younger igneous rocks (2 above) were emplaced.

3.3.1 Metasedimentary Rocks

Most of these rocks occur in a schuppen zone beneath the thrust contact with the Bergen Arcs (map 1). There are two main rock types present, mica schist and quartzite, but during the thrusting these were tectonically interleaved with gneissose rocks belonging to group 1. This process resulted in a mixing of the three lithologies, producing thinly banded, fine-grained quartzose mylonites and phyllonites.

Micaceous Members

There are three varieties of mica schist present (fig.113), the first of these crop out as long, thin (3m), foliation parallel lenses, grey
in colour with a silvery sheen on foliation surfaces. Garnet is present in places in this variety. The second variety is only found on Drangey (06.00,32.00) and forms short thick lenses (150x30,). The rock is purplish when fresh, but weathers brown and is rich in garnets up to 3mm in diameter. The third variety of mica schist also has a restricted outcrop, being found as small pods in the gneisses at (21.30,17.00); it is the only variety to be found away from the schuppen zone. When fresh this variety has a silvery appearance and weathers to a pale green colour with gold tinges; garnet is not found in this variety. The mica schists are generally well foliated with large "oyster shell" muscovites up to 2cm across.

Feldspar and quartz in the rocks are mainly confined to thin isoclinal folded layers. Crystal size usually does not exceed 1mm, except for garnet and potash feldspar which occur as porphyroblasts up to 5mm across, the former frequently showing snowball inclusion trails of quartz and graphite.

Phyllonites. Found as small lenses along strike from or beside type 1 mica schists in the schuppen zone (20.25,06.00); these rocks are distinguished from them by the higher proportion of non-micaceous minerals they contain (fig.113). The micaceous material in the phyllonites rather than forming bands in the rock, forms numerous "pip shaped" lenses, ranging in size from 3x5mm to 4x10cm. Within these pips grain size ranges from 0.1-2mm and the lenses are enwrapped by thin impersistent micaceous layers. The rocks are grey/green in colour and less silvery than the mica schists. They are also much richer in potash feldspar porphyroclasts than the mica schists and tend to show more late stage retrogression with the development of calcite and chlorite. In thin section muscovite is seen to occur as distorted books (fig.114), rather than as continuous sheets.
Fig 114.

**DISTORTED MUSCOVITE BOOKS FROM A PHYLLONITE**

Matrix of Finer Grained Quartz and Feldspar

K-Feldspar

Muscovite

Late Iron Stained Shear Zone

K-Feldspar

Iron Ore

Epidote

Quartz

Fig 115.

**HORNBLENDE-QUARTZ-K-FELDSPAR POD**

Knot of Quartz, K-Feldspar & Hornblende

Plane of Foliation

Light Grey-Pink Gneiss

Pink K-Feldspar

Hornblende

Quartz

Foliation Wrapped Round the Pod

Dark Grey Gneiss Band
Quartzites

These form a very distinctive lithology in the schuppen zone with the type 1 mica schists, into which they frequently grade laterally. The bodies are never more than 6m thick, medium-grained and, when pure, milk white in colour, becoming green/grey as the mica content increases. Petrologically they fall between two end members, one a quartz mica schist and the other a pure quartzite (fig.113). These rocks preserve older fabrics very well; several generations of schistosity are seen, some depicted by the orientation of muscovite flakes, some by quartz lamellae and some by crystallographic lattice orientations (figs.123 & 124). The earliest fabric may be bedding, but this cannot be proved. The rocks are cut by several phases of quartz veins, some of these are isoclinally folded.

Quartzose Mylonites. These rocks differ from the 'normal' sediments in that they are richer in feldspar (fig.116, and analyses 5 and 10 in appendix 2). Characteristically the rocks are fine- to medium-grained, dark grey, with a strong foliation, numerous potash feldspar augen and porphyroclasts of coarsely crystalline rock fragments up to 1cm across. The lithology is associated with the schists, phyllonites and quartzites in the schuppen zone and forms a gradational rock type between the sediments and the group 1 gneisses. The most quartz-rich varieties are distinguished from the quartzites because they are fine-grained (0.1mm) and almost black in hand specimen. The difficulty in differentiating between these rocks and the gneisses other than in thin section means that this rock type is not distinguished on map 1, but the rock type is very important within the schuppen zone.

Discussion

The metasediments are the highest structural unit of the Osterøy gneisses. They have been included within the Bergen Arc lithologies by
## Table of Mineral Content, Modal Ranges

<table>
<thead>
<tr>
<th>ROCK TYPE / GROUP</th>
<th>RANGE OF MODAL MINERAL (%)</th>
<th>METASEDIMENTARY ROCKS</th>
<th>IGNEOUS ROCKS</th>
<th>GNEISSOSE ROCKS</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Quartz Mylonites</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Quartz Rich</td>
<td>Quartz Poor</td>
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<td>15-30</td>
</tr>
<tr>
<td>K-Feldspar</td>
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<td>7-40</td>
<td>25-40</td>
<td>10-25</td>
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<tr>
<td>Biotite</td>
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<td>Muscovite</td>
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<td>0-10</td>
<td>0-3</td>
<td>6-15</td>
</tr>
<tr>
<td>Hornblende</td>
<td>-</td>
<td>-</td>
<td>0-5</td>
<td>-</td>
</tr>
<tr>
<td>Tremolite/Act.</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Epidote</td>
<td>-</td>
<td>-</td>
<td>1-6</td>
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<tr>
<td>Sphene</td>
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<td>T</td>
</tr>
<tr>
<td>Chlorite</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>T</td>
</tr>
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<td>Garnet</td>
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<td>Kyanite</td>
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<td>-</td>
</tr>
<tr>
<td>Staurolite</td>
<td>-</td>
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<td>-</td>
</tr>
<tr>
<td>Graphite</td>
<td>-</td>
<td>-</td>
<td>-</td>
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</tr>
<tr>
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<td>T</td>
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<table>
<thead>
<tr>
<th></th>
<th>Heildalsfjell Gneiss</th>
<th>Pegmatite Rich Gneiss</th>
<th>Rissingen/Granitic Gneiss</th>
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<td>Quartz</td>
<td>35-64</td>
<td>20-33</td>
<td>35-45</td>
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<td>Plagioclase</td>
<td>5-6</td>
<td>10-30</td>
<td>15-20</td>
</tr>
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<td>K-Feldspar</td>
<td>10-25</td>
<td>25-45</td>
<td>35-38</td>
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<td>Muscovite</td>
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<tr>
<td>Hornblende</td>
<td>-</td>
<td>-</td>
<td>T-4</td>
</tr>
<tr>
<td>Tremolite/Act.</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Epidote</td>
<td>-</td>
<td>-</td>
<td>T</td>
</tr>
<tr>
<td>Sphene</td>
<td>-</td>
<td>-</td>
<td>T</td>
</tr>
<tr>
<td>Chlorite</td>
<td>-</td>
<td>-</td>
<td>1-4</td>
</tr>
<tr>
<td>Garnet</td>
<td>T</td>
<td>T</td>
<td>0-5</td>
</tr>
<tr>
<td>Apatite</td>
<td>T</td>
<td>T</td>
<td>-</td>
</tr>
<tr>
<td>Iron Ore</td>
<td>T</td>
<td>T</td>
<td>T-2</td>
</tr>
<tr>
<td>Zircon</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Sericite</td>
<td>0-15</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Tourmaline</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Pyroxene</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Kyanite</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Staurolite</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Graphite</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Calcite</td>
<td>T</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
previous workers (Kolderup & Kolderup, 1940). The implication is that
these rocks formed a parautochthonous cover sequence, restacked and
imbricated during the emplacement of the allochthonous Bergen Arc nappes.

No correlation is made between these rocks and the Northern Gneiss
Metasedimentary Complex. The rocks are different in colour, composition
and in the relationship of the mica schists to the quartzites. A
minimum age of Ca. 800 Ma. has been determined for the Southern gneiss
metasediments (appendix 1). This correlates well with similar ages
determined from xenolith-bearing granites within the Bergsdalen Nappes (Pringle
et al., 1975). It is with these sediments and those of the Samnanger
Complex (Faerseth et al., 1977), of the Bergen Arcs, that these rocks are
correlated.

3.3.2 Igneous Rocks

There are two rock types involved in this group, augen gneisses and
amphibolites, which occur as foliation parallel sheets in the gneisses.
The augen gneisses are more extensively developed than the amphibolites.
Both rock types are subdivided into two separate tectonic members, one
set found in or close to the schuppen zone and the other found throughout
the Southern gneisses.

Augen Gneisses

This quartz monzonitic-granitic rock type is perhaps the most
distinctive in the Southern gneisses. It forms, for the most part,
thick (50-400m) regional foliation concordant sheets up to 1400m long.
Contacts with the surrounding gneisses are sharp and locally discordant
with the local gneiss foliation; it is unclear whether this discordance
is a product of deformation or an original intrusive feature. The rock
appears rough and mottled in outcrop because it contains up to 40 per cent.
large (up to 3 cm across) augen porphyroblasts of potash feldspar.
These are set in a medium-grained (0.5-2mm) dark grey, mafic-rich matrix with a strong foliation. There is a considerable variation in augen content, from dark, augen-poor to pink, augen-rich portions. The variation occurs laterally both along and across the sheets. In the latter case the centres are seen to be richer in augen than the margins (pls.42-45); the augen at the margin are also smaller. Thin section work shows that this variation is the product of tectonism rather than a primary feature. There is also a tendency towards local migmatitic segregation of the matrix into leucocratic and melanocratic layers up to 1cm thick. The leucocratic layers often 'include' the augen to form long, thin, light coloured bands.

The general petrology of these rocks is given in figure 116 and analyses 21-26 and 51-60 (appendix 2), but the garnet-bearing examples come only from the easterly outcrops, while the main outcrops in the west are devoid of garnet. The feldspar porphyroblasts may be single microcline crystals or crystal aggregates, showing several phases of regrowth; they may be poikilitic (pls.51 & 52). Sphene in the rocks forms large euhedral crystals (2-3mm), while the epidotes, except for allanite, are smaller and anhedral. The plagioclase in the rocks, except for odd relics of normally zoned crystals with cores of An 30, is sodic (An 7-8) and forms small interstitial crystals in the matrix and exsolutions in perthitic augen. In general the darker varieties of the rock type are finer-grained than their pink counterparts.

Amphibolites

There appear to be two distinct varieties of amphibolite in the Southern gneisses. Type 1 is only found in the schuppen zone intruding the sediments. It occurs as foliation parallel dark green to black sheets with a strong internal foliation. The sheets range in size from 2x20m to 50x8000m, the longer bodies varying in thickness along their
length. The rocks are generally medium- to coarse-grained (0.5-5mm) but they become fine- to medium-grained (0.01-1mm) in the thinner regions and develop a paler green colour. The fine-grained rocks (0.20, 30.20) are serpentinitic and have a soapy feel in hand specimen.

MINERAL CONTENT:

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Biotite-rich</th>
<th>Amphibole-rich</th>
</tr>
</thead>
<tbody>
<tr>
<td>Actinolite</td>
<td>25-60 per cent.*</td>
<td>80 per cent.*</td>
</tr>
<tr>
<td>Iron ore</td>
<td>0-10 &quot; &quot;</td>
<td>5 &quot; &quot;</td>
</tr>
<tr>
<td>Sphene</td>
<td>0-5 &quot; &quot;</td>
<td>3 &quot; &quot;</td>
</tr>
<tr>
<td>Biotite (green)</td>
<td>0-60 &quot; &quot;</td>
<td>29-35 &quot; &quot;</td>
</tr>
<tr>
<td>Chlorite (and talc)</td>
<td>0-40 &quot; &quot;</td>
<td>3 &quot; &quot;</td>
</tr>
<tr>
<td>Apatite</td>
<td>0-2 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Plag. albite-andesine</td>
<td>0-30 &quot; &quot;</td>
<td>6 &quot; &quot;</td>
</tr>
<tr>
<td>Epidote (and clinozoisite)</td>
<td>0-20 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Quartz</td>
<td>0-23 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Sericite</td>
<td>0-8 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Calcite</td>
<td>0-2 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Garnet</td>
<td>0-2 &quot; &quot;</td>
<td>2 &quot; &quot;</td>
</tr>
<tr>
<td>Potash feldspar</td>
<td>trace</td>
<td></td>
</tr>
<tr>
<td>Dravite</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* thin section estimates.

This range of composition is a response to a retrograde metamorphism and deformation and to appreciate the mineralogies developed in individual rocks, it is best to consider two end members, one biotite-rich, one amphibole-rich.

<table>
<thead>
<tr>
<th>Biotite-rich</th>
<th>Amphibole-rich</th>
</tr>
</thead>
<tbody>
<tr>
<td>Biotite</td>
<td>60 per cent.*</td>
</tr>
<tr>
<td>Epidote</td>
<td>20 &quot; &quot;</td>
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<td>Quartz</td>
<td>15 &quot; &quot;</td>
</tr>
<tr>
<td>Sphene</td>
<td>3 &quot; &quot;</td>
</tr>
<tr>
<td>Iron ore</td>
<td>1 &quot; &quot;</td>
</tr>
<tr>
<td>Calcite</td>
<td>1 &quot; &quot;</td>
</tr>
<tr>
<td></td>
<td></td>
</tr>
</tbody>
</table>

* point counted
Other examples of both end members are richer in plagioclase, which usually forms about 20 per cent. of the rock.

Type 2, again cropping out as foliation parallel sheets, differs from the first in that the bodies are much smaller (18.00, 26.00), and found throughout the Southern gneisses. In outcrop they are dark green and frequently rich in biotite and relict garnet. The mafic minerals in these rocks tend to form clumps (up to 0.5 cm across), imparting a coarse-grained appearance in hand specimen; foliation in these rocks is therefore apparently weaker.

**MINERAL CONTENT:**

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Percentage</th>
</tr>
</thead>
<tbody>
<tr>
<td>Actinolite and hornblende</td>
<td>3-70 per cent.*</td>
</tr>
<tr>
<td>Biotite (green)</td>
<td>10-35 &quot; &quot;</td>
</tr>
<tr>
<td>Flag. (albite-oligoclase)</td>
<td>0-22 &quot; &quot;</td>
</tr>
<tr>
<td>Quartz</td>
<td>5-25 &quot; &quot;</td>
</tr>
<tr>
<td>Potash feldspar</td>
<td>0-15 &quot; &quot;</td>
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<tr>
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</tr>
<tr>
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<tr>
<td>Apatite</td>
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</tr>
<tr>
<td>Epidotes</td>
<td>0-1 &quot; &quot;</td>
</tr>
<tr>
<td>Garnet</td>
<td>0-5 &quot; &quot;</td>
</tr>
<tr>
<td>Calcite</td>
<td>trace</td>
</tr>
</tbody>
</table>

* thin section estimates.

A third rock type grouped with the amphibolites is found in the schuppen zone; epidote-rich pods up to 0.5m across are found scattered in the gneisses, enwrapped by the foliation. Contacts with the surrounding gneisses are sharp and the bodies do not seem to be aligned as boudins; their origin is unclear; they may be igneous or metasomatic. In hand specimen they are light green, fine-grained and comprise:


<table>
<thead>
<tr>
<th>Mineral</th>
<th>Percentage</th>
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<tbody>
<tr>
<td>Hornblende</td>
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<td>Clinzoisite</td>
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<tr>
<td>Quartz</td>
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</tr>
<tr>
<td>Potash feldspar</td>
<td>1 &quot; &quot;</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1 &quot; &quot;</td>
</tr>
<tr>
<td>Apatite</td>
<td>1 &quot; &quot;</td>
</tr>
<tr>
<td>Iron ore</td>
<td>trace</td>
</tr>
</tbody>
</table>

* Point counted.

Discussion

Apart from the epidotites, the amphibolites are all basaltic and very similar in composition, analyses 4, 8, 9, 12, 13, 22, 23, 35, 38, 39, 43, 49, 61 and 68 (appendix 2). Subdivision of these and the augen gneiss rocks into two sub-varieties is done on the basis of minor variations in petrology and sometimes the type 2 amphibolites are less basic than those confined to the schuppen zone, and the augen gneisses in this zone contain no hornblende and less garnet. Moreover, type 1 amphibolites do not contain I3 and the augen gneisses outside the schuppen zone are more gradational with the country gneisses at their contacts. These factors when taken together with the igneous geochemistry (figs. 108, 109 & 110), probably indicate that the two sub-varieties are separate intrusions.

The two-intrusion hypothesis is supported by independent structural evidence. The type 1 amphibolites and augen gneisses confined to the schuppen zone cut the sediments at (06.20, 32.30), xenoliths of quartzite are found in the augen gneiss, while the rest of the augen gneisses and type 2 amphibolites are older than the sediments forming part of the old Southern Gneiss Complex, having deformed with it for a protracted period of its history. Thus not only are there two separate intrusions, but two intrusive episodes represented; this provides an insight as to the
composition and nature of the rocks from which the Southern Gneiss Unit was derived.

3.3.3 Gneissose Rocks

These form the bulk of the Southern gneiss outcrop. For the most part they are well foliated, grey acid rocks with dispersed feldspar augen. Occasionally layers rich in hornblende or quartz or feldspar or biotite are found, and while these variants are not volumetrically important individually, they do impart a great heterogeneity to the gneisses. The more distinct units are mapped (map 1), because their stratigraphy 'proves' the thrust beneath the Bergen Arcs to be crosscutting. These distinct units are given suitable local place names, while the rest of the rocks are grouped as 'Undifferentiated Gneiss'. The named units are distinguishable mainly on their weathering characteristics, a function of both composition and strain state, they are therefore not strictly lithological but lithotectonic divisions.

Helldalsfjell Gneiss

This forms a lensoid body 4000m long and up to 400m thick, best exposed at (15.00, 29.50), on the top of Helldalsfjell. The rock is distinguished by its brown weathering colouration and the high muscovite/biotite content seen in hand specimen (fig.116). It is a paragneiss, probably originally an impure quartzite rich in quartz and micas. It is well foliated, the micas distributed in schistosity parallel bands as well as disseminated throughout the rock. Generally medium-grained (0-2mm), these rocks characteristically show F5 folds at all scales because of the banding and high mica content.

Pegmatite-Rich Gneiss

Both units of this gneiss type (one at (10.50, 33.00) and the other at (18.00, 22.00)), are distinguished by the large amounts of vein and
pegmatite material they contain. The veins range up to 1m thick and crosscut the dominant foliation in the gneisses at a high angle, later to be folded with them, developing a new foliation sub-parallel to that in the host rocks. Granitic in composition, these intrusives are coarse-grained (0.1-5cm) and rich in feldspar and quartz. These pegmatites are seen throughout the Southern gneisses, but in the pegmatite-rich rocks they form up to 30 per cent. of the rock by volume. A description of the intrusives will be given later; the petrology of the host rock is given in figure 116. The gneisses are granitic, fine- to medium-grained (0.01-1mm) with a strong foliation, and rich in small augen of polycrystalline feldspar up to 1cm in diameter.

Rispingen and Granitic Gneisses

Forming very distinctive outcrops, clearly seen on air photographs at (20.80,15.00) and (21.00,10.50) respectively, these two rock types are grouped together because they are so similar. They are hard weathering red/pink granites forming features of positive relief (fig.116). The outcrops are about 3000m long and a maximum of 400m thick and well foliated. The Rispingen Gneiss is coarse-grained (0.5-3mm) and, in hand specimen, appears to be made of a mosaic of leucocratic augen separated by thin diastomosing mafic-rich layers. The augen are generally pink, but may have white kernels. The Granitic Gneiss, on the other hand, is fine-grained (0.01-1mm) and in hand specimen the leucocratic components, again separated by thin diastomosing mafic layers, form long thin lenses rather than augen. Throughout the rock, irrespective of the layering, small (up to 1mm) crystals of iron ore are seen, while hornblende is absent. These features suggest that both rock types were probably originally orthogneisses. The Granitic Gneiss in particular has been modified by later deformation and metamorphism which has destroyed hornblende, reducing grain size and producing iron ore.
Drangskollen Gneiss

This is a well-foliated, medium-grained (0.5-2mm), light grey weathering gneiss, best exposed on the highest point of Drangøy (05°80', 33°50'), after which it is named. It is the clearest and most easily traceable of all the gneissose units, maintaining a homogeneous character right along the 10000m outcrop from one coast of Osterøy to the other; as such it forms a marker band. The width of the outcrop varies from 10m to 800m in some places. It is characteristically poorly banded, but with a strong schistosity, preserving L3 very well. D5 quartz veins are also very well developed in this lithology but few F5 folds are seen. Weathered surfaces frequently show ice chatter marks and rough surfaces expose small (1-2mm), spindle-shaped quartz crystals orientated with L3. Fresh surfaces are dark grey with small (0x3x1cm) white albite porphyroblasts lying in the schistosity defined by the micas.

The rock is adamellite (fig.117), and largely recrystallised. The leucocratic minerals overgrow the mafics and albite porphyroblasts are seen to be late poikiloblastic developments. The restricted mineralogy and general uniformity of this lithology suggests it may have been an early intrusion into the gneisses, deformed, reworked and metamorphosed with them.

Nonkletten Gneiss

Named after the hill on which it is best seen (10°30',32°70'), this gneiss type also crops out on the tip of Drangøy, forming a lensoid body 3000m long and up to 400m thick. The rocks are coarse-grained, well foliated hornblende gneisses (fig.117 and analyses 16 &18, appendix 2). The hornblende, while disseminated throughout the rock, is best developed along thin planes, which define the banding. These are frequently the sites on which pink, porphyroblastic feldspar augen, up to 3cm in diameter, nucleate. The hornblende is always anhedral, intergrown with almandine
Fig 117. Table of Mineral Content, Modal Ranges.

<table>
<thead>
<tr>
<th>ROCK TYPE / GROUP</th>
<th>GNEISSOSE ROCKS</th>
<th></th>
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<tbody>
<tr>
<td></td>
<td>DRANGSKOLLEN GNEISS</td>
<td>NONKLETEN GNEISS</td>
<td>UNDIFFERENTIATED GNEISS</td>
<td>MAFIC</td>
<td>QUARTZ</td>
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<td>Calcite</td>
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and sphene and much replaced by biotite. The strong banding in these rocks means they show the F5 folding well at all scales and while they abut the Helldalsfjell paragneisses, there is no direct evidence that they are paragneisses themselves.

Undifferentiated Gneisses

These form the bulk of the Southern gneisses and show many compositional and textural variations. The rocks fall into three broad groupings: Mafic gneisses, rocks of all grain sizes, of dark colour with or without small infrequent feldspar augen and with the compositional range shown in figure 117; Quartz gneisses, which are lighter coloured except when fine-grained (when they can appear almost black), frequently with large (up to 2cm) feldspar augen and within the compositional range shown in figure 117; and Feldspar gneisses, frequently white or pale pink in outcrop with rarer light grey varieties. There is a broad range of grain size in the Feldspar gneisses and feldspar augen are common, similar to those found in the Quartz gneisses. The compositional range is given in figure 117.

The rocks are schistose and frequently well-banded with concentrations of mafic constituents. Members of the group may themselves be banded, alternating with each other locally across strike. None of the units comprising such a layering is very thick or laterally persistent. Plagioclase in these rocks is rather sodic, usually albitic (An 5-8) and never exceeding An 30. If the crystals are more than 0.1mm in diameter they are usually normally zoned too and large late albite crystals full of epidote and chlorite inclusions are common, but most of the augen in the rocks are single or polycrystalline microclines.

Two compositionally remarkable rock types were found within these gneisses. The first, best seen at (09.43,30.05) contains:
Quartz 25 per cent.*
Potash feldspar 20 " "
Sodic plag. 5 " "
Biotite (green) 1 " "
Chlorite 1 " "
Sphene 2 " "
Hornblende 50 " "
Apatite trace

* Thin section estimates.

The rocks are found in pink or grey feldspathic gneiss as foliation parallel lenses up to 8cm thick (fig.115), which separate the gneisses into bands, some 15cm thick. Knots of subhedral hornblende occur in these lenses, with crystals up to 5cm across. These are set in a matrix of coarsely crystalline pink feldspars in which meandering polycrystalline quartz veinlets occur.

The second rock type is a quartzite, found outside the schuppen zone at (15.61,23.10), as a lens in a quartz gneiss; it consists of 84 per cent. quartz, 15 per cent. muscovite and 2 per cent. kyanite. The kyanite is found as small crystals sheared into the S4 schistosity, clearly from a paragneiss predating this structure and as it is the only kyanite-bearing rock found in the Southern gneisses it cannot be included with the rest of the Southern Gneiss Quartzites; it may be a relic of an older assemblage.

Discussion

There is a great variety of rock types present in the gneissosose unit, including rocks inferred to be of both igneous and sedimentary origin. Analyses 15-17, 20, 34, 36, 37, 45 and 47 (appendix 2), are of rocks from this group.

3.3.4 Origin of the Southern Gneisses

The Southern Gneisses form a thrust wedge. It is inferred that they are a para-autochthonous nappe derived wholly from the northern "basement",...
with a suite of igneous intrusives emplaced along thrust planes during deformation and reworking. All rock types present in the Southern gneisses have their counterparts in the Northern gneisses and they have both been metamorphosed at Amphibolite facies (kyanite grade) early in their development.

This explanation is also consistent with the interpretation of the structures. The differences between the two gneiss units are thought to be due to tectonism.

3.4 ACID VEINS IN THE OSTERØY GNEISSES

The Osterøy Gneisses are cut by numerous veins which show a great range in composition and texture. Varieties which crosscut the foliation are generally thicker, do not have selvedges and are interpreted as intrusions into the gneisses. It is recognised that this intrusive material may originate locally, from within the gneisses. Other varieties with selvedges are usually conformable with the local foliation and are interpreted as having generated locally during migmatisation. The veins are then, in Sederholm's (1967) terminology, arterites and/or veinites respectively.

Compositionally the varieties found are aplites, granites, pegmatites and trondhjemites. An example of granitic veining is seen at (11.00, 34.60). This is a multi-stage intrusion, im thick, preserving xenoliths of earlier magma types. Trondhjemites, the least common vein type, are seen at (17.50, 17.00). Aplites and pegmatites form the bulk of the veins, the latter predominating. Some of the pegmatites are virtually monomineralic composed of 80 per cent. microcline and 20 per cent. quartz. Some of these have the quartz segregated out into layers within the vein. In some pegmatites euhedral garnets are found. These veins are emplaced at many stages in the tectonic development of the gneisses. The garnetiferous pegmatites, because of the presence of garnets, are thought to be emplaced.
during the thermal maxima.

The final type of vein comprises 99 per cent. quartz with occasional potash feldspar knots up to 5cm in diameter. Veins of this composition have been emplaced throughout the history of the gneisses, but by far the most important set were cognate with D5.

Most of the veins in the gneisses are highly deformed and are beginning to merge with the host rocks, which makes them difficult to study from a petrological viewpoint, but as they commonly preserve at least one tectonic fabric, they are useful as tectonic markers.

3.5 METAMORPHISM

A synopsis of the structural character of the north Osterøy gneisses is given in figure 103. Deformation across the area has been shown to be fundamentally heterogeneous (section 2), accounting for the preservation of old mineral assemblages which developed with the early deformation phases. The majority of the rocks in the area are acid in composition and the variations in mineral assemblages seen associated with the older structures are related to 'fossil' metamorphic events rather than variations in rock composition.

The metamorphic petrology and textural development of each metamorphic event will be discussed in conjunction with the deformation structure with which it is associated, starting with the oldest.

3.5.1 Possible Early Metamorphism (M "Early")

Establishment of the structural chronology indicates that the oldest rocks are the Unflattened Gneisses (fig.104); these preserve the oldest structure (S "Early"). Unfortunately, the mineral assemblages accompanying its formation were completely destroyed by later metamorphisms. In view of the mineralogical nature of S "Early" (section 2.1.1), it would appear
that the rocks were probably at Amphibolite facies conditions during its formation.

3.5.2 First Metamorphism (M1)

Possible evidence for the next metamorphism is again found in the Unflattened Gneisses. In the nose regions of the F1 folds, relict crystals of garnet are occasionally found, while in the limb regions of the same folds these relics are absent. In the basic rocks of this group, hornblende is a relatively common mineral. It could be argued that these relics of Amphibolite facies mineralogy belong either to S "Early" or were developed after D1 in a subsequent metamorphism. However, the presence of hornblende in the basic rocks, together with a relict igneous mineralogy, points to the Amphibolite facies minerals having formed during D1. Whether or not there was an intervening period of retrogression between D "Early" and D1, or whether the minerals developed as the product of a continuing metamorphic event cannot be determined.

Uplift

The next event affecting the environment of the Unflattened Gneisses is determined by inference. It is proved (section 3.2.2) that the Gneiss and Migmatite Complex contains rocks, undeformed by F1, which were originally sediments deposited on the Unflattened Gneisses. At the time of deposition therefore, the Unflattened Gneisses must have been at the surface.

Any mineral or textures which may have developed in response to this uplift have, however, been completely masked by the effects of subsequent metamorphic events.

3.5.3 Second Metamorphism (M2)

The second metamorphism is at Amphibolite facies, but is divided into
two stages on the basis of an intervening deformation which separates the
thermal peak from the advent of its decline.

Following their deposition the sediments were migmatised and certain
of the more micaceous members developed kyanite and staurolite, while
a dark green hornblende (α yellowish green, β olive green, γ dark green),
and almandine developed in the rest of the suite. The presence of these
minerals indicates that the migmatisation occurred at conditions correspon­
ding to the medium pressure subfacies of the Amphibolite facies, the
equivalent to the kyanite and sillimanite zone in the Barrovian series
(Miyashiro, 1973). The effects of the migmatisation were not solely
confined to what is now the Gneiss and Migmatite Complex, but also locally
affected the Unflattened Gneiss. At (11.20, 34.60, fig.118), secondary
potash feldspar overgrows the S1 layering, leaving a "ghost" outline of
F1 fold noses.

In thin section all minerals, except quartz, sphene and iron ore can
be seen to be progressively replaced by potash feldspar, so that K feldspar
constitutes up to 60-70 per cent. of the rock. From this point quartz
is progressively replaced. The potash feldspar crystals are anhedral
and up to 2mm in diameter, forming a mosaic in which the quartz is left
as an interstitial mineral. This is clear evidence for at least local
mobility of elements, specially potassium.

Similar processes are seen to affect the sediments, e.g. the quartz
migmatites and quartzites. Here, however, the effects are more widespread,
preumably because their original mineralogical composition was unstable
under Amphibolite facies conditions, while the Unflattened Gneisses already
had a mineralogy stable at approximately these conditions. It is interesting
to note that while the migmatisation is accompanied by fabric development
(S2), it is not accompanied by folding; (S2 and the accompanying minerals
are deformed by F2, which took place under Lower Amphibolitic facies
Fig 118.

SKETCH SHOWING THE EFFECTS OF THE M2a METAMORPHISM ON THE UNFLATTENED GNEISSES

S1 in Sheared F1 Limb

'Ghost' of S1

Late Pegmatite

'Ghost' of S 'Early'

(Pink) Area of K-Feldspar Metasomatism

S 'Early', Fold by F1

0 M 3
conditions). In circumstances where deformation and metamorphism occur together it is unusual for a metamorphic peak completely to precede the deformation (Fyfe et al., 1978, pp346-7).

(M2b)

The minerals which accompanied the development of S3 are again only poorly preserved but hornblende and almandine are found growing in relics of the S3a cleavage. Hornblende also developed as an essential mineral in those basic rocks intruded during D2: the hornblende is very similar to that grown during the main migmatisation. Most of the mineral growth occurred in the second stage of S3 development, the less flattened rocks show very little new mineral growth. The dominant mafic mineral to develop in M2b was biotite, and it is this which emphasises the strong S3b schistosity. All these features point to a continued stability under Amphibolite facies conditions during D2, with new mineral growth occurring only in regions of highest strain, where the S3b schistosity developed. There was however a slight retrogression from the conditions which prevailed during the migmatisation, as growth of kyanite and staurolite apparently stopped.

Uplift

Following D2 another return to near surface conditions is indicated by the presence of more sediments deposited on the gneisses. These are seen in the zone of imbrication beneath the Bergen Arcs. Any mineralogical changes that may have accompanied this uplift are again masked by younger events.

Discussion

The deposition of the second sedimentary sequence marks the penultimate event in the metamorphic development of the Osterfj gneisses. It was followed by another metamorphism at Amphibolite facies, terminating in a
long retrogression to Greenschist facies conditions. This last metamorphism (described below), was largely responsible for the poor preservation of the mineralogical evidence for the metamorphic reactions involved in the earlier events. Discussion of this evidence has therefore been limited to the fabrics with which the paragneissces were associated. With these limited relict assemblages any argument as to the reactions which produced them would be pure speculation; indeed without the structural evidence the metamorphic chronology of any specimen cannot be determined because the minerals produced in M1 and M2 are so similar.

Evidence for the early metamorphic events is taken entirely from the Northern gneisses; in the Southern gneisses the ultimate metamorphic event is too strong.

3.5.4. Third Metamorphism (M3)

The last metamorphism to affect the Osterøy gneisses, M3, reached its peak at Amphibolite facies and retrogressed through the Greenschist facies. It was responsible for the bulk of the mineral parageneses and was thus the most pervasive metamorphism. In the Northern gneisses its effects wax up succession south-eastwards and in the Southern gneisses, it virtually obliterated all evidence for earlier metamorphisms. It differs from M1 and M2 on two counts; firstly, a retrogression following the thermal peak can be proved, secondly, the thermal peak (and retrograde steps) are synchronous with deformation maxima. Four steps in the metamorphic retrogression are identified, all best developed in the Southern gneisses from where they will be described.

(M3a)

This event is seen in the gneisses, metasediments and Type 2 igneous rocks. The gneisses are banded so the mineral paragenesis must be examined in each component of this banding which is itself a product of D3.
In the mafic bands, an interlocking granoblastic texture between biotite (yellow green–dark green), hornblende (green yellow, bluish green, bluish green), allanite, quartz, plagioclase (A2 29-35), almandine, garnet and sphene are seen. The only minerals which are possibly a relic from an older paragenesis are sphene and quartz (pl. 53). These occur in two forms, as small crystals poikilitically enclosed by hornblende and as larger crystals which form part of the granoblastic texture. The crystals in the hornblende are interpreted as relics while the larger crystals are interpreted as relics on which a mimetic overgrowth has occurred during the metamorphism. The hornblende laths are aligned parallel to L3, while the biotite grows to define the S4 schistosity. The garnet is subhedral and the allanite is now metamict and virtually isotropic.

In the felsic bands the paragenesis is less well preserved, it consists of biotite (yellow green–dark green), quartz, sphene, allanite, plagioclase (An 29-35) and potash feldspar. The biotite again defines the S4 schistosity.

Microcline porphyroblasts forming augen are found throughout the rock. Also, lenses or "strings" of polycrystalline, coarse-grained microcline and quartz (pl. 54) are found, which have narrow biotite selvages and are interpreted as products of a slight migmatisation which accompanied this metamorphism.

In the metasediments, the paragenesis of this metamorphism is less well preserved, but almandine porphyroblasts are common. Though presently enwrapped by a younger foliation, these crystals are internally zoned, the marginal zones are clear or contain minute graphite inclusions while the central zones are full of quartz inclusions which form rotational trails, "snowball structure" (pls. 55 & 56). The garnets are interpreted as having grown during M3a, and the rotational trails in the central zones
indicate this growth was syntectonic. The outer rims either grew very rapidly or postdate the deformation. The rotational inclusion trails are generally confined to garnets within the micaceous metasediments. In garnets in the quartzites the zoning is still seen and the inclusions of the central zones are still at an angle to the enveloping schistosity but they are not rotational. This is interpreted as a reflection of the difference in the response of the two rock types during deformation.

In Type 2 amphibolites hornblende (≈ green yellow, φ bluish green, χ bluish green), almandine and plagioclase (An 30-35) are seen lying in S4, and the hornblendes are again aligned parallel to L3. The Augen Gneisses, which intrude the sediments, are similar to the host gneisses except the hornblende lacks sphene inclusions and the sphene forms very large (up to 2mm) crystals of typical lozenge shape. The microcline porphyroblasts too are different, showing a wide range of perthite exsolution textures (pls. 52, 58, 59, 64, 78 & 103).

Discussion

The alignment of the hornblende laths with L3, biotite with S4 and the rotational inclusion trails within the garnets, prove the syntectonic nature of this metamorphism. However, the garnets show at least three rims. The centres are full of quartz inclusions, while the outer rims show only minor graphitic trails which do not have rotational geometries. This indicates that the thermal event outlasted the tectonism.

The mineral paragenesis is similar to that developed in M2b. As a consequence of later metamorphisms it is incomplete, for example, the only mineral still preserved in the metasediments is garnet. It is therefore difficult to define exactly the metamorphic environment, but the blue/green colour of the hornblende and the An 25-30 content of the
plagioclase put it somewhere in the Amphibolite facies (Miyashiro, 1973, p. 299), probably somewhere between the low pressure and medium pressure type. A question then arises as to which K-feldspar polymorph formed during this event. The polymorph presently seen is microcline, but this is known to have been crystallised in the subsequent metamorphic "steps" and could therefore have replaced original orthoclase. A replacement like this would be consistent with the complex perthite textures observed, and it is inferred that perthites (e.g. Pls. 52, 58, 59, 64, 78 & 103) are a part of the M3a paragenesis.

The metamorphic differentiation which produced the banding, selvedges, microcline porphyroblasts and K-feldspar and quartz segregation lenses (pl. 54) prove a marked chemical mobility during this metamorphism. This mobility probably also explains the hornblende knots and lenses found in the Undifferentiated gneiss (fig. 115). Allanite seems to be an essential mineral of the paragenesis. The crystals are large, euhedral and, as with sphene, it is present in both the mafic and felsic regions of the rock. It is also possible that apatite was a stable mineral, but this cannot be proved, as like microcline it is clearly recrystallised in subsequent metamorphic events. The presence of allanite is unusual; it is interpreted as forming as a 'house' for rare earths mobilised in the Amphibolite facies environment.

(M3b)

This event was responsible for the initial retrogression of the M3a paragenesis and is called the clinozoisite event.

In the gneisses, hornblende and biotite are replaced by secondary biotite (yellow green - dark green), defining a new schistosity, S5 (pl. 60), at a slight angle to S4. Hornblende is also replaced by actinolite (α yellowish green, p green, α dark green) and tremolite,
which sometimes rim hornblende crystals, leaving unaltered cores. However, actinolite and tremolite have usually grown as a new mineral during formation of S5. The early hornblende and biotite are also corroded and embayed by quartz and feldspars (pls. 60 & 61), so that they become interstitial, anhedral relics in the new mineral paragenesis. Clinozoisite formed both as anhedral, metamict rims round allanite nuclei (pl. 62), and as discrete euhedra.

Both K-feldspar and plagioclase replaced the mafic minerals partly by enlargement of pre-existing crystals and by crystallisation at new sites. The later plagioclase is albitic (An 6-8), while relict plagioclase from the M3a paragenesis is strongly zoned (pl. 63), with sodic rims. The early plagioclase is also largely replaced by quartz and albite in a similar way to the mafic minerals (pl. 63). The zoning in the relics is interpreted as a product of secondary leaching of calcium rather than an original growth feature, because it is patchy. The M3a microcline porphyroblasts are partially replaced by myrmekite (pl. 64), but also became clear as they recrystallised during the metamorphism. New porphyroblasts also formed in the S5 schistosity; in general these are less perthitic than their M3a counterparts (pls. 65, 51 & 59). Sphene continued to crystallise in S5 by further nucleation on pre-existent cores and quartz continued to segregate in the new banding forming 'ribbons' (pl. 54) and 'strings' (pl. 66). Similarly, potash feldspar forms segregation veins and lenses, though these only have very weakly developed selvedges of biotite.

In the amphibolites, similar reactions to those discussed above are seen, but more clinozoisite forms and actinolite and tremolite become the dominant amphiboles. In the Type 1 amphibolites, all the amphibole is tremolite; it constitutes as the main mafic mineral in the rock, indicating intrusion and crystallisation during the M3b event. This view is supported by the fact that none of the minerals are aligned with L3 and
by the occurrence of relics of interstitial plagioclase, grown with the
tremolite, with an An composition above An 20. After crystallisation
these minerals were overgrown and poikilitically enclosed in dravite
(watery yellow - straw yellow) porphyroblasts; no clinozoisite formed in
this particular rock type, which distinguished it from the rest of the
amphibolites. The Type 1 amphibolites were probably intruded towards
the end of this stage of M3.

(M3c)

This event (the epidote event) is characterised by epidote growth and
is concurrent with S5 and constitutes a retrogression of M3b. Tremolite
and actinolite are replaced by a third growth of new biotite which is aligned
with S5, while remnants of hornblende are replaced directly by biotite.
The secondary biotite of M3b is overgrown by epidote, and epidote rims
formed on clinozoisite grains. Epidote also formed rims on allanite and
a third rim on allanites already rimmed by clinozoisite. The volume of
epidote formed at this stage in the retrogression far exceeds the total
volume of allanite or clinozoisite produced earlier. It constitutes the
only epidote mineral in the Western Amphibolites, which supports the view
that these were intruded during late M3b. In general the behaviour of
the rest of the mineral paragenesis is a continuation of that seen in M3b,
but myrmekite formation round microcline porphyroblasts is more noticeable
(pl.67), and replacement textures, attributable to solution and replacement
effects are better preserved (pl.55, fig.119). Otherwise microcline con­
tinued to form porphyroblasts and albite (An 5-6) is the dominant plagioclase.

(M3d)

This event is characterised by the development of muscovite. The
mineralogical changes which characterise the event started during the forma­
tion of S5 and continued through D5 to accompany the formation of the S6
foliation.
Fig 119.

STRESS CONTROLLED CORROSION OF GARNET IN M3c

SECTION CUT PARALLEL TO S6

First Shell of Garnet Full of Inclusions

Secondary Clear Rim

PROFILE SECTION THROUGH F5

Garnet

Epidote

Replacement by Quartz Only Occurs Parallel to the F5 Axial Plane

[solution across the plane]

Remnant of Clear Rim Here
Epidote and biotite continued to crystallise, but in places, especially in the metasediments, biotite was replaced by muscovite with an accompanying exsolution of iron ore. The ore is of several varieties: skeletal ilmenite developed intergrown with sphene; magnetite and pyrite formed euhedral crystals dispersed through the rock; minute magnetite crystals formed in the cleavage planes of the developing muscovite as it replaced biotite, and there is much submicroscopic iron ore deposited along grain boundaries, especially along the axial planes of microscopic F5 folds. These ore crystals overgrow all except the S6 fabric.

There is evidence for the presence of mineralising solutions during M3d; shear zones formed along the axial planes of F5 folds are filled by vein quartz with accessory microcline. The sheared rock is annealed by recrystallisation of the felsic rock constituents. This recrystallisation caused embayment of the mafic constituents, which, realigned into S6 by the shearing, survive as anhedral interstitial relics in the new paragenesis. This mobility of acid material is also seen elsewhere; quartz in the metasediments annealed to form laths which define S6 (p.46), microcline and albite both developed as porphyroblasts with markedly anhedral "amoeboid" outlines (p.68). The An content of the albites is unknown because the crystals do not twin; characteristically they include laths of mica and epidote (pls.68,69,70 & 102). The inclusions are not always from the matrix because they are euhedral and crystals like these are not found in the matrix. Calcite also crystallised during this phase of the retrogression. Though it is seen in all rock types, it is best developed in the metasediments, where, augened by the S5 schistosity, it is distinguished from interstitial calcite of later age.

In the amphibolites this metamorphism finds expression not in muscovite development, but in the production of talc and epidote from the breakdown of tremolite and plagioclase (interstitial high An plagioclase). Schorlire (buff olive), also developed in these rocks and the micaceous metasediments.
The crystals are euhedral, zoned (with paler centres), and overgrow the S5 schistosity.

**Discussion**

The three metamorphic steps described above define a regime of declining temperature through the Greenschist facies (Miyashiro, 1973). This thermal retrogression, overlapping several deformations, was obviously gradual and implies an accompanying reduction in hydrostatic pressure. A notable point is the close association of the new parageneses with the developing structural fabrics. The ability of the mineralogy to equilibrate with the retrogressive environment is closely controlled by tectonism. The reasons for this are discussed in section 4.2.1.

Chemically the retrogression is best described in terms of the movement of the cationic components of the minerals, especially 'Fe'+, 'Ca'+ and 'Al'+, though there must also have been some rearrangement of the 'Al - Si - O'- anions. The movement is accommodated through the processes of recrystallisation and by movement of fluids through the rocks.

The original M3a paragenesis, developed throughout the rocks may be summarised:

(Hornblende + allanite + andesine + biotite + orthoclase + almandine + apatite + quartz). The fact that the plagioclase is in the andesine range means that with a reduction in temperature, the recrystallising plagioclase enters the peristerite field, yet only albite (An less than 10), is seen. The stages in the retrogression of the M3a paragenesis can therefore be summarised:

M3b

Andesine + hornblende $\rightarrow$ Tremolite/actinolite + Clinozoisite + albite producing a new paragenesis which is:

(Tremolite/actinolite + clinozoisite + albite + biotite + microcline + apatite + quartz).
Andesine + hornblende $\rightarrow$ biotite + epidote + albite

Tremolite/actinolite + albite $\rightarrow$ biotite + epidote

producing a new paragenesis which is:

(Biotite + epidote + albite + microcline + apatite + quartz).

Andesine + hornblende $\rightarrow$ biotite + muscovite/talc + iron ore + epidote + calcite + albite

Tremolite/actinolite + albite $\rightarrow$ biotite + muscovite/talc + iron ore + epidote + calcite

Biotite $\rightarrow$ Muscovite + iron ore

producing a new paragenesis which is:

(Muscovite/talc + iron ore + albite + microcline + epidote + biotite + calcite + apatite + quartz).

For all the above reactions, there is no need to invoke any large-scale metasomatism in the rocks, the proportions of the ions involved in the reactions would appear to balance out. Also in the amphibolites, most of the reactions occurred at the contacts between the reacting phases. In the acid gneisses, however, the new minerals are not directly localised round nuclei of outgoing phases, the new minerals such as epidote, grow randomly in the rocks. This implies some local mobility of ions but not necessarily large-scale metasomatism.

This mobility of ions suggests the presence of a fluid phase in the rocks during metamorphism, allowing migration in an intergranular 'film'. The development of calcite/(dolomite) in M3d supports this view, for while the $\text{Ca}^2+$ can have origins from within the rocks, the $\text{CO}_3^{2-}$ must have come from a volatile phase because no minerals found in older parageneses contain the $\text{CO}_3^{2-}$ ion.

There is a problem with regard to apatite. Euhedral crystals are present in the parageneses right up to M3d and it is unclear whether they
are euhedral because they continually recrystallised or whether they are relics from the M3a paragenesis, unaffected by subsequent events.

(M3e)

Unlike the previous stages of the retrogression, this stage is not specifically associated with a developing schistosity and occurs throughout the rocks. The characteristic minerals formed, however, grew preferentially filling the planes of S7 fracture cleavage. These minerals are green chlorite (anomalous blue birefringence) and a grass green epidote. Away from these fractures, biotite and garnet are replaced by chlorite. In the garnet chlorite grew in random orientations, whereas in the biotite replacement started along the biotite cleavages from the ends of the 'books' and moved into the centre, producing a dumbell shape (fig.120). Hornblende and tremolite/actinolite are also replaced by chlorite to some extent both in the amphibolites and the acid gneisses and iron ore is exsolved as small euhedral laths along the cleavage planes. Calcite also crystallised in this metamorphic stage to form an interstitial mosaic in some of the rocks in the schuppen zone beneath the Major Arc. It was also deposited in hydraulic fractures formed at this time (pl.48). Any rock sheared in this stage of the retrogression (the D6 thrusts), did not anneal as in M3d, and remained as a friable rock flour.

Discussion

This metamorphism is interpreted as the last stage in the retrogression of the M3a paragenesis, the changes being the result of ground waters migrating through the gneisses, hydrating some of the minerals. The mineralisation of the S7 cleavage is the result of the passage of these solutions out of the rocks. The fact that there is no annealing in the shears suggests that at this stage the thermal environment is too low for activation of annealing in the felsic rock components.
Fig 120.

CHLORITE REPLACING BIOTITE WITH ACCOMPANYING EXOLUTION OF IRON ORE, AN M₃ₑ PHENOMENON

(Deformation Predates Replacement)

Exolved Iron

Chlorite Starts to Replace the Biotite Along the Obvious Cleavages Especially Where they are Bent

Chlorite

Uncorroded Outlines

Book of Biotite, Bent with Corroded Outlines

0 mm 0.5
This metamorphism only affects the D8 joints and their wall rocks. Within the joint zones the rocks are shattered and form a breccia of rock fragments set in a matrix of fine crystal fragments. This matrix is extensively replaced by green epidote and calcite; in one instance crystals of stilbite were also found. For about twenty meters either side of the largest joints the feldspars and gneisses are red, because of the presence of minute hematite grains. These probably originated from solutions passing up the joints, for, although the gneisses are generally sericitized, it is only in association with the joints that this red colour is seen.

Discussion

The event is interpreted as a final dewatering of the rocks as they were uplifted to their present level. The mineralising solutions probably originated from below where the rocks were in an M3e type environment.

3.5.5 Conclusion

There are four distinct metamorphic events that have affected the Osterøy gneisses, the last and most important of which is best developed in the south of the area, but is also responsible for most of the mineral assemblages seen in the north.

These metamorphisms are not maxima in a continuing event because they are separated by periods of sedimentation. Therefore, there are at least three complete metamorphic cycles preserved, each widely separated in time. The gaps between the cycles and the duration of the metamorphisms themselves, except in the final case, are unknown but the cycles clearly relate to separate orogenies; the final metamorphism is Caledonian. These observations fit well with observations made as to the structural history of the area and together they are used in the formulation of the deformational
subdivisions outlined in the Structural section. Figure 149 provides a summary of the metamorphic history of the North Osterøy gneisses.
During the course of the present investigation it was realised that many of the rocks on Osterøy show features which match those described in the literature from mylonitic zones (Lapworth, 1885; Holmquist, 1910; Tiermier & Boussac, 1911; Sander, 1911; Sander, 1912; Staub, 1915; Quensel, 1916; Horne, 1930; Peach, 1930; Christie, 1960; Higgins, 1971; Bell & Etheridge, 1973; Nicolas & Poirier, 1976 and Sibson 1977). Whilst in places these are found in discrete zones and thus are interpretable as forming under classical conditions of mylonite formation, viz. (Lapworth, 1885; Higgins, 1971; Sibson, 1977 and figs. 121 & 122), large areas of gneisses (3 to 4km thick and 12 to 15km long) with similar characteristics cannot be thus interpreted because of their great thickness and regional development.

The occurrence of mylonitic features on such a large scale demands explanation beyond that normally given for mylonite formation and thus requires both a review of classical concepts and development of theoretical arguments. Thus the "mylonitic" rocks of Osterøy are described in terms of their modes of occurrence and characteristic megascopic, mesoscopic and microscopic features, and a broader theoretical consideration is given to their possible formation and that of classical mylonites.

4.1 THE OSTERØY MYLONITES

4.1.1 Description

The rocks in the area between the Tysse Thrust and the eastern boundary of the Major Bergen Arc, the Southern Gneisses (fig.10), have been affected by two major periods of deformation, each giving rise to foliations (sections 2.2.2 & 2.2.3). Where intensely developed, these deformations have resulted in an apparent structural simplicity despite the fact that the rocks are the most structurally evolved; D3 was coincident with an amphibolite facies metamorphism and produced a completely penetrative fabric
TEXTURAL CLASSIFICATION OF FAULT ROCKS

[After:– SIBSON, 1977]

<table>
<thead>
<tr>
<th>RANDOM – FABRIC</th>
<th>FOLIATED</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Fault Breccia</strong></td>
<td>?</td>
</tr>
<tr>
<td>&gt; 30% fragments</td>
<td></td>
</tr>
<tr>
<td><strong>Fault Gouge</strong></td>
<td>?</td>
</tr>
<tr>
<td>&lt; 30%</td>
<td></td>
</tr>
<tr>
<td><strong>Pseudotachylyte</strong></td>
<td>?</td>
</tr>
<tr>
<td><strong>Crush Breccia</strong></td>
<td>&gt; 0.5 cm fragments</td>
</tr>
<tr>
<td><strong>Fine Crush Breccia</strong></td>
<td>0.1–0.5 cm</td>
</tr>
<tr>
<td><strong>Crush Microbreccia</strong></td>
<td>&lt;0.1 cm</td>
</tr>
<tr>
<td><strong>Protocataclasite</strong></td>
<td>Protomylonite</td>
</tr>
<tr>
<td><strong>Cataclasite</strong></td>
<td><strong>Mylonite</strong></td>
</tr>
<tr>
<td><strong>Ultracataclasite</strong></td>
<td><strong>Ultramylonite</strong></td>
</tr>
<tr>
<td><strong>Phyllonites</strong></td>
<td><strong>Blastomylonite</strong></td>
</tr>
</tbody>
</table>
**MODEL FOR THE PRODUCTION OF MYLONITES**

[After :=Sibson, 1977]

NOTE, The Overall Regime is One of SIMPLE SHEAR [Faulting]
throughout the rocks (S4 & L3); D4 occurred under middle to upper green-schist facies conditions and produced discrete bands of new fabric (S5). Both deformations produced folds which became flattened into the developing foliations but the rock types produced in each event are quite distinct.

D3 Mylonites

D3 affects rocks which were very similar in character and composition to those described from the Northern gneisses (pl.71); it homogenises them metamorphically and texturally (pls.72, 73 & 74). Except in discrete lensoid bodies up to 30m long (tectonic pips), all the early characteristics of the rocks are lost and effectively a new rock type develops (pl.75).

The changes involved in D3 include a general reduction in grain size with accompanying growth of new minerals and formation of new textures from the decomposition of the old ones (pl.76). In detail, certain minerals, such as sphene, are seen to fragment and spread out in the plane of the developing foliation (pl.77) where the fragments act as nuclei for later sphene growth. Other minerals, such as biotite and hornblende, are destroyed in the margins of the "tectonic pips" (pl.60). They recrystallise with a deccusate texture in bands within the new schistosity; the new schistosity bears no causal relationship to that preserved in the pips because the schistosity within the pips cannot be traced into the new younger schistosity developed outside them (pl.75). Porphyroblastic minerals, which are often poikiolitic, including small crystals from the fine-grained matrix (pl.53), grow in the rock. In the case of garnet, the inclusions sometimes form rotational trails (pl.56).

As well as the relict "tectonic pips", at a microscopic level, porphyroclasts of early minerals are occasionally seen enwrapped by the new schistosity. These occur as single crystals or polycrystalline aggregates which may be polymineralic. Feldspar frequently forms porphyroclasts,
early plagioclases are found with bent twin planes and undulose extinction
and are much corroded by new, developing minerals (pls. 57 & 63).
Frequently these are also fragmented. Potash feldspar often forms augen
in the rocks, either as single crystals showing undulose extinction
(pl. 79), or as many crystal fragments or subgrains of varying size (pls. 76,
80 & 81). In some cases subgrains recombine to develop a single crystal
which may be smaller (pl. 82), of the same size (pls. 83 & 84), or larger
(pls. 51, 85 & 86) than the original; the original size being determined
from the deflection of the enveloping schistosity (pl. 84). In the last
case the margins of the new crystals are crenulate (pls. 65, 86 & 87),
extending out into the matrix like "pseudopodia" (pl. 85) enclosing matrix
crystals (pl. 87) to form a poikilolitic texture (pls. 51, 85, 86 & 87).
These augen are thus both porphyroclasts and porphyroblasts; in both cases
the foliation is deflected round the nucleus (pls. 81, 82, 83 & 84).
Recrystallisation effects are also accompanied by the complex perthite
relationships described previously (pl. 58).

Coeval with the processes described above was the segregation of
quartz into small ribbons parallel to the foliation, with very irregular
margins clearly cross-cutting matrix minerals (pl. 54); their growth then
was partially metasomatic because they are replacing other matrix minerals.
The quartz in these ribbons and elsewhere in the rocks shows undulose
extinction (pl. 88), subgrain development and annealing features similar
to those described by White (1971, 1973a, b, c, 1975a, 1976) and White and
Treagus (1975), from rocks deformed at high strain rates. Where the rocks
are composed of more than 80 per cent. quartz interlocking crenulated
crystal margins develop (pl. 101). It is inferred that this texture is
the product of annealing and recrystallisation of a rock in which the crystal
size was much smaller because of the interlocking pattern of the crystal
margins and the way in which isolated islands of crystals in one orientation
are found in a crystal with another orientation. Where recrystallisation occurred under stress, quartz laths showing undulose extinction developed from the strained grains (pl. 89). The laths mostly lie in the plane of the developing schistosity (pl. 46).

During the development of the D3 mylonitic textures a crystallographic orientation of quartz developed in the quartzites (best seen in thin section using a sensitive tint). Figure 123 illustrates that the orientation is a monoclinic "c" axis fabric with the axis of symmetry around 030/40°. This cluster of "c" axes coincides with a strong mesoscopic rodding fabric (L3) seen in the field. The quartz fabric has therefore persisted through at least two later deformations and must have originally been much stronger. A very weak girdle dipping at 170/30° is also present, which lies within the plane of the schistosity. On Osterý the monoclinic fabrics are unique to the quartzites, a plot of the quartz "c" axes of a granitic gneiss collected a few yards away from the quartzite analysed above, shows no strong fabric (fig. 124), but there is again a weak girdle dipping at 265/20°, within the plane of the schistosity. Both these rocks contain the rodding fabric and intrafolial fold structures equally developed, and both have the same metamorphic and deformational histories.

In the polymineralic rocks, apart from the minerals described above, the remaining minerals play a minor role in the development of distinctive structures. For the most part they form a fine-grained groundmass of irregularly sized anhedral crystals (pl. 90), but in some places the texture becomes almost granoblastic (pl. 91); crystals of the same species in these areas are usually of the same size.

**D4 Mylonites**

The over-all effect of D4 is a further reduction in grain size and an enhancement of the banding or a development of platiness in the rocks (pls. 92, 93 & 94). There is also a darkening in colour of rocks of all
PLOT OF QUARTZ 'C-AXES', [Rock No FH A24]

The contours represent the following percentages:

- 0.0% - 3.0%
- 3.0% - 6.0%
- 6.0% - 9.0%
- 9.0% - 12.0%
- 12.0% - 15.0%
- 15.0% - 18.0%
- 18.0% - 21.0%
- 21.0% - 24.0%
- 24.0% - 27.0%
- 27.0% - 30.0%
- 30.0% - 33.0%
- 33.0% - 36.0%
- 36.0% - 39.0%
- 39.0% - 42.0%
- 42.0% - 45.0%
- 45.0% - 48.0%
- 48.0% - 51.0%

Note: A value of 0.5 > 0.9 is less than and greater than.
Fig 124.

PLOT OF QUARTZ C-AXES, [Rock FH A73]

This plot has undergone -0- rotations using coordinate card 0.

Sampling 2 across -10 down. Poles per 1 percent area 300 data eq ar.

The contours represent the following percentages:

00.9 <= 1 - 25.9, 9x = 3-42.9, 9x = -5.62; 9x = 7-82.08, 9x = 9-102; 9x = 11-122; 12.9
12.9 <= 13-14% 14, 9x = 15-16% 16, 9x = 17-18% 18, 9x = 19-20%; 20.9x = 21-22% 22.9
22.9 <= 23-24% 24, 9x = 25-26% 26, 9x = 27-28% 28, 9x = 29-30%, 9x = 31-32% 32.9
32.9 <= 33-34% 34, 9x = 35-36% 36, 9x = 37-38% 38, 9x = 39-40% 40, 9x = 41-42% 42.9
42.9 <= 43-44% 44, 9x = 45-46% 46, 9x = 47-48% 48, 9x = 49-50%, 9x = 51%.

Note a value of >0.5-0.9x, LT 12% and GT 0.0.
compositions resulting from a decrease in grain size and dispersion of micas.

As a consequence of \( D_4 \), \( S_4 \) and \( S_5 \) lie very close to each other and the effects accompanying the increasing dominance of \( S_5 \) are similar to those described above for \( D_3 \) (\( S_4 \)). Under the microscope, bands of incipient \( D_4 \) shear are seen to start in biotite-rich layers of \( S_4 \), the biotites shearing and fragmenting along their cleavages. The fragments then act as nuclei for new biotite growth but the crystals never regain their former size. The zones of shear pass out of mafic regions into felsic bands, where they form discrete fine-grained bands containing biotite (pl.95).

Development of the \( D_4 \) shear bands is accompanied by all the features which accompanied the development of the \( D_3 \) mylonites with the following slight modifications:

Quartz segregations are frequent but tend to form "strings" along the \( S_5 \) schistosity (pl.66). The cycle of destruction and regrowth of potash feldspar continued but now myrmekite formed round the augen (pls.64 & 67) and fragments of this became sheared into the matrix (pl.96). A poikilolitic form of albite joins potash feldspar as a porphyroblastic mineral forming augen in the \( S_5 \) fabric (pls.68, 69, 70, 97 & 102). \( D_3 \) quartz ribbons developed undulose extinction and deformation laths during \( D_4 \) and in places recrystallised parallel to the plane of the new schistosity (pl.54).

The most striking features developed during \( D_4 \) are found in association with the thrust contact with the Major Arc. At grid ref. (07.80, 29.80) augen gneiss is in contact with dark coloured metasediments (fig.129). At the top of the section, within the contact zone, augen shaped fragments of augen gneiss up to 20cm in diameter are suspended in dark schist characterised by large "oyster shell" muscovites lying along the planes of schistosity. Moving away from the contact the size of these fragments
Fig 129.

**Sketch of the Contact Between the Augen Gneiss and the Metasediments in the Thrust Zone Beneath the Major Arc [A D4 Mylonite Zone]**

- **Augen Gneiss**
- [K-feldspar augen]
- Dark Schist with porphyroclasts of K-feldspar & augen gneiss
- Fine grained black quartzoide schist with K-feldspar augen
- Dark schistose quartzite
- Schistose psammite
decreases until only isolated potash feldspar augen (2-3 mm in diameter) remain. These contact rocks are interpreted as hybrid mylonites formed by the tectonic mixing of sediments and augen gneiss.

There is a relationship between the D3 and D4 folds and the locations of the mylonitic rock types; the new mylonitic fabrics are strongest in the limb regions of F3 folds, where the new schistosity parallels the old; the noses of F3 folds form "tectonic pips". During D4 the S5 schistosity was best developed where S4 was strongest and the grain size was smallest; it has thus in effect reactivated old F3 limbs. It is also developed along the short limbs of the F4 folds; (pl. 23).

The effect of D4 on the rocks described above are therefore as widespread as those of D3 but only become dominant in some areas. Along the junction with the Major Arc, S5 is the dominant fabric, from here it dies out north-eastwards and is seen in the middle of the Southern gneiss outcrop as discrete bands of new schistosity. While these bands are found in the middle of homogeneous rocks, they tend to be preferentially sited along the contacts between gneiss types; D4 is thus responsible for the variation in grain size described approaching the margins of the augen gneisses (see section 3.3.2).

Other mylonitic rocks

Mylonitic rocks occur elsewhere in the gneisses in association with F5 folds. The middle limbs of these show a marked mechanical degradation of the minerals, but characteristically while the felsic minerals have annealed, maintaining the coherence of the rock, the biotites have not re-crystallised, though they do define a new schistosity (S6). The D5 mylonites are therefore interpreted as having formed at a lower temperature than those of D4.
A second variety of mylonite is found along the D6 faults, where a fine-grained friable cream coloured rock which contains small angular fragments of rock and crystals, is interpreted as a rock flour formed during faulting. Mylonite is also associated with the D6 joints. Within the joint zones are rocks composed of gneiss and crystal fragments, set in a matrix of almost black isotropic powder (pl.50). This material occurs in sharp sided diastamosing bands (pl.47), and may contain voids into which drusy quartz has grown (pl.49). Calcite veins are common, especially in the larger rock fragments (pl.48) and their presence, together with voids, epidote and chlorite mineralisation, indicates formation by the explosive release of pore water during jointing. While such hydraulic pressure breccia is cohesive neither it nor the rock flour contain a foliation.

Discussion

Most of the features described above are characteristic of mylonites (Lapworth,1885). The Southern gneisses are thus a suite of mylonitic gneisses, though they do not conform to the usual description of mylonites in that they occur regionally rather than being confined to narrow zones.

Similar rocks seen elsewhere in the Bergen Arcs have also been called mylonites (Sinha Roy,1977c; Faerseth et al.,1977) and it is clear that mylonitic rocks form a considerable volume of rocks in the region.

Based on Sibson's 1977 classification of mylonites (fig.121), the rock types found on north Osterøy are: proto mylonite, mylonite, phylonite, blastomylonite, ultracataclasite and fault gouge. Most of the mylonites in the Southern Gneisses are blastomylonites.

4.1.2 Mode of Occurrence

The mylonites occur in three distinct situations and on this basis they may be grouped (figs. 130 & 131).
ENvironments in Which Rocks With Mylonitic Textures Were Found on Osterøy (Part 1)

Simple Shear Mylonites

A1 (Simple) Shear Zone Mylonites

Mylonite

A2 Thrust Zone Mylonites

Thrust Zone

Abutment Mylonites

B1 Hard Resistant Rock

Mylonite Forms at the Contact

B2 Tectonic Pip of Resistant Rock
Fig 131.

ENVIRONMENTS IN WHICH ROCKS WITH MYLONITIC TEXTURES WERE FOUND ON OSTEROY, (Part 2)

C1 FOLD ASSOCIATED MYLONITES

(Pure) SHEAR ZONE MYLONITES
[tightening of folds]

Fold Train

Fold Nose & Boudin Relics

Mylonite

C2 MIDDLE LIMB MYLONITES
Fault Mylonites

These occur in close association with faults on Osterýy (fig.130, A2) and this variety is only represented by two rock types; the fault gouge and the D₄ mylonite along the contact with the Major Arc.

Tectonic Abutment Mylonites

Most of the D₄ and some of the D₃ mylonites are of this type. The mylonites are found at the junction of two rock types of differing mechanical strength (fig.130, B1 & B2). The mylonitic rocks are better developed in the weaker member. Such locations of mylonitic rocks can be likened to a strain shadow round a porphyroclast where the schistosity is intensified across the layering at the points of maximum stress and reduced either side of the clast in the protected regions (Harker, 1939; Simpson, 1964). On Osterýy the strain effects are however on the kilometer scale. The "tectonic pips" appear to resist deformation and persist through succeeding deformations unaltered. During subsequent deformation it appears that the highly strained zones are the first to be reworked because veins cross-cutting early mylonites are most deformed in subsequent deformations, producing mylonites where the early mylonitisation was most intense (pl.23).

How the "tectonic pips" first form is problematic; some are remnants of early fold noses but others seem to form from the limb regions of the folds by a process akin to boudinage. Sinha Roy (1977b) suggests that small scale "tectonic pips" or porphyroclasts and high strain regions round them form by the intersection of sets of simple shear zones but this is considered unlikely on Osterýy because no evidence for simple shear zone arrays was found.
Regional Mylonites

These mylonites, which are the product of D3 and, to a lesser extent, D4, form the bulk of the Southern gneisses, which are therefore best described as a regional, flat lying mylonite belt. Characteristically these rocks are closely associated with folding and develop a strong "L" tectonite shape fabric. It has been shown that the tightness of these folds decreases with depth (sections 2.1.5, 2.2.2 & 2.3.1 for D3 & 2.2.3 & 2.3.2 for D4) and since where the folds are open (section 2.1.5), no mylonite develops, it is apparent that mylonitisation is largely the product of "pure shear" flattening during the formation of the tight D3 and D4 folds (pl.99, fig.131-C1). In contrast the D5 mylonites are the product of "simple shear" in the short, middle limbs of asymmetric folds (pl.18, fig.131-C2) and are therefore akin to simple shear zone mylonites (fig.130-A1).

4.1.3 Pilot Geochemical Study

Thin section analysis of some of the mylonitic rocks in varying degrees of deformation suggested that apart from the development of new mineralogies and texture, there was a slight alteration of the bulk chemical composition with increasing deformation. To examine this possibility a suite of samples was collected from an augen gneiss at locality (19.20,09.52), where it enters a D4 (abutment type) shear zone. In the centre of the augen gneiss body the augen are 1.5 to 2cm in diameter but towards its southern contact the size of these decreases sharply. At the contact the over-all grain size is considerably reduced and the augen eliminated; the rock also develops a new parallel banding. Plates 42, 43, 44 and 45 show some of the textural variation among specimens. No specimens were collected within a meter of the contact to avoid contamination due to mixing.

In thin section the principal mineralogical components of the rocks are quartz, potash feldspar, sodic plagioclase, biotite and accessory
chlorite, muscovite, epidote sphene and iron ore. The chemical composition of this suite of samples was determined, the lists of analyses are given in appendix 2 (analyses 52-60).

There are problems in representing the variation in data across shear zones in a way that clearly shows any geochemical trends that might be present. No such method exists in the literature; Beach (1976) defines his chemistry in classical two-component field plots; Sinha Roy (1977b) presents the raw data and states the broad trends. Neither of these methods is suitable because they do not show the details of the trends.

Herein data is presented on an Enrichment/Depletion plot, which is constructed by plotting relative chemical enrichment or depletion against degree of deformation. The data must be plotted left and right of the origin for symmetrical zones or on one side only, as in this case, for asymmetrical zones. The specimens may also be spaced evenly as in this case or scaled to actual measured distances.

The enrichment/depletion value for each specimen is calculated by subtracting the weight per cent. oxide for each element of the most deformed rock from those for each of the specimens. A scaling factor is employed in some cases for the sake of clarity.

On such a plot any trend starting in the negative field and moving towards the positive field represents a relative enrichment in that element during deformation, and vice versa, especially if the slope is well defined. The over-all behaviour of the lines compared with each other may define patterns of behaviour amongst groups of elements.

Figures 125 to 128 present the analyses of major and trace elements for two sets of samples. It is clear that there is an over-all apparent enrichment in SiO₂ accompanying the deformation and a corresponding consistent pattern of apparent depletion in other major elements: MgO, CaO, TiO₂, Fe₂O₃ and P₂O₅.
Fig 125. MAJOR ELEMENT ENRICHMENT / DEPLETION DIAGRAM FOR THE FH-B29 MYLONITE SUITE.

The most deformed rock is...
Fig 126.

TRACE ELEMENT ENRICHMENT / DEPLETION DIAGRAM FOR THE FH B29 MYLONITE SUITE

(X.R.F. Geochemistry)
Fig 127. Least Deformed, M1, M2, Most Deformed, M3

(MAJOR ELEMENT ENRICHMENT / DEPLETION DIAGRAM FOR THE M1-3 MYLONITE SUITE)

X.R.F. Geochemistry

- MgO x 10
- TiO$_2$ x 10
- MnO x 100
- P$_2$O$_5$ x 10
- Fe$_2$O$_3$
- CaO
- Na$_2$O x 10
- Al$_2$O$_3$
- K$_2$O x 10
- SiO$_2$
TRACE ELEMENT ENRICHMENT / DEPLETION DIAGRAM FOR THE M1-3 MYLONITE SUITE

THE MOST DEFORMED ROCK is DEPLETED in
ENRICHED in Ppm

(Y/10, Sr/100, Zr/100, Nb/10, Rb/10)

(X.R.F. Geochemistry)
$K_2O$, $Na_2O$ and $Al_2O_3$ do not follow either of the two trends described above, but rather define an over-all apparent enrichment in $K_2O$ with increasing strain and an apparent depletion of $Na_2O$ and $Al_2O_3$, yet the lines for all three when taken together form a pattern of their own.

Similarly, with the trace element plots, all except Rb are relatively depleted, Sr strongly so. The elements Rb and Zr define a trend pattern separate from the others; that for Rb is clearly similar to that described for $K_2O$.

Discussion

It is impossible to decide from the geochemical evidence alone whether the apparent $SiO_2$ enrichment is the result of addition of $SiO_2$ to the system or the subtraction of all other elements. However, thin section study reveals that the grain size of the mafic minerals decreases with increasing deformation, suggesting that their stability may have decreased because their surface area increased; thus subtraction may have caused the apparent silica enrichment.

A similar problem arises with the apparent $K_2O$ enrichment. Thin section study reveals that microcline, the mineral most likely to carry $K_2O$, forms very large crystals in all the rocks. Thus during deformation microcline was stable and therefore its constituent elements are likely to have remained in the rock. However, it has been ascertained from the enrichment/depletion trends that $K_2O$, $Na_2O$ and $Al_2O_3$ all behave in a related manner. As $Na_2O$ and $Al_2O_3$ show depletion trends, it is inferred that while microcline was stable, plagioclase was unstable and was therefore removed. This conclusion is confirmed by the fact that plagioclase, rather than forming large crystals, is one of the small matrix minerals. The apparent $K_2O$ enrichment is therefore similar to that shown by $SiO_2$, and is not, as classical interpretation would predict, the result of K-metasomatism.
The trace element plots show that Rb runs with the $K_2O$, and was therefore left behind whilst Sr in the rock was strongly depleted. This has important implications when considering ages based on Rb/Sr ratios.

The Y and Nb behave in a manner similar to the bulk of the mafic components. The Zr values must always be treated with caution in these plots, as large variations can be attributed to an uneven distribution of the mineral zirconium in the rock.

Thus the process of deformation is accompanied by the loss of the constituents of mafic minerals, aluminium and sodium.

4.1.4 Discussion on Possible Meaning of Textures

In thin section it is possible to subdivide the mineral textures seen in the Osterøy mylonites into sets. The presence, shape and size of any one crystal (especially plagioclase, K feldspar and quartz) is the result of the interaction of many processes (figs. 134, 135 & 136), but it is possible in theory to determine the texture likely to result from the action of particular processes. The textures in any one rock are the product of two sets of these processes with diametrically opposed results; one set acting to increase crystal size, the other to decrease it. Assuming that any rock is the product of a dynamic equilibrium between these two sets and because the rocks are anisotropic it is possible, on the basis of the relative dominance of certain structures and textures, to identify individual processes responsible for any particular texture.

Textures resulting from processes causing grain size reduction

Textures from chemical processes. These textures are mainly the product of metamorphism which, apart from mechanical processes, is very important in determining the textures present in the rock.
a) Rim myrmekite (symplectites) (pls. 64 & 67)

b) Solution features (pl. 55): These features, because of their very nature are very difficult to identify and are therefore probably much more common than can be proved.

c) Relics (pls. 60 & 61).

Textures from mechanical processes. Textures produced by these processes are further subdivided into those which result from high pore fluid pressures and those formed as a result of high grain/grain contact pressures.

d) High fluid pressure textures (pls. 47, 48, 49 & 50)

The mechanical pressure structures are again subdivided on the basis of whether the stresses were applied at a rate faster or slower than, or equal to the rate at which the mineral can accommodate the strain.

Faster:

e) Cataclastic textures (pls. 77, 96 & 100)

Slower or Equal:

f) Sub grains (pls. 76, 78, 79, 80, 81 & 82)

g) Undulose extinction (pls. 57, 79 & 88)

h) Quartz lamellae (pl. 46)

i) Augen and porphyroclasts (pls. 55, 76, 80, 83 & 84)

j) Shear bands (pl. 93)

k) Quartz strings (pl. 66).

Textures resulting from processes causing increase in grain size

There are two main processes by which mineral grains may increase in size. The first involves the rearrangement of the crystal lattice so that it conforms to, and merges with, adjacent grains of the same species. The process results in reduction of the grain boundary surface area in the rock. The second involves chemical reactions at adjacent crystal boundaries,
or redeposition of a mineral phase at a site of nucleation in the rock, while at other locations this phase is dissolving.

Lattice rearrangement structures

1) Crenulate crystal boundaries (pl.101)
2) Clearing features (pl.91)
3) Granoblastic texture (pl.91)

Chemically induced structures

4) Snowball garnets (pl.56)
5) Poikiloblasts (pls.53, 56, 70, 102 & 86)
6) Amoeboid growth (pls.51, 86, 85, 87 & 68)
7) Porphyroblasts (pls.55, 69, 70, 86, 97, 68 & 102).

These form in two ways, by recrystallisation of a grain originally porphyroclastic in nature, or by nucleation and growth of a new crystal.

8) Embayment and cross-cutting structures (pls.61 & 100)
9) Mimetic rims (pls.55 & 62)
10) Ribbons (pl.54).

Definition of Terms

The terms "crystallisation" and "recrystallisation" used in the following discussion are defined as follows:

Crystallisation: the appearance of a mineral at a site in the matrix for the first time; involves nucleation and growth of a new or similar phase.

Recrystallisation: A mineral is recrystallising if, following subgrain formation, it clears itself of strain and the subgrains reunite, reforming a discrete crystal; this may attain or exceed the size of the parent grain.
4.1.5 General Discussion

The Southern gneisses are unique because while there are other mylonite zones of equal thickness, for instance the Namtin and Namarodo mylonites from the south-west Nile region of Uganda (Hepworth, 1964), the main foliation in the Southern gneisses is subhorizontal rather than steeply inclined or vertical.

There is also a fundamental difference between these thick tracts of mylonite, termed here "regional mylonites", and those typically described from fault zones, for example the Outer Isles Thrust (Sibson, 1977), in that the genesis of regional mylonites is not clearly associated with major thrusts. Indeed it is doubtful whether the term "mylonite" which was originally applied to rocks formed "where two superposed rock systems moved over each other as solid masses" (Lapworth, 1885), is strictly applicable to regional mylonites where no overriding is seen.

A large body of opinion, led by Higgins (1971) regards mylonites as having a brittle, disruptive (cataclastic) origin and therefore diagnostic of fault zones. Recently however it has been proved that the microstructures seen in mylonites (polygonisation, subgrains, serrated grain margins, kink bands and grain aggregates) are equally attributable to ductile deformation, recovery and recrystallisation. Therefore, in part, the processes responsible for mylonite generation can be plastic rather than brittle (Bell & Etheridge, 1973; Nicolas & Poirier, 1976).

All the literature on the formation of mylonites regards the over-all deformation regime as one of simple shear because of the association with fault zones (viz. Higgins, 1971; Sibson, 1977). Whilst this regime is applicable to mylonite zones like the Outer Isles Thrust, it is inadequate in other cases where it is evident that there is a strong pure shear component in the generative deformation (Johnson, 1967).
The Southern gneisses are an example of a pure shear mylonite zone. Apart from the mechanical problems of generating mylonites of this thickness over such a broad region by a simple shear mechanism, it is evident from the mesoscopic structures (see section 2.2), that deformation accompanying mylonite genesis is one of flattening without large-scale displacement (the situation seen in fig.131-C1).

The geometry of the "mylonites" on Osterøy points to the conclusion that mylonites are not a rock type specific to faulting regimes, but are one aspect of a continuum of rock deformation and metamorphism. The theoretical basis for this conclusion and an explanation of the significance and presence of the structures described above is given below.

4.2 THEORETICAL CONSIDERATIONS

It is generally agreed (e.g. Sibson, 1977), that mylonites are produced when the strain rate in a rock exceeds the annealing rate of its component minerals. It is usually thought that they are the product of a single tectonic phase. The rock is stressed, and strain is concentrated in a band in which the rock develops a fine-grained texture and a foliation. The unit so formed may then recrystallise to form a blastomylonite, depending on its depth of burial (Sibson, 1977). (It is, however, understood that subsequently the fault zone may reactivate (Christie, 1960), thus producing subsidiary mylonites which may incorporate portions of those previously formed.) Zones like this clearly exist; the Outer Isles Thrust described by Sibson (1977) being one of them.

The genetic origins of these mylonites are known, and it is possible partially to quantify the strain rates appropriate to their production (White, 1975b; Price, 1975). Figure 132 shows a table of geologically realistic strain rates and their associated structures. Mylonites are produced at strain rates between $10^{-1}$/sec. and $10^{-8}$/sec.
GEOLOGICAL STRAIN - RATES

METEORITES, VOLCANOES & FAULTS

INTRUSIONS

SOFT MUD FOLDING & GRAVITY SLIDES

HARD ROCK FOLDING

MYLONITE PRODUCING DEFORMATIONS

ISOSTASY

STOP

(After PRICE 1975)
The mylonites comprising the Southern gneiss unit on Osterøy do not however fit Sibson's (1977) model, as it can be demonstrated that the mylonites developed concurrently with regional folding in the gneisses, and development continued through two phases of such folding; the metamorphic grade during these processes waned from almandine amphibolite facies, through to lower greenschist facies (section 3.5.4).

There is therefore a paradox, the tectonic facies (mylonites) indicate high strain rates whilst the structural features associated with mylonite formation (folding of hard rocks, i.e. gneisses), suggest the strain rate should be in the order of $10^{-12}$/sec. or less. The mode of occurrence is also different: the Osterøy mylonites are not confined to a discrete band, but are regional in extent.

Any model explaining mylonites must be based on the behaviour of its component minerals and their reaction to stress. Recently advances have been made in the understanding of molecular deformation paths, and a review of these is necessary.

The deformation theory used in geology is based on that developed for metallurgy and ceramics. Plates 104 and 105 model a plane through a crystal. The areas (A) model ideal packing. The bands (E) model crystal boundaries. Across the boundaries at which two crystals of different orientations meet, is an area of mismatch; these areas are almost 50 per cent void space. The bands (F) model subgrain boundaries; here the mismatch and void space are not so great. Areas (E) model strained crystal lattices; microscopically these would be areas of undulose extinction (pl.38). Planes (C) model dislocation planes or stacking faults. The strain in (B) areas can regularise to form (C) planes. Frequently in doing this a hexagonal net structure develops (X, pl.105), (Wenk, 1976; Nicolas & Poirier, 1976). Points (D) model vacancies or point defects. (Areas (E)
and (F) could be considered to be sites of accumulation of (D) type structures.) There is a dependence of vacancy density on grain size, as grain size decreases the proportion of vacancies per unit volume increases. The dependence is controlled by surface area and varies:

\[ \frac{3A}{D} \text{ cm}^2/\text{cm}^3 \]

where

\[ D = \text{average grain diameter} \]
\[ A = \text{the number of vacancies in a 1 cm length of crystal boundary.} \]

Triaxial compression tests have yielded data on the conditions required to generate and mobilise dislocations (Heard, 1960, 1963; Serdengecti & Boozer, 1961; Rutter, 1972(a)). There are at least six independent distinguishable paths by which steady state deformation may be accommodated that have been discovered. All but one (Twinning) involve motion of defects through or around grains. These paths are:

1. Defectless flow (plastic flow without defect formation)

   Plates 104 & 105

2. Glide (movement of edge or plane defect across lattice planes)

   .......... (C) & (E)

3. Creep (this occurs at elevated temperatures and includes the process called climb. It involves movement of an edge or plane defect by addition or subtraction of ions from the plane. The movement is therefore confined within the plane).

   .......... (C) & (E)

4. Coble Creep (movement of point defects around grain and subgrain boundaries)

   .......... (F) & (E)

5. Nabbo Herring Creep (movement of point defects through the body of a crystal)

   .......... (D)

6. Twinning.
The basis of all deformation theory is Frick's first and second laws which describe the movement of point defects (vacancies, foreign or interstitial ions).

The first law governs the rate at which stresses anneal from crystals; it relates the flux of ions at right angles to a direction to the concentration gradient parallel to it.

\[ J = -D \frac{dc}{dx} \]  

Where:  
- \( J \) = The flux of defects  
- \( D \) = The diffusive coefficient  
- \( \frac{dc}{dx} \) = The concentration gradient along the axis.

The second law gives the profile of the gradient.

\[ \frac{dc}{dt} = D \frac{d^2c}{2} \]  

Where:  
- \( t \) = time

Each mechanism can therefore be described mathematically (Elliot, 1973; Wenk, 1976; Nicolas & Poirier, 1976) and Ashby (1972) proposed a composite diagram, "The Deformation Map" (fig. 133), which expresses how rock would deform at steady state under any given environmental conditions. There are, therefore, five main processes by which rock may deform; only the last two readily achieve a steady state condition.

(1) **Defectless flow** (Ashby, 1972). The rate controlling factor is the viscosity of the rock.

(2) **Twinning**, only significant in rocks composed of minerals which twin easily. Deformation is finite because a lattice cannot twin indefinitely in direction or amount.

(3) **Cataclasis** (fracture). Involves crystal breakdown, which is pressure dependent and frictional sliding between resultant fragments. A disruptive rather than continuous process. (This is the mechanism thought by Higgins (1971) to be responsible for mylonite genese.)
Fig 133.

THE DEFORMATION MECHANISM MAP

[After: ASHBY 1972 & RUTTER 1976]

DIFFERENTIAL STRESS \( \text{MN/m}^2 \)

TEMPERATURE °C

THEORETICAL SHEAR STRESS

DISLOCATION GLIDE

DISLOCATION CREEP

NABBRO-HERRING CREEP

COBLE CREEP & PRESSURE SOLUTION

NORMALIZED DIFFERENTIAL STRESS = \( 0 / \text{Shear Modulus} \)
(4) Intra-crystalline processes (Wenk, 1976; Nicolas & Poirier, 1976). This involves the dislocation glide and climb mechanism described above.

(5) Mass transfer or diffusive processes. Included here are coble creep, nabbro herring creep and pressure solution. Deformation is accommodated by migration of point defects (vacancies) from sites of low stress to sites of high stress (to accommodate matter moves in the opposite direction).

Rutter (1976) and McClay (1977 a & b) have added a new field to the Ashby deformation map, that of pressure solution. The mechanism is basically one of the coble type, but the process is enhanced by a thin film of fluid at the grain boundary. The site of redeposition can therefore in this instance be remote from the dissolving mineral.

There is some debate as to whether one or more mechanism at a time can operate on a single mineral aggregate. Mitra (1978) in contradiction of Rutter (1976), concludes from his work on quartzite that a combined flow law, composed of pressure solution and intra-crystalline processes can operate.

Another process not considered by Ashby has recently been proposed. Boullier and Gueguen (1975) in considering mylonite formation suggested a process of "Superplastic Flow". The process is quasi-viscous, involving essentially grain boundary sliding, and allows unusually large tensile deformation, without necking, at high strain rates. Such a process necessitates a marked reduction in viscosity during mylonite formation. Calculations by S. White (pers.comm.1977) for the viscosity of a quartz mylonite suggesting viscosities similar to those for molten granite ($10^{-9}$ poise), support this. Other work on superplasticity is reviewed by Davies et al. (1970), Johnston (1970) and Edington et al. (1976).
The possible paths of deformation in rock are summarised in figures 134, 135 and 136. The processes can be further grouped on the basis of their effect on grain size:

<table>
<thead>
<tr>
<th>DESTRUCTIVE</th>
<th>NO CHANGE</th>
<th>CONSTRUCTIVE</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dislocations</td>
<td>Coble creep</td>
<td>Pressure solution (b)</td>
</tr>
<tr>
<td>Pressure solution (b)</td>
<td>Pressure solution (a)</td>
<td>Chemical changes (as</td>
</tr>
<tr>
<td>Cataclasis</td>
<td>Dislocations</td>
<td>regards the stable</td>
</tr>
<tr>
<td>Chemical changes (as</td>
<td>Grain boundary sliding</td>
<td>phase)</td>
</tr>
<tr>
<td>regards the unstable</td>
<td>Nabbro herring creep</td>
<td></td>
</tr>
<tr>
<td>phase)</td>
<td>Twinning</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Annealing</td>
<td></td>
</tr>
</tbody>
</table>

The Rate of Deformation Processes

The ease with which dislocations migrate is governed by the rate at which the largest ion can move. In rock the largest common ion (94% by volume of the Earth) is (O)²⁻, and the diffusion rate varies from mineral to mineral, depending on bond strengths. The mobility depends also on the ambient temperature. Thermal oscillations cause momentary breaks in these bonds. There is a temperature unique to every mineral (the Activation Temperature or Energy), however, below which the rate at which these breaks occur is insignificant. The variation in this activation temperature from mineral to mineral is seen in the Osterøy mylonites in the comparative amount to which minerals have recovered from deformational events. Giletti, Semet and Yund (1977) investigated the diffusion of oxygen in feldspars; they concluded that these minerals have a low activation temperature and would therefore recrystallise in geological settings at comparatively low temperatures. This is confirmed in the Osterøy rocks, the feldspars are one of the first and last minerals to recrystallise in the mylonites.

The Grain Boundary

This acts as a plane of weakness in rock, it is the source and sink for dislocations and the plane at which chemical changes occur. It is
POSSIBLE DEFORMATION PATHS IN POLYCRYSTALLINE ROCKS, Part 1

SINGLE CRYSTAL PATHS

DISLOCATION Glide & CREEP

CYCLES OF DEFORMATION

ANNEALING & RECRYSTALLISATION

ENLARGED SUBGRAINS

CRYSTAL REORGANIZES

DISLOCATIONS

STRAINED LATTICE

SUBGRAINS

SUBGRAIN WALLS

CATACLASIS

FRACTURE

CRYSTAL FRAGMENTS

(A High Strain Rate - Low Temperature Phenomenon)
Fig 135.

**POSSIBLE DEFORMATION PATHS IN POLYCRYSTALLINE ROCKS, Part 2**

**SINGLE CRYSTAL PATHS**

**COBLE CREEP**

Movement Round the Grain Boundary

(By Way of Dislocation Glide & Creep)

Movement Round Grain & Subgrain Boundaries

**PRESSURE SOLUTION [COBLE CREEP]**

Material Dissolved

Material Deposited

**NABBRO-HERRING CREEP**

Movement Through the Body of the Crystal

(A High Temperature Process)
POSSIBLE DEFORMATION PATHS IN POLYCRYSTALLINE ROCKS, Part 3

POLYCRYSTALLINE MECHANISMS

PRESSURE SOLUTION (COBLE CREEP) (b)

GRAIN BOUNDARY SLIDING

CHEMICAL CHANGES
therefore fundamental in determining crystal size, shape and stability.

The surface energy properties of solids are discussed by Spry (1969) using analogies with liquids; the situation for solids is, however, more complex. There is a high state of order imposed in solid packing, bonds are of the strong ionic type rather than Van der Waals forces and this imposes symmetry and shape on the surfaces, increasing surface area above the theoretical ideal. Furthermore all non-cubic crystals are anisotropic; this anisotropy is reflected in the energies obtaining at the crystal faces. Faces with high energies are the smallest and straightest. In polycrystalline rocks these considerations are of fundamental importance, as they determine which of two adjacent mineral phases will grow if both are stable and crystallising. The presence of defects in or near the grain boundary, because of the ionic nature of most minerals, will have significant effects on the surface energy.

The likely effects of grain boundary parameters on mylonites can therefore be summarised:

(1) Surface molecules are at a higher energy level than enclosed molecules; the surface of the mineral will therefore be the site of all chemical reactions.

(2) Similarly, unless this energy is partially satisfied by adjacent crystals (e.g. symplectite formation) or adsorbed material (e.g. pore fluid), any means by which the surface area may be reduced will be favoured. This can be achieved in two ways:

(a) Surfaces that are crenulate can become planar
(b) The volume of the mineral may be increased (grain growth).
4.2.1 Thermodynamics

It has been stated (2 above), that surface energies tend to minimise. These energies, however, form only one component of the total energy state of rock, the Gibbs Energy \((G)\).

\((G)\) complies with thermodynamic laws and tends always to a minimum possible value. A rock, then, should represent the minimum energy state possible for matter in the conditions under which it was formed. Should the conditions change with time, then so will the state of the matter in order to maintain the minimum possible \((G)\), which we will call \((Gm)\). The change in the state of matter, whilst obeying the equilibria laws will, because matter does not behave as a continuous medium, go in a series of steps. \((Gm)\) will therefore have stepped values too, while \((G)\) will be able to assume all intermediate positions because it is not directly dependent on matter but also on other variables (Powell, 1978).

We can, therefore, examine what happens to \((G)\) as one of its component variables changes. The values of \((Gm)\) as determined by the matter in the rock can be mapped on the trace of the variation in \((G)\). For example, let us consider the effect of varying the thermal energy of a system. In the first instance there will be an addition of energy and in the second a subtraction. For simplicity this system can be represented as in figure 137, a plot of \((G)\) against time.

With the addition of thermal energy, two possible states of matter can exist, phases A and B. It can be seen, however, that the new \((Gm)\) \((Gm')\) for these phases is not reached by a direct route. Rather \((G)\) builds up to a maximum then falls to the new equilibrium position. Then, as more heat is added and all the matter has swapped to its new state, it continues to climb. There are two reasons for this. One, a small part, is caused because the rate of supply of thermal energy exceeds the rate at which
GENERAL PLOT OF GIBBS FREE ENERGY / TIME FOR A SIMPLE ROCK UNDERGOING METAMORPHISM (PROGRADE METAMORPHISM)

Point at which Thermal Energy began to be added to the system.
the matter can react to absorb it by shifting the equilibrium. At some point though the excess energy speeds up the reaction, causing the graph to dip rather than level out (a 10°C increase doubles the rate of a chemical reaction). The second part, and the more important, is caused by the fact that before a shift in the state of matter is possible, the original state has to be destroyed; this takes energy. The peak in the graph, therefore, represents the activation energy \((Q)\) required to do this.

The well-known Arrhenius equation expresses this and, in fact, is an expansion of the \((D)\) term in Frick's laws \((2)\) \((3)\).

\[
D = Kr = A
\]

where:

\(D\) = the diffusion coefficient

\(Kr\) = the rate constant

\(A\) = the pre-exponential constant (the vibration of matter because of its temperature)

\(R\) = the gas constant

\(T\) = temp. in °F

\(Q\) = the activation energy.

The term \((q)\) can be subdivided \(Q = W + E\), where \((E)\) = enthalpy to form a defect, and \((W)\) = enthalpy to move a defect.

To explain his equation, Arrhenius suggested that in every system an equilibrium existed between "normal" and "activated" molecules, and that only the latter could take part in chemical reactions. Heat, he suggested, increased the number of "activated" molecules, and therefore the rate at which the reaction progressed. Point \((B)\) would be the equilibrium position selected by this system because the \((Gm)\) value is less than for \((A)\). The attainment of the equilibrium position may involve an endothermic or exothermic reaction; in a situation of prograde thermal variation the former would be favoured.
Figure 138 shows the situation that develops when thermal energy is removed from the system. Here (G) never reaches the value (q), so the rate at which the state of matter A or B is attained is infinitely slow. On a temperature basis alone then, retrogressive reactions in geological situations should be impossible. To some extent this is true; if conditions seen in figure 138 did not hold, then at the surface only minerals in equilibrium with surface pressure and temperature conditions would be found.

The fact that we do see retrogression reactions in geology means that the Arrhenius equation as first envisaged is only a partial solution to the attainment of equilibria. This is because there are terms other than temperature to be taken into consideration. Still accepting that molecules need to be "activated" in order that the reaction proceed at a reasonable rate, the most important variable other than temperature is the state of strain. It is possible mechanically to push a molecule from "non-activated" to "activated" condition.

Let us examine the system as shown in figure 139. Here the thermal variation is the same as that for figure 138, but the energy is applied to the system in the form of mechanical stress. (We see immediately that the value of (q) remains the same, thus the term in the Arrhenius equation that changes or incorporates the stress function is (A).) Now the matter is given the requisite energy, the new state B is achieved and the rock will record the fact that (G) has passed that (Gm) value at some point in time. (Note: the rarer (Gm) states are likely to be utilised in these conditions.) A good example of just such a situation is that of the classical (Ramsay & Graham, 1970) shear zone, where the phases present inside and outside the zone show different (Gm) values, simply because of the strain in the zone of shear. This sort of metamorphism is therefore stress dependent. It should not be confused with retrogression resulting
Fig 138.

GENERAL PLOT OF GIBBS FREE ENERGY / TIME FOR

A SIMPLE ROCK UNDERGOING METAMORPHISM

[RETROGRADE METAMORPHISM]

Gibbs free energy of the system [G]

Activation Energy

Q

Gm(A)

A

Gm(B)

Original State

Point at which to be subtracted

Thermal Energy began from the system

TIME
Fig 139.

GENERAL PLOT OF GIBBS FREE ENERGY / TIME FOR
A SIMPLE ROCK UNDERGOING METAMORPHISM

[DEFORMATION ACCOMPANYING THERMAL RETROGRESSION]
from subsequent prograde (thermal) phases, which do not progress to the level of those previously seen in the rock.

Confining the arguments to the situation where a mineral recrystallises in situ or at a site remote in the rock, a further understanding of the effects of strain is possible. The generalised plot for such a situation is seen in figure 140(a). It is clear from this, that no matter what the mechanism, \( q \) for the mineral will remain constant, and that in this situation of recrystallisation chemical energies need not be considered.

Figure 140(b) shows the individual components of the system where the rock is stressed. For a fixed thermal value it is possible to increase \( g \) by the input of mechanical energy. If the product of the two exceeds \( q \) for the mineral it will recrystallise PROVIDED the ambient thermal energy exceeds a critical value. This value is unique to every mineral species. Thus in a polymineralic rock some minerals will recrystallise while others will persist in the strained condition if the rock is subject to constant strain input at a fixed temperature. (The potash feldspar in the Osterý mylonites is a good example of a recrystallising mineral while sphene remains deformed.) If, therefore, we know the absolute value of \( q \) for a given mineral and the limits of its thermal environment, then, by examining the state of recrystallisation, estimates as to the strain rate should ultimately be obtainable.

The Critical Thermal Value

It is possible to attain \( q \) virtually entirely by stress. In this case the return to the ground state will occur, but only over a long period. The gradient of the graph from \( q \) to \( (\mathrm{Gm}^*) \) is therefore temperature dependent. Below a certain critical temperature \( (\mathrm{Tm}) \), which varies from mineral to mineral, the rate of return to the ground state is infinitely slow.
Fig 140

**GENERAL PLOT OF GIBBS FREE ENERGY VS TIME FOR A MINERAL UNDERGOING RECRYSTALLISATION**

(a) Recrystallisation of the same phase of a mineral

(b) Strain energy portion of $G$

Thermal level remains constant
This means there is a threshold limit to the \((T)\) term in the Arrhenius equation for single phase recrystallisation when strain alone is responsible for producing the "activated" state. Figure 141(a) shows the progress of the system when above \((T_m)\). Figure 141(b) shows it when below \((T_m)\).

It is therefore possible to define three new critical thermal positions:

\((T_m)\) The "Mechanical" activation temperature, the point at which the entropy barrier caused by the rigidity of the crystal is overcome.

\((T_i)\) The "Initiation" activation temperature, the point at which one or more of the phases present dissociate and recombine to produce one or more new phases. \((T_i)\) is applied to the new phase.

\((T_t)\) The "Termination" activation temperature, the point at which a mineral phase ceases to be stable and dissociates. \((T_t)\) of one mineral is therefore \((T_i)\) of another. There are two such points for each mineral, where \((T_t) < (T_m) < (T_i) < (T_t)\). This is because, as has been demonstrated, before a mineral can be "destroyed" \((Q)\) must be reached. In the prograde situation this is achieved dominantly by thermal energy, in the retrogressive by mechanical energy. In each case the new mineral formed is different, reflecting the nature of the metamorphism. (Pressure, too, is a relatively minor consideration here.)

From the observations made on the rocks of the Southern Gneiss Unit, which show a retrogressive metamorphism, it is possible to rank minerals on the basis of their order of "closure" \((T_t)\) on the thermal decline:

- (1) Kyanite
- (2) Garnet
- (3) Hornblende
- (4) Actinolite
- (5) Epidote
- (6) Biotite
- (7) Sodic feldspar
- (8) Chlorite
- (9) Potassic feldspar
- (10) Quartz

In general minerals with a broad separation of their two \((T_t)\)s, i.e. stable over a wide range of P-T conditions, AND with a low \((T_m)\) are likely to
Fig 141.

GENERAL PLOTS OF GIBBS FREE ENERGY vs TIME FOR A STRAINED CRYSTAL IN TWO CONTRASTING THERMAL ENVIRONMENTS

(a) Effect of Stress on a system where strain is the only increasing variable, & the system is above Tm.

(b) Strain Energy [Stress] stops being applied

Effect of Stress on a system where strain is the only increasing variable & the system is below Tm.

Tm = Critical value for the Thermal component of Gt, [Gt] if recrystallisation is to proceed at a reasonable rate.
form the largest crystals in metamorphic terrains with a high tectonic component.

If the same minerals are now examined as to the degree of idiomorphy, then it is possible to rank them in order of decreasing idiomorphic tendency:

1. Rutile, sphene, magnetite
2. Tourmaline, kyanite, garnet
3. Epidote
4. Amphiboles
5. Micas, chlorite
6. Calcite
7. Plagioclase, quartz
8. Orthoclase, microcline.

This table, therefore, is a reflection of the energies developed at the grain boundaries as described above. (Note: all minerals will eventually develop the euhedral form given enough time above (Tt) and (Tm).)

It is also possible to observe how minerals of a given size resist stresses tending to reduce their size. They resist in two ways:

1. Directly: here the crystal structure has high energy bonds, poor cleavage, dense packing in the lattice and high symmetry (e.g. garnet), making the crystal difficult to deform as compared to its neighbours.
2. Passively: minerals can maintain their size and coherence by virtue of having a low (Tm). Stressing minerals below their (Tm) causes subgrain formation. Above (Tm), however, they anneal fast enough to maintain coherence, and in some cases maintain their original shape (e.g. potash feldspar).

At any one temperature the rate of grain growth decays with time as the surface area volume ratio increases. This means that for one strain rate there is an equilibrium grain size for every mineral. Two lists can be constructed grouping the same minerals in descending order of their ability to resist stress. The first deals with minerals that resist stress directly, the second with those which resist passively:
In general, high bond strength and good symmetry etc. will tend to prolong the life of a crystal as a clast less well than the annealing mechanism when strains become extreme. It is therefore the potash feldspar which forms the characteristic porphyroclasts in mylonites.

### 4.2.2 Processes in the Osterøy Mylonites

**Crystallographic Orientation Fabrics**

Crystallographic orientation fabrics are frequently described from mylonites. The Osterøy mylonites are no exception. The presence, strength and orientation of the fabric is dependent on, and relatable to, the dominant processes acting during the bulk deformation of the rock. Often these fabrics are accompanied by a strong shape fabric, visible on the mesoscopic and macroscopic scale. The development of these fabrics is schematically represented in figure 142; the likely process of formation and accompanying structure are tabulated in figure 143. As can be seen the simplest and commonest path is a straightforward rotation and growth of inequidimensional grains into the X/Y plane of the strain ellipsoid; frequently the longest axis becomes aligned parallel to X (Spry, 1969).

It is the crystallographic alignment of equidimensional minerals relative to macroscopic features that presents problems. These alignments are not necessarily always in the X/Y plane, but may form an orthorhombic

<table>
<thead>
<tr>
<th>(1)</th>
<th>(2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Potash feldspar</td>
<td>Soda feldspar</td>
</tr>
<tr>
<td>Garnet</td>
<td>Quartz</td>
</tr>
<tr>
<td>Sphene</td>
<td>Amphibole</td>
</tr>
<tr>
<td>Amphibole</td>
<td>Kyanite</td>
</tr>
<tr>
<td>Amphibole</td>
<td>Muscovite</td>
</tr>
<tr>
<td>Amphibole</td>
<td>Biotite</td>
</tr>
<tr>
<td>Amphibole</td>
<td>Chlorite</td>
</tr>
</tbody>
</table>

(The numerics rank them in order of ease of annealing.)
SCHEMATIC REPRESENTATION OF PATHWAYS BY WHICH CRYSTALLOGRAPHIC FABRICS MAY DEVELOP

1) MECHANICAL ROTATION OF INEQUIDIMENSIONAL CRYSTALS

2) INTERNAL PLASTIC DEFORMATION OF EACH CRYSTAL

3) SLIDING OF CRYSTALS & SUBGRAINS WITH LITTLE INTERNAL DEFORMATION

4) SOLUTION AND GROWTH OF CERTAIN GRAIN ORIENTATIONS

5) GROWTH OF PREFERENTIALLY ORIENTATED GRAINS DURING METAMORPHISM

6) ENLARGEMENT & RECRYSTALLIZATION UNDER STRESS
### Table of Likely Deformation Mechanisms Involved in the Pathways of Figure 142 and the Nature of the Fabrics Produced

<table>
<thead>
<tr>
<th>Mechanism(s)</th>
<th>Fabric</th>
</tr>
</thead>
<tbody>
<tr>
<td>1) Grain Boundary Sliding (+ other processes)</td>
<td>Good Shape Fabric</td>
</tr>
<tr>
<td></td>
<td>Poor Crystallographic Fabric.</td>
</tr>
<tr>
<td>2) Glide &amp; Creep + Grain Boundary Sliding or Nabbrro Herring Creep or Coble Creep</td>
<td>Very Good Crystallographic Fabric</td>
</tr>
<tr>
<td></td>
<td>if Glide &amp; Creep were the Mechanisms</td>
</tr>
<tr>
<td></td>
<td>but NOT if GBS is important too.</td>
</tr>
<tr>
<td></td>
<td>OTHERWISE, a Good Shape Fabric.</td>
</tr>
<tr>
<td>3) Grain Boundary Sliding (G.B.S.)</td>
<td>Only a weak Mesoscopic Shape Fabric.</td>
</tr>
<tr>
<td>4) Pressure Solution Coble Creep [+G.B.S.][+Glide &amp; Creep]</td>
<td>Good Shape Fabric &amp; a weak to moderate Crystallographic Fabric, depending on the amount of G.B.S. which may have occurred too.</td>
</tr>
<tr>
<td>5) Chemical Changes</td>
<td>Good Shape Fabric &amp; with respect to the new mineral, probably a moderate Crystallographic Fabric.</td>
</tr>
<tr>
<td>6) Glide &amp; Creep + Coble processes + Grain Boundary Sliding</td>
<td>Very Good Shape Fabric &amp; possibly a very weak Crystallographic Fabric.</td>
</tr>
</tbody>
</table>
monoclinic or triclinic symmetry about one strain axis. Current theories explaining these fabrics are based on the work by March (1932) and subsequently Means and Patterson (1966). The process is essentially one of slip on suitably orientated planes of weakness in the lattices, causing rotation of the lattice without bodily rotation of the crystal. The process therefore is only possible when the rock is deforming dominantly in the area of the Deformation Map where climb and glide are the dominant processes.

The only other field of the Deformation Map where crystallographic orientation fabrics are likely to form, is that of pressure solution. It has been demonstrated above that anisotropic mineral grains have faces with varying surface energies. Following the principles described, crystals in the process of dissolving, nucleating and growing, which tend to be orientated so that the faces with the lowest energies will be normal to the maximum principal stress axis, are favoured. Crystals not so orientated will either dissolve faster or grow slower; the bulk of the crystals will therefore show a preferred alignment.

The Osterøy mylonites are different, as has been shown, in that only certain rock types have crystallographic orientation fabrics and as the only difference between the specimens is their mineralogy, it must be this that is responsible for their differing fabrics.

These observations shed light on the nature of the deformation of the mylonites. It is clear that processes other than those here described do not produce fabrics because the deformation paths for the rest are independent of crystal structure. Rock type, too, has an effect on fabric production, the homogeneity of monomineralic rocks regularises and restricts the deformation process, enhancing the fabric.
Thus fabrics form:

(a) In conditions when glide, climb, pressure solution, coble creep and crystallisation (twinning too to a certain extent), are the dominant deformational processes, i.e. no other process acting in conjunction must have proceeded at a rate which would negate or reverse the tendency towards fabric development.

(b) When the rock is composed of very few mineral species.

They do not form:

(a) When grain boundary sliding is a process of deformation.

(b) During deformations in which there have been rapid changes in the grain size of some grains, the local volume adjustment causes rotations of neighbouring minerals.

(c) Where cataclasis or disaggregation of subgrains occurs, as the fragments tend to rotate.

(d) If recrystallisation, nucleation, or solution processes have proceeded at a rate fast enough to overcome surface energy anisotropies.

(e) If the stress field is short lived or rotating.

(f) If the rock is polymineralic in composition because each mineral is likely to deform by a different process. (Even if these processes acting alone would produce fabrics, they negate each other acting together.)

Polymineralic rocks are mechanically heterogeneous at the microscopic level; they are therefore unlikely to fulfil any conditions except those preventing fabric formation.

Thus in the quartzites on Osterfy, the deformational processes are clearly defined (solution, climb and glide). In the granite gneisses, the processes are also clearly defined (grain boundary sliding; the result of mechanical anisotropies, enhanced by volume change effects as the mineralogy responded to the metamorphism).
These two rocks tell us that all processes (except perhaps nabbro herring creep) were operative simultaneously in the Osterfjord gneisses during mylonitisation.

Changes in Bulk Chemistry

There is little published work on the geochemical changes accompanying mylonite genesis. Field evidence from known mylonite examples (e.g. the Moine thrust) clearly shows that as a rule, the rock product of intense mylonitisation is greatly enriched in quartz as compared to the parent rock. On a smaller scale there is evidence for the local chemical mobility in the production of the characteristic banding (Sclar, 1965), though there are some who believe this is a deformational feature (Myres, 1977; Vernon, 1974). These observations may imply that there is some chemical mobility during mylonite production, on both the local and the "open system" scale. Work on shear zones by Beach (1976), where it was found that sodium and potassium were extremely mobile, supports this view. Beach also found that the shear zones were enriched in water, but he found no evidence for silica enrichment or depletion.

The pilot Enrichment/Depletion plots of the Osterfjord augen gneiss however, proves mylonitisation of these gneisses was probably chemically an "open system" process, at least with respect to the size of the outcrops sampled. The deformation was accompanied by loss of mafics or femics, aluminium and sodium from the rocks. This proves that there were volume changes during mylonitisation and that surface active processes were operative.

Processes responsible for Mineral Textures

An understanding of the processes operative during the formation of the Osterfjord mylonites is only possible after relating the textures observed to theoretical considerations and the deformation map. This should provide
information as to the significance of the mylonites as a unit in the context of their environment of formation.

Textures produced by Chemical Processes

This group of textures is the result of metamorphic changes, i.e. (Tt) and (Tt) effects. The changes often require mobility of ions in the rock for the reactions to progress. These ions are best transferred in solution; coble creep and pressure solution must therefore have been active during these changes. It is also reasonable to postulate that grain boundary sliding and mineral rotation would occur in the rock in response to local volume changes.

Rim Symplectites (Myrmekite)

Much is written as to the causes of this texture (Phillips, 1974). In the Osterøy mylonites the structure is a product of the retrogression, and is chemically induced, because it works from the outside inwards. The reaction is not a solid/solid process as the texture forms regardless of the adjacent mineral species. This means the instability is the product of an external vector. An intergranular fluid must therefore be present, indicating that coble creep and pressure solution were active. The morphology of the texture indicates that at the time of formation the surface energies of quartz and sodic plagioclase were complementary. (The intergrowth as seen in these rocks is not actually worms but sinuous branching sheets of quartz in the plagioclase. This can be seen because no matter what the section through the intergrowth, it never shows a circular section.)

Solution features

Pressure solution is almost entirely responsible for these; the solution occurs at the points of highest stress. The process, while common, is difficult to detect except in the case where a marker is present,
as in plate 55. There must always be grain boundary sliding accompanying this process to accommodate the volume loss.

Relic structures

These are the product of metasomatic metamorphic reactions; chemicals therefore need to be mobile through the body of the rock. The transport of ions to and from the waxing and waning crystals must be via an intergranular fluid; solid state ionic diffusion cannot alone explain the texture. With the presence of an active intergranular fluid it is reasonable to assume pressure solution is occurring elsewhere in the rock. The feature, which characteristically shows little marked stress orientation, also indicates there could have been little grain boundary sliding during formation. Indeed the local stress field during formation must have been virtually isotropic; the feature therefore is the product of a pause in tectonic development.

Clearing and Granoblastic texture

These are the product of annealing processes on strained rock. Granoblastic texture represents the most stable configuration for the minerals composing the rock. Internally each crystal approaches the strain-free state (clearing) and develops equilibrium configurations with its neighbours.

Given specific P T conditions each mineral present will attain a constant size and shape according to the dictates of surface energy and (Ti), (Tt) factors, provided no two minerals of the same species share a common boundary. The absolute value of this grain size is dictated by the extent to which the minerals have been ground in the first instance. Once all the grains of one mineral in a rock are of equal size, the surface energy ceases to be an important factor. A granoblastic texture can therefore develop and be stable at all possible grain sizes. This means
that once a finer grained rock is generated from a coarse grained rock, in the absence of subsequent P T and tectonic changes, it is difficult to re-establish the original grain size. The fine grained state is an equilibrium condition; it can therefore develop during deformation and is not only a post-deformational texture.

The processes operating are therefore two-fold:
Those acting at grain boundaries to relieve fossil stresses; grain and sub-grain boundary migration, coble creep, solution, crystallisation (pressure solution) and a little grain boundary sliding.
Those acting through the body of the mineral allowing re-ordering of the lattice; glide and creep.

Crenulate Boundaries

These are seen in rocks with an essentially monomineralic composition. They are features very like those seen in granoblastic texture, but here two minerals of the same species share common boundaries. The ultimate grain size is therefore a function of time and temperature. The irregularity of the boundaries reflects the relative energies developed at faces disposed at varying angles; the constant enlargement of the grains results in jigsaw-like junctions.

Processes operating in the formation of this structure are dislocation glide and creep causing grain and subgrain boundary migrations. The process is driven by surface energy forces, thus the structure may form during deformation or recovery.

Snowball Garnets

These are important structures, the mineral is obviously stable in high strain rate conditions. The growth pattern (Powell & MacQueen,1976) indicates that they nucleate and grow during deformation. This requires there to be a mobility of chemicals, and points to a fluid phase present during deformation.
The processes likely to be operating, as indicated by the structures preserved in this mineral are: grain boundary sliding, coble creep and pressure solution. As the mineral maintains coherence during rotation glide and creep must have been operative too.

**Poikiloblastic Texture**

The reason for the existence of inclusions in the body of a developing crystal is basically a surface energy phenomenon. The presence of the inclusions cannot unduly increase the surface energy equilibrium of the crystal. The inclusion must then satisfy the free bonds at the boundary between the two grains (a phenomenon akin to the "wetting" properties of liquids). The species of the inclusion, therefore, provide an insight into the properties of the including mineral. The usual inclusion in the Osterøy mylonites is quartz, though graphite can also be an inclusion in garnet.

**Amoeboid Growth**

This growth form is the result of regrowth of a crystal that has been mechanically dispersed into the matrix during deformation. On recrystallisation, the centre of nucleation approximates to the central portion of the original crystal, and moves out from this joining up the scattered pieces. Thus portions now separated by screens of matrix material produce pseudopodia-like extensions from the centre of nucleation. It is clear that these pseudopodia are the product of the migration of chemicals either in solution or by solid state diffusion around grain boundaries. Coble creep and pressure solution are therefore indicated, though the texture is mainly a chemical (Ti) product.

**Porphyroblasts**

These structures post-date the fabric of the rock in which they grow. In polymetamorphic terrains this relationship may be masked by
subsequent deformations. There are situations where the growth is syn-
tectonic, whereupon the crystal displays both clastic and blastic features
simultaneously. The processes are in the first instance nucleation of
the mineral at a new site in the rock, a (Ti) effect requiring the
presence of a fluid medium (coble creep and pressure solution could there­
fore have been active). In the second instance, where the deformation
continued during blastesis, glide, creep and grain boundary sliding must
also be operative. The feldspar porphyroblasts/clasts in the Osterfy
mylonites are of this second type. This means textural patterns developed
at the margins of these structures (Sinha Roy, 1977 a ; Ferguson & Harte,
1975) must be treated with extreme caution because they do not represent
relics of the original rock.

Embayments

These features are produced in a manner similar to relic structures,
a purely chemical process requiring transport of ions. Coble creep is
therefore operative and if there is a significant fluid phase, pressure
solution takes over.

Rims and Crystallisation

These are metamorphic features; the mobility of ions required
in their formation indicates there was a fluid phase present. Pressure
solution and coble creep were therefore active, as was grain boundary
sliding.

Recrystallisation

This term covers all processes by which a mineral regains its
original structure. The processes involved are: dislocation glide and
creep; these are responsible for re-ordering the lattice and enlargement
of the crystal by annexation of neighbouring grains. Coble creep and
pressure solution; the mechanisms by which material is obtained from
non-adjacent grains. Implicit in any superplastic model explaining mylonites is that clearing and recrystallisation (annealing) operate during deformation, and not just in the post-tectonic period.

There is considerable controversy about the role of recrystallisation in the development of mylonitic banding (Vernon, 1974; Myers, 1978; Sclar, 1965). From the Osterfy mylonites it is clear that the banding results from shearing of once restricted pockets of mafic material into long streaks parallel to the foliation. The banding is then enhanced by migration of leucosome from areas rich in mafic fragments to the less mafic regions; selvedges develop. The scale of migration implied by the banding seen in the mylonites indicates the importance of the recrystallisation processes.

The movement of material from the mafic-rich regions happens because the energy for a mafic/leucocratic mineral grain boundary is less than that for a leucocratic/leucocratic mineral boundary. Rocks banded on the mesoscopic scale cannot then obey the granoblastic rule, the surface energies never equilibriate locally. It is by this process, too, that quartz ribbons form.

From these observations it is apparent that chemical effects, phase changes, metamorphic assembly changes (T1/Tt effects) and the chemical transport component of recrystallisation, are extremely important when accommodating deformation. This view is supported by the recent work of White and Knipe (1978).

Mechanically formed Textures

Textures in this group are the product mainly of processes acting within the body of crystals. The tendency, contrary to most of the chemical processes, is always towards a diminution of grain size. Except at very high strain rates and pore fluid pressures, the grain size reduction
and shape change is achieved without the grain losing internal coherence. There is usually a substantial amount of grain boundary sliding and coble creep accompanying these shape changes. These processes continue after the internal processes stop.

**Cataclasis**

This is one of the high strain rate products; the grains lose internal coherence. Dislocations are generated in the lattice at a rate faster than glide and creep can dissipate them. The whole process of dislocation motion becomes more difficult as the dislocations get "tangled up"; this is work hardening. If stressing continues, brittle fracture occurs and the lattice fragments. These fragments are then strung out into the plane of the foliation; this is done by grain boundary sliding. There is another method for producing cataclasis in a rock, namely by high pore fluid pressures. The process is slightly different: the crystal is not work hardened before rupture, rather local pressure differences within and between crystals are responsible. These develop in areas where pore fluids escape from the rock; the rock and crystals are literally exploded apart. In the Osterøy mylonites, cataclasis of this last variety only occurs at a late stage in the tectonic history and produces the D8 mylonites. At this time the migratory fluids were rich in quartz, epidote and calcite, thus by inference pressure solution must also have been active somewhere in the rock.

Reduction of grain size by these methods is not a steady state process; cataclasis therefore has no place on the deformation map.

**Undulose extinction**

As dislocations build up in a lattice, distortions occur. Viewed optically these are seen as variations in extinction position in the crystal. The processes, therefore, are: defect formation and then migration by dislocation glide and creep.
Subgrains

Following on from undulose extinction, dislocation glide and creep carry defects through the lattice. In some areas bottlenecks occur; these define sub-areas in the original crystal. The subgrain boundaries, sites of high disorder, facilitate dislocation migration. Coble creep is therefore active in these regions. This process of grain size reduction differs from cataclasis, as a steady state is achieved. The diameter of the subgrains decreases (numbers of them increase) to a point where coble creep round the boundaries can dissipate the stress without initiation of new dislocations within the crystal.

Quartz Lamellae

These are subgrains and bands of undulose extinction that form in quartz. There is a definite alignment relative to the causative stress field. The processes involved are identical to those for undulose extinction and subgrain formation.

Augen and Porphyroclasts

In the Osterøy mylonites these can be divided into three types:

1. Those which are seen to be unaffected by the stresses which have affected the matrix.
2. Those in which the outer margin of the clast is strained like the matrix.
3. Those in which there is extreme internal deformation.

In the case of varieties 2 and 3, there are further complications. The minerals can recrystallise concurrently with deformation, thereby obscuring strain that is seen elsewhere in the rock.

Type 1 porphyroclasts can only form from minerals with a "rigid" lattice stable at high strain rates, i.e. with little or no mechanical anisotropies and high bond strengths. The matrix flows round these crystals
by grain boundary sliding. The process for type 2 is grain boundary sliding too, but the crystal margins friction causes dislocation build-up and attrition occurs by glide and creep process. The subgrains are stripped off and there is no over-all change in axial ratio for the clast. Porphyroclasts formed by this method are unsuitable as markers for studies on the relative variation in strain ellipse across an area.

Type 3 clasts develop in the same way as the others, but glide and creep become more important, producing subgrains within the clast. The clast, therefore, changes shape during deformation. Once the aggregate of subgrains is formed, this then changes shape rapidly as the constituents rotate by grain boundary sliding. The clast becomes flattened into the plane of the foliation, but still remains an entity.

The porphyroclasts from Osterøy are of all three types; the commonest are made from potash feldspar which forms types 2 and 3 clasts. Type 1 clasts are best developed from garnets.

Shear Bands

The process of initiation of shear bands is important when discussing mylonites. In the Osterøy rocks, not as proposed by Sinha Roy (1977), it is clear that most of the shear zones that are distinctly seen in the rock are a relatively late phenomenon in the history of deformation.

All processes of grain boundary sliding involve simple shear at the grain boundaries. Thus, no matter the nature of the macroscopic deformation, it can be resolved in part into simple shear at the microscopic level. When the shear is between two grains of markedly differing mechanical properties, the "weaker" member (e.g. mica) will break up faster. Locally, therefore, the grain size of both components will be smaller as this region of the rock has acted to magnify the stress field. This grain size variation is obvious from the Osterøy mylonites. As has been
discussed above, reduction in grain size accelerates chemical effects. It also provides more boundaries for coble creep, grain boundary sliding etc. to work on. Once shearing of this sort starts on a microscopic level, then it is a runaway process, and a large shear band develops. It is the variation in mineralogy that is the primary reason for the initiation of a shear band at any particular locus. When the confining pressure is high the process happens at all such boundaries in the rock (sometimes a body of rock can behave like a crystal - "tectonic pips"), it is only as the confining pressure decays that the process becomes heterogeneous. The main processes then are the grain boundary processes; these start and continue after mechanical constraints have stopped glide and creep. Of the grain boundary processes, it is grain boundary sliding that is most important. All the boundary processes are accelerated by the presence of a fluid phase (Westwood et al., 1967; Rutter, 1972(b)), a Rebinder effect. The chemical processes prove that there was a fluid phase throughout the deformation.

4.2.3. **Formation of the Osterøy Mylonites**

The characteristics of the mylonites can be summarised:

(i) They are regional in development involving vast thicknesses (> 3.5Km), of both allochthon and basement.

(ii) The internal structures and fabrics indicate superplastic behaviour during their formation.

(iii) All the processes on the deformation map were operative during deformation. (It is assumed that nabbro herring creep was operative too, but never contributed significantly to any single textural development.) In addition chemical changes, both types of cataclasism, and grain boundary sliding were operative.
(iv) 99 per cent. of the mylonites are polymineralic rocks with at least four major mineral phases present. This means the rocks were strongly anisotropic, both mechanically and chemically, on the microscopic and mesoscopic scale.

(v) A fluid phase was present throughout the deformation, resulting in the acceleration of all surface operative processes. This reduced the mechanical strength of the rock significantly and also aided chemical movement.

With these conclusions it is possible to build a comprehensive model describing the nature of superplastic deformation as seen in the Osterøy mylonites. At the onset of deformation, the waxing stress field caused a build-up of dislocation density in the component crystals of the rock. The process proceeded, subgrains formed, and when the grain size reached the point where most of the deformation was accommodated by grain boundary processes (some coble creep but mostly grain boundary sliding, as is proven by observation of deformation structures in mafic areas and the porphyroclasts), a stable grain size was reached. Dislocations still formed within grains, but the grains were small enough, and surface processes active enough, for them to be few enough in number to migrate through the lattices without locking up. Then for a given mineral and strain rate there would be a unique stable grain size. Thus there is an equilibrium grain size state for any particular rock type and strain rate and the first response to stress is the generation of this state. Deformation then proceeded mainly by grain boundary sliding. This and other surface processes were accelerated by the presence of the fluid phase in the rock.

Part of the stress was stored in the rock when generating this fine-grained state, partly as "activated" molecules and partly as surface area energies. This then facilitated chemical reactions and recrystallisation, both of which provided "fresh" crystals to be deformed. This, too, accelerated the rate at which the stress was dissipated.
Superplasticity then is not a single process, but the utilisation of all possible processes, especially the surface active varieties, coble creep, pressure solution, chemical processes and grain boundary sliding. The most important factor for these surface processes is the grain boundary area in the rock. The amount of surface area is controlled by the activity of the internal processes, glide and creep. This means that superplasticity is a state of dynamic equilibrium between constructive and destructive processes, acting to dissipate the stresses.

The model can be expressed as a cybernetic loop relating: grain size, finite strain, strain rate and stress field (fig.144). This means we can consider two geological conditions. In both cases the rocks are of the same composition and in the same geological environment. The second body, however, has in it a zone in which the grain size is smaller than for the rest of the rock. On deformation the first body of rock will enter the loop, and as deformation proceeds will tend to deform homogeneously. The strain is accommodated by all parts of the rock, and the grain size achieves equilibrium evenly throughout the rock.

In the second instance, the presence of the inhomogeneity of grain size in one portion means that that portion starts deformation in a more "advanced" state than the rest of the rock. The rate of deformation in this area will therefore be high. This modifies the stress field which means the rest of the rock will remain undeformed, and all the deformation will be accommodated in the finger-grained portion. Should the rate of application of energy increase, two things can happen: either the grain size in the zone of shear can decrease to facilitate the operation of the surface active processes, and/or the volume of rock in the zone of shear can be increased by attrition from the margins. This, too, increases the total grain surface area present in the zone of shear. In reality both are likely to happen, the attrition process occurring first until the grain
Fig 144.

THE STRAIN-RATE TEMPERATURE SYSTEM FOR THE
GENERATION OF THE OSTERøy MYLONITES

---

PRODUCES - CONTROLS - CONTRIBUTES TO
size in the zone has a chance to equilibrate. Then the only factor
tending to enlarge the zone will be the frictional resistance to sliding,
which will only cause a small amount of attrition.

In reality, rock is never homogeneous. When rocks deform super-
plastically they do so forming shear bands at the points of maximum
anisotropy (see previous description for modes of occurrence of the Osterýy
mylonites). The dynamic approach of the superplastic model explains why
restricted zones of shear develop and perpetuate; why sites of previous
shear are so frequently reactivated by later deformations rather than new
zones developing; how tectonic pips or pods of undeformed rock are created
and why, once formed, they persist unaffected by subsequent deformations.
Indeed all the above mentioned features are indicative of a superplastic
flow regime, where fine-grained rock acts as a "Strain Mop".

The relationship between differential stress, temperature and grain
size (the area defined by the stippled box in figure 144), can also be
described as shown in figure 145. The abscissa is the time axis, the
ordinate the grain size axis with the origin representing the original
grain size. All values below this represent reduction in grain size and
vice versa. The mechanism influencing grain size through time can be
described as a tensor with two major components, one of which can be
represented as \( f(\frac{de}{dt}) \), the variation of strain rate with time, and the
other as \( f(\frac{dT}{dt}) \), the variation of temperature with time. The solution
of this tensor determines the slope of the line for the variation of grain
size with time. A positive slope indicates grain growth, a negative slope
grain size reduction. When the gradient is vertical the rock has either
melted or fractured. A horizontal path indicates a steady state condition,
grain growth = grain size reduction. If the path is horizontal at a time
of deformation in the rock, the deformation process is likely to be
superplastic.
Fig 145.

SEPARATION OF METAMORPHIC ENVIRONMENTS ON A PLOT OF VARIATION IN GRAIN SIZE THROUGH TIME

Gradient of graphs = f(t).

\[ \begin{align*}
\text{\textbullet} &= \text{Grain growth exceeds Subgrain formation} \\
\text{-} &= \text{Grain growth equals Subgrain formation} \\
\text{\textbullet} &= \text{Subgrain formation exceeds Grain growth}
\end{align*} \]

Temperature increasing at constant T

STOCK STATE CONDITION: NO CHANGE IN CRYSTAL SIZE, THERMAL & DEFORMATIONAL EVENTS BALANCE OUT

A = Thermal metamorphism - [Melting]
B = Regional metamorphism
C = Osterøy mylonite type metamorphism
D = Traditional fault zone mylonite type metamorphism
E = Faulting
Examination of the paths for real rocks in this field enables different patterns to be distinguished which characterise the different possible metamorphic environments.

4.2.4 Implications of the Osterøy Mylonites

The Osterøy mylonites, therefore, describe a tectono/metamorphic system lying on a plane, defined by strain rate \((e')\) and temperature, normal to the pressure axis of a diagram relating time \((t)\), \((e')\), temperature \((T)\) and pressure \((P)\). This diagram in fact describes completely the passage of rocks (metamorphic and structural history), through time (as opposed to the classical plots such as the PT diagram or the Deformation Map, which only describe one plane cut through it).

The three cartesian axes of this diagram are simplifications of parameters in an equation that defines the finite state of rock, its chemical composition and mineralogy as a function of what these variables were originally and how they have changed with time.

\[
E_{GP} = f(d f(E_0. C_0. P_0.)/dt)
\]

where: \(E\) = finite strain \(E_0\) = original strain state
\(C\) = finite chemistry \(C_0\) = original chemistry
\(P_h\) = finite mineral phases \(P_h_0\) = original mineral phases.
\(f\) = some function

This equation can be expanded:

\[
E_{GP} = f(d f(f(dC/dC_0).f(dP_h/dP_h_0).f(dT/dT_0).f(dP/dP_0).f(d(d(C_1-C_3)/dt)c))/dt)
\]

where: \(C_c\) = the change in chemistry
\(P_h_c\) = the change in mineral phases
\(T_c\) = the change in temperature
\(P_c\) = the change in pressure
\(C_1\) = the maximum principal stress
\(C_3\) = the minimum principal stress
\((d(C_1-C_3)/dt)_0\) = the original strain rate
\((d(C_1-C_3)/dt)c\) = the change in strain rate.

This can be more simply written as:

\[
E_{GP} = f(d(f(C).f(P_h).f(T).f(P).f(d(C_1-C_3)/dt))/dt)
\]
It is clear that this expression requires a large number of terms for its solution, in fact the number approaches infinity when all possible mineral phases are considered. The equation would then be useless to the geologist, but it can be simplified. We know from field observations that when the strain rate becomes excessively high, the rock fractures, and that at slightly below this rate the chemistry alters by removal of all material except quartz. We know that above a certain temperature rock melts, and we only need to consider geologically realistic pressures. When considering the chemistry and mineral phases present, there are only a finite number that are of geological importance.

Real geology, therefore, only occupies a small field within that space theoretically possible. Thus considering one moment in time it is possible to write an equilibrium reaction to explain what is happening in a rock:

\[(P_{eq} \cdot T_{eq} \cdot e'^{eq}) = G(P_{(to)} \cdot T_{(to)} \cdot e'(to) \cdot A \cdot B)\]

where:

- \(t_{o}\) = time at the start
- \(e\) = time at equilibrium
- \(G\) = a function
- \(A\) = a function of the original chemistry at time zero
- \(B\) = a function of the original phases present at time zero.

This equation has a solution for the geologist; this can be plotted on a three-dimensional diagram constructed by adding a strain rate axis (similar to that shown in figure 132) to the classical PT diagram (fig.146).

The equilibrium positions for the rock at given times can be joined up. The resulting line represents the behaviour of the rock with time, the history of metamorphism and deformation as described by:

\[dEPh/dt = f ( f(Ph) \cdot f(T) \cdot f(P) \cdot f(e') )\]

This equation requires only twelve terms in its solution if the phases present are known and the chemistry constant.
Fig 146.

THREE DIMENSIONAL STRAIN-RATE / PRESSURE / TEMPERATURE DIAGRAM, ON WHICH METAMORPHIC ROCKS ARE MORE ACCURATELY PLOTTED.

VOLUME, in which metamorphic rocks actually fall

PT. PLANE of classical metamorphic petrology
The Use of the $e^*$ P T Diagram

The classical PT approach to metamorphic problems is useful and in general an adequate method for describing rocks. This approach has, however, virtually reached its limits with works such as Powell (1978) and Frazer (1977), and there are still inexplicable anomalies between theoretical and geological situations using this diagram and method (Hollister, 1969). It does not take account of the texture in metamorphic rocks either.

There is therefore a need to be able to describe rocks in a more complete manner. This need is highlighted by the Osterøy mylonites, which define a plane at right angles to that of the classical PT diagram, while classical thrust mylonites and pseudotachylites like those described by Sibson (1975, 1977) fall on a plane at right angles to both of these. It is clear, therefore, that any metamorphic event with a tectonic component will cause the rock to describe a path in the body of the cube defined by the $e^*$, P, T axes. To project this path onto any one plane is deliberately to ignore an important part of the history of the rock. The $e^*$ P T diagram, then, is the marriage of structural and metamorphic geology.

It is perhaps best to consider the effects of a third axis to the PT diagram by describing the effect it might have on one of the better known reactions, namely that of the aluminium silicate polymorphs.

The phase boundaries of this system will, with the addition of the $e^*$ axis, describe a three-dimensional geometric form in the diagram. There are problems in determining the true form of this figure, as no specific experiments have been run to determine its shape as we move up the $e^*$ axis. Fortuitously, many of the experiments that have been performed included significant proportions of effects that are attributable to $e^*$. Indeed, authors have been aware for a long time that these effects
significantly alter the positions of the phase boundaries, and put much effort into eliminating them.

Brown and Fyfe (1971) state: "Near equilibrium the $\Delta s$ of the andalusite-sillimanite reaction is only about 0.3 cal mol$^{-1}$ deg$^{-1}$. This means that if we wish to know the phase boundary with a certainty of at least $\pm 50^\circ C$ we must also use procedures which will give a free energy with an accuracy of $\pm 15$ cal mol$^{-1}$. For myanite reactions where $\Delta s$ is about 2 cal mol$^{-1}$ deg$^{-1}$, again $\Delta G$ must be known within the limits of at least $\pm 100$ cal mol$^{-1}$. These are trivial quantities when it is noted that the lattice energies of the three polymorphs are near 5 $10^5$ cal mol$^{-1}$.

It is clear when these requirements are considered, that small effects, strain energy, surface energy, impurity factors, will be critical in such systems. If materials are ground to micron size range, surface energies alone will be of the order of 100 cal mol$^{-1}$, and it is equally certain that strain energies induced in grinding to such a size will greatly exceed the surface energies. Thus it seems that the only experiments likely to yield significant results will be those in which unstrained macrocrystalline materials are used".

Similarly, Holdaway (1971) stated: "Intense grinding may shift phase boundaries. Newton (1969) has shown that prolonged grinding of a kyanite-sillimanite mixture preferentially degrades the kyanite as deduced by X-ray line broadening. The result of such grinding could be to give a falsely high pressure indication of the kyanite sillimanite phase boundary. Newton's work indicates that a boundary 2kb above the equilibrium boundary could result if grinding procedures similar to those of Althous (1967) are used. Shearing squeezer results of Bell (1963) may also be in error for this reason." "Fibrolite in sillimanite mineral samples has an effect on the andalusite-sillimanite boundary. The boundary is raised by as
much as 200°C for runs of short duration." Similar statements are made by Fyfe (1969).

When the field occurrence of the aluminium silicate polymorphs are examined, the literature is in the same confusion; apparent anomalies develop where one phase behaves metastably and there are arguments as to whether all of the polymorphs seen to be present in the rock formed at the same time, or were the result of two phases in the metamorphic event.

It is possible that the introduction of the $e'$ axis returns an order to this confusion in both the natural and experimental situations. It should be recognised that the variation in the positions of the phase boundaries as described perhaps present a truer picture of geological situations than that presented by the positions of the boundaries actually on the base plane. Similarly, in the field apparent anomalies such as those described by Hollister (1969), who detected two distinct trends in the aluminium silicate polymorphs in what seemed to be a single phase event. In one trend kyanite pseudomorphed andalusite and was then replaced by sillimanite. In the other andalusite was replaced directly by fibrolite, or first replaced by kyanite then rimmed by fibrolite. To explain this he proposed a complex series of metastable paths. However, an examination of his descriptions and photographic plates of the two sorts of rock show clearly that those rocks in which the second trend was seen are sheared, while the remainder are not. Thus on the basis of the evidence presented above it is suggested that a more satisfactory explanation would be to state that the two trends represent identical PT conditions for both, but that the rocks of the second trend progressed along a vector with a larger component of $e'$ than did the other. It thus moved more into the body of the $e'$ PT cube and into an equilibrium situation that will be described below.
This field evidence, together with the experimental work, allows some very general conclusions to be drawn as to the shape of the polymorph diagram in three dimensions.

1. The kyanite - sillimanite boundary plane curves towards the $P_e^*$ plane with an increasing component of $e'$. 

2. The andalusite - sillimanite boundary plane curves away from the $e'$ axis with increasing component of $e'$. 

3. The triple point moves firstly towards the origin then curves away from it again, describing a spiral towards the $P_e^*$ plane with increasing $e'$. 

4. The kyanite - andalusite boundary plane seems to move towards the $e'T$ plane with increasing $e'$. 

5. A new phase, fibrolite, develops some way up the $e'$ axis. The phase comes in by branching the upward growth of both the kyanite - sillimanite and andalusite - sillimanite boundary planes. There therefore exists the possibility of a four-point junction at some specific $e'$. 

Figure 147 embodies these ideas, and shows how Hollister's (1969) trends may plot on the diagram. The immediate objection to this proposal is that fibrolite is commonly described from the contact aureoles of granites which are believed to be passive intrusions. A set of teaching thin sections across one of these intrusions was studied. The FANAD Granite Pluton, outcropping on the three adjacent peninsulas of Rosynoll, Fanad and Inshowen in northern Donegal is a classical passive stoping intrusion (Pitcher & Berger, 1972), yet it immediately became clear that the minerals developed in the aureole are deformed. Andalusite, the first aluminium silicate to form, shows rotational inclusion trails (fig. 148) and fibrolite, which replaces andalusite nearer the granite is deformed round andalusite nuclei, forming augen structure; some of the andalusite cores are even bent.
Fig 147.

POSSIBLE STRAIN-RATE | PRESSURE | TEMPERATURE
DIAGRAM FOR THE ALUMINUM-SILICATE
POLYMORPH SYSTEM

a) ANDALUSITE → KYANITE → SILLIMANITE
b) ANDALUSITE → KYANITE → FIBROLITE
c) ANDALUSITE → FIBROLITE
Fig 148.

TEXTURES IN ALUMINIUM SILICATE POLYMORPHS
FROM THE AUREOLE OF THE FANAD GRANITE

ANDALUSITES SHOWING KINKED OR ROTATIONAL FABRICS

small crystal nucleating on a kink

small crystal keying a kink

andalusite

muscovite

rotational inclusion trails in a larger crystal

ANDALUSITE WITH FIBROLITE 'SHEATH'

fibrolite bends

late biotite

fibrolite 'sheath' forming from the andalusite

note augen shape

andalusite

fibrolite

structure deformed during or after the fibrolite development

0 1 2mm
Admittedly these features do not amount to very high strains, but the rocks are deformed and it is not the amount of deformation but the strain rate that is important.

Another possible stress stable mineral may be chrysotile; this is always confined to cross fibre veinlets in serpentine. Read (1962) in his description of the mineral, notes that the veinlets are in fact zones of shear. Chrysotile, then, could be a polymorph of serpentine stable at high strain rates.

4.3 SIGNIFICANCE OF THE MYLONITES OF THE SOUTHERN GNEISSES

The Southern Gneiss Unit forms a belt of rock between a tectonically older complex of gneisses to the north-east and the Major Bergen Arc to the west. This rock unit shows features compatible with high strain rates and temperatures at virtually constant pressures. The rock types developed are similar to those seen in classical mylonite terrains. Their regional extent and vast thickness, however, preclude their development at the base of major thrusts, as the frictional forces required in their generation would make thrusting on the scale seen in the Bergen/Jotunheim area impossible (Hubbert & Rubey, 1959, 1 & 2). One must therefore conclude that the period of plasticity and grain size reduction predated the brittle events, and that the mylonites of the Southern Gneiss Unit represent deformation occurring in the "root zone" of an orogenic belt, and were moved from this region as a basement nappe and given their flat attitude during the later brittle events.

The rocks do not conform to the original definition of mylonite (Lapworth, 1885), in the sense that they are demonstrably not formed by the passage of two rigid plates past one another. They are more the product of a pure shear flattening. It is considered that mylonites of this type
are probably more common than hitherto recognised (Johnson, 1967); that the general confusion in the literature over their formation and the plethora of terms for their description is due to workers trying to apply Lapworth's original concept and description to related yet genetically different rocks. It is proposed, therefore, to remove confusion when discussing rocks with a high strain rate component, to call all such rocks "Tectonites". By the addition of a prefix, "P" or "T" one could then subdivide "P Tectonites", those defined by Lapworth, from "T Tectonites", those discussed here.
5. GENERAL CONCLUSIONS

The conclusions are divided into three parts: a summary of the Osterøy gneisses and their formation; a regional synthesis; suggestions for further research.

5.1 THE HISTORY OF THE NORTH OSTERØY GNEISSES

5.1.1 Summary of Lithologies

The rocks forming the north-eastern part of Osterøy comprise portions of basement gneisses, migmatised metasediments, and acid and basic plutonics. The rocks are now all gneissose and preserve a complex history of metamorphism and deformation. The gneisses belong to two major units: one autochthonous, the Northern Gneiss Unit; the other para-autochthonous, the Southern Gneiss Unit. A zone of translation, the Tysse Thrust Zone, separates the two units. The Southern gneisses are derived from and lie above the Northern gneisses.

The Northern gneisses are divided into three sub-units, primarily on a deformational basis, but the divisions also contain distinct lithologies. The oldest unit, the Unflattened Gneisses, consists of gneisses dated at 1833± 242 Ma. (Gray,1978), which resemble the Jostedal Complex (Bryhni, 1966). This is overlain by a sedimentary complex which has been extensively migmatised, possibly at 1509± 81 Ma (Gray,1978), and infolded with the basement prior to 1021± 84 Ma. (Gray,1978). This, the Gneiss and Migmatite Complex, resembles the Fjordane Complex (Bryhni,1966). Moving up the structural section the Gneiss and Migmatite Complex is reworked to form the Flattened Gneiss and Migmatite Complex.

The Southern gneisses are a 3 km thick sequence of mylonitic gneisses formed during two mylonite-producing deformations. The mylonitised rocks
are derived from the Northern gneisses, but the unit contains two suites of rocks not found in the north. The first, a suite of mica schists and quartzites is found in a schuppen zone beneath the allochthonous Bergen Arc nappes. These rocks are correlated with those of the Samnanger Complex (Faerseth et al.,1977; Kolderup & Kolderup,1940), within the Arcs by tracing the continuous outcrops into the Samnanger region. The second contains two groups of plutonic rocks, firstly amphibolites intruded into the mylonites towards the end of their formation, and secondly, granitic augen gneisses intruded into the mylonites, mica schists and quartzites early in their deformation. The intrusive augen gneisses are dated at $794.08 \pm 18.82$ Ma. (appendix 1) and contain xenoliths of the Samnanger sediments from the schuppen zone. This implies that the Samnanger sediments, and much of their deformation, are Precambrian and not Lower Ordovician as has been hitherto assumed (Faerseth et al.,1977). This discovery removes much of the basis of Sturt and Thon's (1976) criticism of Gee and Wilson's (1974) claim that the Caledonian deformation and metamorphism in the major part of Scandinavia was a single late event and that "there is no evidence of major unconformity within the Lower Paleozoic sequence of the Scandinavian Caledonides". The major unconformity between Samnanger and Holdhus (Faerseth et al.,1977) rocks described by Sturt and Thon (1976), is therefore within the Lower Paleozoic, but between rocks deformed in the Precambrian and fossiliferous, Ashgill (Faerseth et al.,1977) sediments.

5.1.2 Summary of Metamorphic and Deformational History

There are four stages to the development of the North Osterøy Gneisses, each stage a complete orogenic cycle punctuated by a period of sedimentation (fig.149). The first three stages were very similar and involved a cyclic repetition of coaxial folding on a north-east - south-west axis, with a sense of translation, given by the vergence of the folds, to the
TABLE SUMMARIZING DEVELOPMENT OF THE
NORTH OSTERØY GNEISSES

<table>
<thead>
<tr>
<th>DEFORMATION PHASES</th>
<th>FOLDS</th>
<th>SCHISTOSITIES</th>
<th>LINEATIONS</th>
<th>BASIC INTRUSIONS</th>
<th>ACIDIC INTRUSIONS</th>
<th>METAMORPHISMS</th>
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<td>D&lt;sub&gt;Early&lt;/sub&gt;</td>
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<td>?</td>
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<td>S1</td>
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<td>F2</td>
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<tr>
<td>D8</td>
<td>JOINTING</td>
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[Diagram showing the sequence of events with joints and thrusting.]
north-west. At each stage the latest sediments became incorporated into the basement.

The first episode involved isoclinal folding of an early banding (1855 ± 242 Ma., Gray, 1978) in an amphibolite facies environment, followed by uplift and sedimentation.

In the second episode the sediments were migmatised at middle amphibolite facies (at around 1509 ± 81 Ma., Gray, 1978). This was followed by retrogression to lower amphibolite facies, accompanied by further isoclinal folding (prior to 1021 ± 84 Ma., Gray, 1978), uplift and a new sedimentary cover.

The third stage differed from those preceding because the folding at amphibolite facies produced a thick sequence of mylonites (794.08 ± 18.82 Ma., appendix 1) and, at a late stage, translation of allochthonous and para-autochthonous nappes derived from sediments and the mylonites. These moved in a direction perpendicular to the north-east - south-west trend of the fold axes. The third suite of sediments was deposited on these nappes sometime during the Ashgill (Faerseth et al., 1977).

The final episode differed from the rest in that while deformation started more or less coaxial to the earlier structures, it later developed to produce north-west - south-east trending folds. There are several fold phases involved in the episode, each occurring in progressively lower greenschist facies environments. The thrusts formed in the preceding episode were reactivated, more mylonites were produced and new brittle thrusts developed in the nappes, cross-cutting the earlier structures. Translation was again perpendicular to a north-east - south-west axis.

Later folding on a north-west trend, folded these thrusts and was responsible for the Bergen Arc outcrop pattern.
This history, summarised in the cross sections (figs. 158 & 150) is not immediately apparent from the outcrop patterns (map 1), which are controlled by major F2 and F5 folds. The linearity of outcrops is produced by erosion of the flat limbs to leave intervening steep limbs. The complete history is only deduced through detailed structural and lithotectonic mapping. The key to the understanding of the history lies in appreciating that the main effect of deformation was flattening, and that the deformations were heterogeneous, increasing in intensity up the structural section to the south-east. The lowest part of the structural pile shows the oldest structures, which are flattened out of recognition at higher levels. The mylonitic para-autochthonous nappe being the highest structural unit, it therefore contains the largest number of deformations, but this is superficially the simplest unit because cumulative superposition of these flattenings has largely obliterated the early structures.

The mylonitic unit, once formed, 'cushioned' the underlying basement from further deformation by absorbing all the stresses in superplastic deformation, up to the point where metamorphic retrogression precluded further recrystallisation in the mylonite system (D5). The Tysse Thrust then acted as a plane of transport and is a structural boundary. The thrust separates the first two episodes, which may be regarded as separate Precambrian basement forming cycles, similar to those seen in Telemark (Torske, 1977), from the last two. These, although chronologically separate are more closely related, a conclusion reached from a consideration of the morphology of the parental regional structures. Both are the result of metamorphism and deformation in a thermal environment of 400-500°C, at depths of 15-18 Km and of strain rates exceeding $10^{-11}$/sec. These conditions are only attained in the root zones of orogens. The mylonites therefore can be regarded as a deep level ductile equivalent
Fig 150. DIAGRAM SUMMARIZING DEVELOPMENT OF THE NORTH OSTERØY GNEISSES

Episode 1

1. Sediments

Episode 2

2. Sediments

Episode 3

3. Sediments [Ashgillian]

Episode 4

Northern Gneiss Unit

TYSSE THRUST

BERGEN ARC THRUST

Bergen Arcs

D2 THRUST

Southern Gneiss Unit

S3b (sediments [1] mixed with basal gneisses)
to brittle thrusting at the surface. The brittle thrusts cut the mylonites and are therefore later; they are considered to have formed after uplift (generated by ductile deformation at depth) and are responsible for the final removal and emplacement of the mylonitic unit.

5.2 REGIONAL SYNTHESIS

North Osterøy forms part of a triangle between the Bergen Arcs and Bergsdalen Nappes, and is ideally placed for the study of the response of the basement to the development of these structures. By detailed mapping across the basement/Arc contact on Osterøy and from the deformation history of the Arcs (Faerseth et al., 1977), correlative histories have been established for a protracted period of their history. This correlation is summarised in figure 151.

Assuming that the Northern gneisses are foreland and not part of a larger basement nappe, they can be used as a base line to which structures elsewhere may be related. It is proposed that the basement formed a permanent nucleus, which deformed in sympathy with its adjacent rocks. The cover rocks must therefore contain at least as many deformation phases, from the point of their sedimentary or tectonic emplacement, as the basement. A point of confluence in structures in the two areas is therefore a time of emplacement. On this basis it has been established that the Bergen Arc nappes were emplaced relatively early, during the third stage of development of the basement.

Folding before, during and after this emplacement, was on a north-east axis, and this must be significant in any consideration of an origin for the Bergen Arc nappes. The only major structure with a north-east trend is the Bergsdalen/Jotun mountain range. Unfortunately, the contact with the Bergsdalen nappes lies to the south-east of Osterøy, but based
**Fig 151. CORRELATED DEVELOPMENT HISTORIES FOR THE BERGEN ARCS & N.E. GNEISSES**

<table>
<thead>
<tr>
<th>BERGEN ARCS</th>
<th>N.E. GNEISSES</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Bergen Arc Synform</strong> (NS)</td>
<td></td>
</tr>
<tr>
<td><strong>Non Cylindrical Folding on a N.W. Trend</strong></td>
<td></td>
</tr>
<tr>
<td><strong>Folding and/or Flattening</strong></td>
<td><strong>Flattening</strong></td>
</tr>
<tr>
<td><strong>MINOR THRUSTING AND MYLONITE FORMATION</strong></td>
<td></td>
</tr>
<tr>
<td>Sediments and Pegmatites Isoclinally Folded on a N.N.E. Trend</td>
<td>Pegmatites Folded on a N.N.E. Trend</td>
</tr>
<tr>
<td>Ashgillian Holdus Sediments Deposited, Pegmatites Intruded</td>
<td>Pegmatites Intruded</td>
</tr>
<tr>
<td><strong>MAJOR BRITTLE THRUSTING EVENT</strong></td>
<td></td>
</tr>
<tr>
<td>Ductile Thrusting, Mylonite Formation, Isoclinal Folding on a N.E. Trend with the Production of a Strong 'L' Tectonic Fabric (807 Ma)</td>
<td></td>
</tr>
<tr>
<td>Sammanger Sediments Deposited</td>
<td></td>
</tr>
<tr>
<td>Thick Cover (Volcanics)</td>
<td>Thin Cover (Quartzites)</td>
</tr>
<tr>
<td><strong>SEPARATE PRE THRUSTING HISTORIES FOR EACH THRUST UNIT</strong></td>
<td></td>
</tr>
<tr>
<td></td>
<td><strong>Minor Thrusting</strong></td>
</tr>
<tr>
<td></td>
<td><strong>Flattening</strong></td>
</tr>
<tr>
<td></td>
<td>Migmatization of Sediments and Folding on a N.E. Trend (1509 Ma)</td>
</tr>
<tr>
<td></td>
<td>Sediments Deposited</td>
</tr>
<tr>
<td></td>
<td>Intrafolial Isoclines, New Foliation Develops</td>
</tr>
<tr>
<td></td>
<td>Oldest Detectable Foliation (1855 Ma)</td>
</tr>
<tr>
<td></td>
<td>?</td>
</tr>
</tbody>
</table>
on the structures described from the Osterøy basement (this thesis), Eksingedalen and the lower Bergsdalen Nappe (Gray, 1978); the lower Bergsdalen Nappe (fig. 152), (Winter, 1978; Evans, 1978) and excursions made from the basement into the Bergsdalen Nappes east of Osterøy, a rudimentary deformation history for the Bergsdalen Nappes has been established (fig. 153).

There are several problems in this compilation:

1) There is good agreement with Gray on the Sequence of basement structures, but she does not integrate these with the deformations observed in the Bergsdalen Nappes.

2) Gray describes folding in the easternmost part of the Bergsdalen area on steep east-south-east plunging axes. These are not included in figure 153, because they are not seen further south and not understood in terms of deformation in the Osterøy basement.

3) In the Bergsdalen area, Gray, Evans and Winter describe their earliest folding as isoclinal, affecting quartzites and sediments and cut by syntectonic granite intrusions. These granites are dated at between 1274 ± 48 Ma. (Pringle et al., 1977) and 971 ± 71 Ma. (Gray, 1978). This situation is similar to that described from Osterøy (above). The isoclines are correlated with F3 on Osterøy, and the sediments with those in the Samnanger complex of the Bergen Arcs.

4) After the isoclines there are two phases of folding on a north-east trend before folding on a north-west axis. These latter north-west folds are correlated with F5 on Osterøy on the basis of their style and axial direction.

5) Gray would like to put the Lower Palaeozoic unconformity after the first of the two north-east trending fold phases on the basis that only the second affects the Palaeozoic sediments; Forster (per. comm. 1979) however, describes a hidden intrafolial fold phase in these sediments.
GENERALIZED GEOLOGICAL MAP OF THE BERGSDALEN QUADRANGLE

After KVALE (1964).
**Fig 153. CORRELATED DEVELOPMENT HISTORIES FOR THE BERGEN ARCS, N.E. GNEISSES AND BERGSDALEN**

<table>
<thead>
<tr>
<th>BERGEN ARCS</th>
<th>N.E. GNEISSES</th>
<th>BERGSDALEN</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Bergen Arc Synform (Thrusting Over Bergsdalen)</strong></td>
<td></td>
<td><strong>Much Weaker</strong></td>
</tr>
<tr>
<td>Non Cylindrical Folding on a N.W. Trend</td>
<td></td>
<td><strong>Folding on a N.E. Trend and Flattening</strong></td>
</tr>
<tr>
<td>Folding and/or Flattening</td>
<td><strong>Flattening</strong></td>
<td></td>
</tr>
</tbody>
</table>

**MINOR THRUSTING AND NYLONITE FORMATION**

<p>| | | |</p>
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Sediments and Pegmatites</strong></td>
<td><strong>Pegmatites Folded on a N.N.E. Trend</strong></td>
<td><strong>Sediments and Pegmatites Folded in Tight to Inclined Folds on a N.N.E. Trend</strong></td>
</tr>
<tr>
<td>Inclinedly Folded on a N.N.E. Trend</td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Ashgillian Holdens</strong></td>
<td><strong>Pegmatites Intruded</strong></td>
<td><strong>Fossiliferous Sediments of Bergsdalen/Jotun Deposited, Pegmatites Intruded</strong></td>
</tr>
<tr>
<td>Sediments Inclined, Pegmatites Intruded</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**MAJOR HLITITE THRUSTING EVENT**

Ductile Thrusting Nylonite Formation, Inclined Folding on a N.E. Trend with the Production of a Strong 'L' Tectonic Fabric  
(807 Ma — 1227 Ma)

<table>
<thead>
<tr>
<th>Samnanger and Bergsdalen Sediments Deposited</th>
<th>Thin Cover</th>
<th>Thick Cover</th>
</tr>
</thead>
<tbody>
<tr>
<td>Separate Pre Thrusting Histories for Each Thrust Unit</td>
<td>Minor Thrusting</td>
<td>?</td>
</tr>
<tr>
<td></td>
<td>Flattening</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Migmatization of Sediments and Folding on a N.E. Trend (1509 Ma)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Sediments Deposited</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Intrafolial Isoclines, New Foliation Develops</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Oldest Detectable Foliation (1855 Ma)</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
It is therefore suggested that the unconformity occurs immediately after the early isoclinal folding, a sequence similar to that in the Arcs (Faerseth et al., 1977).

6) The first thrusts in the Osterøy basement were either late D2 or early D3 and trend north-east (map 1, fig. 1). They displaced the early migmatised sediments and could be regarded as the first stages in the development of the Bergsdalen Nappes. This early thrusting is however, regarded as separate from the deformation which produced the Bergsdalen Nappes and as the product of pre-Bergsdalen tectonic activity along the site in which the Bergsdalen sediments accumulated, these early thrusts are brittle structures and it has been shown that the early deformation accompanying emplacement of the Bergsdalen Nappes was super-plastic rather than brittle. This earlier tectonic activity may have provided a source for some of the sediments seen in the Bergsdalen Nappes.

The summary of deformation in the region presented in figure 153 serves as a common framework to be modified by later workers. The immediate conclusion to be drawn from it is that the deformation sequence in the Bergsdalen Nappes is very similar to that of the Bergen Arcs. The Arcs, however, continued to deform after the Bergsdalen Nappes with the formation of the north-south aligned synform, with accompanying transport over the Bergsdalen Nappes. One might speculate, from the presence of tectonised Devonian sediments north of the Arcs (Hoisaeter, 1971), that this stage of development may also be post-Devonian!

The conclusions drawn from the examination of the North Osterøy basement are:

1. The Northern Gneiss Unit forms a foreland to the Bergen Arcs and Bergsdalen Nappes.
2. From a structural point of view, the Southern Gneiss Unit is an integral part of the Bergen Arcs, thus the Tysse Thrust is the "basal thrust" to the Bergen Arc structure.

3. The gneisses have to some extent deformed with the Bergsdalen and Bergen Arc lithologies and are good indicators as to when the nappes were emplaced.

4. The gneisses have a long history of metamorphism, sedimentation and deformation, prior to their involvement with the Bergsdalen and Bergen Arc Nappes.

5. The Bergen Arcs are a portion of the Bergsdalen/Jotun Nappes, affected by later local deformations. It is the Bergen Arc Synform which is responsible for the marked change in trend between the Bergsdalen and Bergen Arc rocks.

6. The Precambrian date on the granite from Osterøy and the granites in the Bergsdalen Nappes (e.g. Pringle et al., 1975), prove that much of the deformation in the nappes was in fact Precambrian (Sveconorwegian, Andresen et al., 1974), and that, at least in the Bergen/Bergsdalen area, some of the sediments are also Precambrian. The evolution of the Bergen Arc/Bergsdalen/Jotun structure was therefore in two stages, the first Sveconorwegian, the second Caledonian.

7. The Sveconorwegian event involves more ductile deformation than the Caledonian, and in the Osterøy area is probably the more important of the two events.

5.2.1 Speculative Tectonic Reconstruction

The generally accepted view (e.g. Hossack, 1978) as to the origin of the Bergen, Bergsdalen and Jotun nappes, is that they were rooted somewhere in the region now occupied by the North Sea, and travelled south-east over
the North Osterøy basement, to their present positions. It is also generally accepted (e.g. Sturt & Thon, 1976; Faerseth et al., 1977; Pringle et al., 1975; Hossack, 1978) that nappe emplacement and accompanying deformation was entirely Caledonian.

It has been shown above that the formation of the Bergsdalen Nappes and Bergen Arcs was a two-stage process; in the early Sveconorwegian event, deformation occurred at deep crustal levels, whilst in the later Caledonian event deformation occurred at shallower crustal levels. In both events the sense of transport, determined from the vergence of the ductile structures which accompany the thrusting, was from south-east to north-west.

These discoveries conflict with the generally accepted view, and from the Osterøy viewpoint, a preferred model is one of a proximal root for the nappes to the south-east. This model is similar to that proposed by Banham et al. (1979) for the Sygnefjell area along strike from Osterøy. It implies the presence of a suture along the Bergsdalen/Jotun axis; the arguments for and against this are discussed in Banham et al. (1979). The difference here is that the suture is now regarded as Sveconorwegian-Caledonian rather than just Caledonian. The model for the development of this suture is explained below.

Model

It is proposed that the present site of the Bergsdalen and Jotun Nappes has acted as a "mobile zone" since 1800 Ma. Its development since then is summarised in a series of north-west - south-east sketch sections across the suture (figs. 154, 155 & 156).

Figure 154

a) Situation at the time of the pre-D2 migmatisation.
b) D2, F2 folding followed by thrusting to the north-west.
MODEL FOR DEVELOPMENT OF THE
SVECONORWEGIAN-CALEDONIAN JOTUN SUTURE, PART 1

NW

OSTERØY

SE

1500 Ma

?FJODANE COMPLEX?

SEDIMENTS

MIGMATISATION

? JOSTERDAL COMPLEX ?

Early / D1 Basement

D2

1020 Ma

This symbol will be used to refer to D2 basement from here on

SITUATION FOR THE DEPOSITION OF THE SAMNAGER ROCKS

Erosion

Samnanger sediments deposited

OCEAN

ANORTHOSITIC

Activity starts on this margin now

ANORTHOSITES

Psammitic

Gabbros & Volcanics

Pelitic
c) Erosion and deposition of the Samnanger rocks, which included gabros and volcanics to the east. At this stage an ocean existed south-east of Osterøy.

**Figure 155**

d) The ocean begins to close, deforming the sediments. Granites are intruded into the sediments and basement at this time.

e) The deformation, which started at the toe of the continent in (d) has now affected the foreland. The nappes of Samnanger rocks are emplaced and part of the continental toe (anorthositic) is pushed over them.

f) Activity on the northern margin now ceases, the new mountains are eroded and Ashgillian sediments deposited over them, and in the remains of the ocean which still existed south-east of Osterøy.

**Figure 156**

g) The Caledonian orogeny finally closes the ocean by activity on the southern margin, transporting the Palaeozoic nappes south over the southern continent. The southern continent underthrusts its northern counterpart, reactivating the old thrusts and sending more Palaeozoic nappes northwards, producing 'Ivrea type' flakes out of the older nappes (see Gabrielsen et al., 1979).

h) An east-west section through the Arcs and Bergsdalen shows the effects of the later deformation, which also affects the Bergen region.

5.3 FURTHER RESEARCH

5.3.1 Suggestions for Further Research on Osterøy

Geochemistry: The geochemical study of the mylonites (section 4.1.3) is incomplete; more samples should be collected both across and along strike to test the inferences drawn in the pilot study. The geochemistry
MODEL FOR DEVELOPMENT OF THE
SVECONORWEGIAN-CALEDONIAN JOTUN SUTURE, PART 2

NW

D3 (ductile phase)
1000Ma still undeformed here.
Granites
800Ma basement

OSTERØY

SE

Diachronous Deformation
of sediments

D3 (brittle phase)
800Ma

ANORTHOSITES (from continental toe)

800-500Ma SITUATION AT THE CLOSE OF THE FIRST PHASE OF ACTIVITY

eroding mountain chain

ANORTHOSITES [deposition of sediments starts 'Sparagmites']

Cambrian shelf sea sediments

Activity on this margin now ceases

Southern continent

DEPOSITION OF THE LOWER PALAEOZOIC ROCKS

500Ma-Silurian?

ANORTHOSITES
Holdhus sediments

remnants of JOTUN OCEAN

volcanics

Northern continent
pelitic sediments

Activity starts on the southern margin now

Southern continent

quiescent margin
NW

D4

[Post Ashgillian]

reactivation of some D3 thrusts & formation of some new ones

unaffected D3 thrust

North-Eastern Gneiss Basement

unaffected ANORTHOSITE forming an Ivrea type flake

Underthrusting of Northern continent by the Southern continent

(g)

SE

FINAL STAGE OF CLOSURE

Flat lying Palaeozoic Nappes

(g)

W

LATE DEFORMATION ALONG THE AXIS OF THE SUTURE ENDS

ACTIVITY

BergenArcs

D5 + D6

- Devonian?

Øygarden basement

Late emplacement of the Arcs

over

the Bergsdalen

Jotun

(OSTERØY)

(h)

F6

F6

Late emplacement of the Arcs

over

the Bergsdalen

ANORTHOSITE

(h)

(for Map showing Lines of Section, see Fig 4.)
of the migmatites merits more detailed future study, it may be possible to separate the basement gneisses from the migmatised metasediments geochemically.

Geochronology: High strains in the gneisses make dating difficult. However, a date on the veins cutting the F3 and F4 folds (pl. 21) would probably be feasible and useful because this would date the end of D3.

Structure: The study of mylonites began on Osterøy should be extended to another area (e.g. a Laxfordian shear zone in Scotland), so that a comparative study of deformational styles may be made.

The F5 folds present an enigma; their style suggests they are the top limb of a fold overturned to the west. Why these late, non-cylindrical folds should close towards the nappe pile with a sense of transport away from the basement is not understood.

5.3.2 Suggestions for Further Research in the Region

The Øygarden basement and the area south-east of Osterøy, where the Bergsdalen, basement and Bergen Arcs meet are being mapped as part of the Bedford College project. The mapping should be extended to the north-west of Osterøy to trace the Tysse Thrust. The rest of the North Eastern Gneisses (fig. 2) also need mapping to confirm that they are in fact basement and not part of a larger nappe.

The work on the Bergen Arcs, basement and Bergsdalen Nappes should be continued along the northern margin to tie in with the work at Sygnefjell and extended into the nappes in an attempt to prove or disprove the presence of a suture in this region.
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143-150.


APPENDIX I

RADIOMETRIC DATING

Two sets of samples from the North Osterøy Gneisses were selected for Rb/Sr whole rock radiometric dating:

<table>
<thead>
<tr>
<th>1</th>
<th>2</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGII 3</td>
<td>FH 4</td>
</tr>
<tr>
<td>AGII 5 From</td>
<td>FH 5 From the structurally oldest portion of the AGII 8 the AGII 9 Southern Gneiss Unit AGII 12 FH C AGII 13 FH N Northern Gneiss Unit</td>
</tr>
<tr>
<td>AGII 12</td>
<td>FH F</td>
</tr>
<tr>
<td>AGII 13</td>
<td>FH G</td>
</tr>
</tbody>
</table>

The second group was a suite of background gneisses and did not provide any useful data.

The first group was taken from the augen gneiss body outcropping at (10.30, 28.00) and this plotted as shown in figure 157. The plot does not constitute an acceptable isocron but it does provide a rough indication of an age for the suite as shown. The high $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratio indicates the gneisses, which are determined to be of igneous origin from their petrological and structural characteristics, must be derived from Precambrian crustal material.

While the age obtained is meaningless by itself, it does correlate with similar ages on similar bodies cropping out to the north-east of Osterøy (Grey 1973), and furnishes an example of how difficult gneisses can be to date. Even if it were an isocron it would be difficult to decide whether the age represents the time of original intrusion or the close of the last almandine amphibolite facies metamorphism affecting the rocks.
Fig 157.

WHOLE ROCK Rb/Sr PLOT FOR THE FH, AGII

AUGEN GNEISS GROUP

YORK-WILLIAMSON LEAST SQUARES FIT

AG II GROUP

<table>
<thead>
<tr>
<th>Line</th>
<th>87Rb/86Sr</th>
<th>s.d.</th>
<th>87Sr/86Sr</th>
<th>s.d.</th>
<th>rho</th>
<th>Xres</th>
<th>Yres</th>
<th>Sample</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1.542</td>
<td>.008</td>
<td>.72717</td>
<td>.00015</td>
<td>0.00</td>
<td>.00775</td>
<td>-0.00024</td>
<td>AG II 13</td>
</tr>
<tr>
<td>2</td>
<td>1.611</td>
<td>.008</td>
<td>.72895</td>
<td>.00015</td>
<td>0.00</td>
<td>.00307</td>
<td>-0.00096</td>
<td>AG II 12</td>
</tr>
<tr>
<td>3</td>
<td>1.587</td>
<td>.008</td>
<td>.72706</td>
<td>.00015</td>
<td>0.00</td>
<td>-0.03133</td>
<td>.00084</td>
<td>AG II 8</td>
</tr>
<tr>
<td>4</td>
<td>1.716</td>
<td>.004</td>
<td>.71772</td>
<td>.00014</td>
<td>0.00</td>
<td>.00162</td>
<td>-0.00023</td>
<td>AG II 9</td>
</tr>
<tr>
<td>5</td>
<td>1.226</td>
<td>.006</td>
<td>.72261</td>
<td>.00014</td>
<td>0.00</td>
<td>-0.01062</td>
<td>.00052</td>
<td>AG II 3</td>
</tr>
<tr>
<td>6</td>
<td>1.502</td>
<td>.008</td>
<td>.72630</td>
<td>.00015</td>
<td>0.00</td>
<td>-0.00188</td>
<td>.00006</td>
<td>AG II 5</td>
</tr>
</tbody>
</table>

No. of cycles 3
Centroid is 1.341 .72456
Slope .011340 .003270
Intercept .73325 .00020
MSWD= 32.56
Variance ratio (F)= 32.21
AGE 794.08+ 18.6 Ma (1-sigma)
The total Rb/Sr ratios for these rocks were determined by XRF analysis at Bedford College and the isotopic ratios determined by mass spectrometry at Cambridge University. The wet chemistry for this was done by Mr H. Lloyd.
## APPENDIX 2

### GEOCHEMISTRY AND TECHNIQUES

Analyses were performed on the specimens listed below:

<table>
<thead>
<tr>
<th>Rock No.</th>
<th>Grid References</th>
<th>X</th>
<th>Y</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>FH 4</td>
<td>11.58 34.97</td>
<td></td>
<td></td>
<td>Pink banded acid gneiss</td>
</tr>
<tr>
<td>FH 8</td>
<td>11.27 34.76</td>
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For a complete list of specimens considered and their locations see the microfiche included in Volume 2.

Specimens for Geochemical analysis were prepared using a roll jaw crusher and agate tema, these were thoroughly washed and dried between each specimen. The analytical method used was X-Ray Fluorescence Analysis on a Philips PW1212 X-Ray spectrometer. The powdered samples were prepared for analysis in two ways; for
restricted major element analysis and trace element analysis, 6 grams of rock powder was mixed with 1 gram of PF Bakelite resin then pressed into a pellet at 15 tons. The pellets were then hardened for 20 minutes at 105°C in an oven.

For major element analysis 0.4 grams of rock powder was weighed with 2.4 grams of Johnson Matthey Spectraflux 104 in a Platinum Gould crucible, and fused at 1000°C in a furnace. Duplicate samples were weighed in each case. The fusion was then formed into a bead while still molten, using a former and press heated to 200°C. As the beads are hydroscopic, the samples were cooled in a desiccator and analysed as soon as possible after preparation.

The Bedford College Spectrometer had no previous calibration for rocks with silica contents as high as those for some of the specimens from North Osterøy. A calibration therefore was erected using Published standards (Abby, 1978; Flannagan, 1973) and a set of rocks from the Osterøy Gneisses. These had been previously analysed (by Mr H. Lloyd) using standard wet chemical techniques. Published standards were also analysed with my samples to check the calibration and instrumental drift. The results on these checks (figs. 159-163) show good correlations between determined chemistry and the published analyses.

The analyses determined from the samples listed above are given in figures 164-169. Attempts at interpretation of the meaning of these analyses in the Osterøy context did not prove fruitful so no further analyses on the CO₂, F, H₂O⁺, H₂O⁻ or FeO₂ contents of the rocks were made. This accounts for the low totals seen in some of the analyses.
### Figure 159

#### Table of Analyses Checks

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| Y / Zr  | 32  | 39  | 5 | 34 | 19 | 61 | 25 | 33 | 6 | 66 | 132 | 30 | 48 | 16 |
| Cr / Zr | 181 | 281 | 149 | 100 | 178 | 266 | 253 | 80 | 8 | 176 | 369 | 43 | 255 | 61 |
| Mo / Zr | 14  | 8  | 1 | 11 | 6 | 16 | 10 | 20 | 1 | 30 | 18 | 15 | 6  | 2  |
| Ti / Zr | 412 | 305 | 194 | 347 | 345 | 67 | 67 | 438 | 723 | 103 | 77  | 344 | 242 | 334 |
| Ti / Co | 10 | 67 | 2 | 91 | 107 | 63 | 10 | 89 | 157 | 25 | 25 | 119 | 63 | 104 |
| Y / Zr | 0.13 | 0.29 | 0.02 | 0.23 | 0.26 | 0.31 | 0.11 | 0.20 | 0.32 | 0.27 | 0.19 | 0.22 | 0.21 | 0.08 |

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Figure 168

MAJOR ELEMENT ANALYSES

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TRACE ELEMENT ANALYSES PPM

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TABLE OF GEOCHEMICAL LISTINGS
### Figure 169

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APPENDIX 3

MINERAL IDENTIFICATION AND TECHNIQUES

Most minerals were identified using standard transmitted light techniques on rock slices (0.03 mm thick) using a Leitz polarising microscope. The relative abundance of mineral species was largely estimated visually, with regular checks on the accuracy of estimation using a mechanical stage and point counter.

Where difficulty was found in identifying minerals, alternative techniques were employed: if a sample of the pure mineral was not easily obtainable from the crushed rock, the Universal Stage was used to determine the precise optical characteristics; where a pure sample could be obtained, by magnetic separation or heavy liquid separation and hand picking, the RI of the mineral was determined using RI liquids, an Abbe Refractometer and spindle stage; if enough of the sample was obtained, an X-Ray Powder Photograph, from which the "d" spacings were calculated, was made of the finely powdered sample. The photographs were taken with a Phillips 14.6 mm radius Debye-Scherrer camera with a Straumanis film mounting, in Ni filtered Cu(Kα) radiation.

The Universal stage was also used to determine the optical orientation of quartz crystals, this being done to determine whether there be any crystallographic fabric present in the rocks. The technique was applied to three right angle sections cut from field oriented specimens.

Details of the techniques used are to be found in (Hartshorne and Stuart, 1970), (Zussman, 1967).
APPENDIX 4

STRUCTURAL ANALYSIS

The nature of the project and the volume of readings made manual manipulation of structural data impractical. It was therefore decided to employ the aid of a computer to process it. To this end a data base was created, and stored on file in the computer. This contained all the structural information on a grid referenced basis (see program instructions below).

A fortran program (Sort 2) was then written to handle and sort this data on the basis of either structural type or grid reference and structural type.

The data were then plotted onto a stereo net, programs to do this have already been written, and to avoid duplication, one of these was obtained from Dr W. Owen (University of Birmingham). It was found that this program was unable to handle the quantity of data to be processed, it did not contour on a percentage basis, and the rotational facilities were inoperative. The program was therefore rewritten to correct these shortcomings (Program Sterio 2). Further modifications were made to enable it to handle "c" axis data obtained from universal stage work, (Program Sterio 3).

The University of London computer has a microfilm facility, this is ideal for geological computing. A program (Draw) was written therefore using these facilities. This produces an output on film which is a contoured stereonet. This can be used as a transparency for projection, or can be enlarged and printed.
All these programs were loaded onto a library in the computer for public use; instructions as to the use of this library are given below.

The computer was also used when calculating and plotting the geochemical data.

Note. In the listings given below certain symbols have been transliterated.

e.g. : has become 0

Reference as to what these transliterated symbols should be can be found in the ULCC Microfilm documentation.
STATISTICAL STEREOonet CONTouring PROGRAM, MICROfilm OUTPUT.

THIS PROGRAM PACKAGE STATISTICALLY MANIPULATES AND DISPLAYS
NUMERICAL STRUCTURAL DATA 'PRESENTED IN VARIOUS FORMS'
ONTO 35 MM BLACK AND WHITE FILM. THE OUTPUT IS IN THE FORM
OF A CIRCLE, FILLED WITH LINEAR CONTOURS OF VARYING INTENSITY,
A LEGEND, AND KEY.

THE PHILOSOPHY BEHIND THE CONSTRUCTION OF THE PACKAGED-

THERE ARE THREE PROGRAMS,
(1). SORT2 (F.W.M.HOPPER, 1977)
(2). STERIO2 (DR.W.Owen, MODIFIED BY F.W.M.H., 1977)
(3). DRAW (F.W.M.HOPPER, 1977)

(1) TAKES DATA FROM A RANDOM 'PILE', JUST AS IT WAS TAKEN IN
THE FIELD, ANDSortS IT INTO 'BLOCKS' ON THE BASIS OF STRUCTURAL TYPE
OR ON STRUCTURAL TYPE AND LOCALITY. TO THE 'BLOCKS' THE PROGRAM ADDS A
STRING OF 'STEERING' PARAMETERS, SPECIFIED BY THE USER.
THUS IN ONE RUN VARYING FORMS OF DATA CAN BE DEALT WITH. THE DATA
'BLOCKS' ARE WRITTEN ON TO A TAPE WHICH IS THEN IMMEDIATELY USED BY
THE SECOND PROGRAM.

(2) CALCULATES AND WRITES THE DATA ON TO ANOTHER TAPE WHICH IS THEN
IMMEDIATELY USED BY THE THIRD PROGRAM. ALL OTHER OUTPUT FROM THE
SECOND PROGRAM IS SUPPRESSED.

(3) CALCULATES AND MANIPULATES THE DATA, AND USING DIMFILM,
(U.L.C.C.) DRAWS A PAGE WITH TITLEING, KEY AND CIRCLE. IT THEN
FITS THE MATRIX CONTAINING THE DATA OVER THE CIRCLE AND CONTOURS IT
IN ACCORDANCE WITH LIMITS PRESET WITHIN THE PROGRAM. THE OUTPUT IS
IN THE FORM OF A ROLE OF NEGATIVE FILM.
THE OUTPUTS ARE SELF EXPLANATORY.

USING THE PACKAGE.

THE USER MUST READ THESE INSTRUCTIONS SO-
THE PACKAGE IS DESIGNED SO THAT THE 'WORKS' OF BOTH PROGRAMS SHOULD
NOT NEED TO BE ALTERED. THE PACKAGE IS FLEXIBLE, THE FLEXIBILITY IS
ACHIEVED BY A SET OF STEERING CARDS EXTERNAL TO THE PROGRAM,
FED IN AS DATA. THUS A SENSIBLE USER, KNOWING WHAT HE WANTS,
AND FOLLOWING THE INSTRUCTIONS GIVEN BELOW,
NEED NOT KNOW HOW IT WORKS.

THERE ARE 4 SECTIONS OF INSTRUCTIONS-
I. THE JOB CONTROL LANGUAGE (ORANGE CARDS) 'A SPECIMEN OF WHICH IS GIVEN'
II. THE STEERING CONTROLS FOR THE FIRST PROGRAM
III. THE FORM OF THE STRUCTURAL DATA INPUT.
IV. CONSTRUCTION OF THE WHOLE PACKAGE 'SPECIMEN GIVEN'.
II. JOB CONTROL

AS THE PACKAGE IS THREE PROGRAMS, ONE MUST BE COMPILED (USE MNP) AND RUN WITH THE DATA THEN THE NEW DATATAPE (SCRATCH) IS PASSED ON TO THE SECOND PROGRAM WHICH MUST BE COMPILED (MNP) RUN. SIMILARLY, THE THIRD PROGRAM MUST BE COMPILED (MNP) AND RUN USING THE NEW DATA TAPE (FILMPUT).

THE FOLLOWING 4 LISTS SHOW THE JOB CONTROL STRUCTURE POSSIBILITIES.
A, PROGRAM + DATA FED IN AS CARDS
B, PROGRAM ON FILE + DATA AS CARDS
C, PROGRAM + STRUCTURAL DATA ON FILE + STEERING DATA ON CARDS
D, PROGRAM + STEERING DATA ON CARDS + STRUCTURAL DATA ON FILE

THE PROGRAM STORED ON FILE AT BEDFORD IS ALREADY COMPILED IT IS THEREFORE ALREADY IN BINARY; THE TWO SETS OF JOB CONTROL INSTRUCTIONS REFER TO 1) THE COMPILED FILE, (2) THE CARD DECK.

IT IS ASSUMED THAT SITUATIONS A, OR B, ABOVE, WILL BE THOSE MOST COMMONLY USED (B = RELEVANT TO BEDFORD, A = GENERAL).

A

JOB (..........., J9, T120, M7600) (NAME)
ATTACH (DIMMILF, DIMMIF, ID=PUBLIC)
ATTACH (MFILM, MICROFILM, ID=PUBLIC)
LIBRARY (DIMMILF, MFILM)
MNP (E=2, R=0, L=0, I=INPUT)
LGO (INPUT)
RETURN (LGO)
MNF.
LGO (SCRATCH)
RETURN (LGO)
MNF.
LGO (FILMPUT)
<0 R>

JOB (..........., J9, T120, M7600) (NAME)
ATTACH (PLOTST, ID=LIBBED)
ATTACH (DIMMILF, DIMMIF, ID=PUBLIC)
ATTACH (MFILM, MICROFILM, ID=PUBLIC)
LIBRARY (PLOTST, DIMMILF, MFILM)
SORT2 (INPUT)
STER102 (SCRATCH)
(XXXXXXXXXXXXXXXXXXXXXXX SEE BELOW XXXXXXXXXXXXXXXXXX)
DRAW (FILMPUT)
<0 R>

(XXXXXXXX IF YOU WANT A PAPER OUTPUT AS WELL XXXXXX)
INSERT HERE 0--
STER101 (SCRATCH)

OTHERWISE REDUCE THE SIZE OF THE ARRAYS IN PROGRAM 1 FROM 5000-1000
THIS STILL PROVIDES ROOM FOR 1000 REA DINGS (DATA CARDS)
THE JOB TAKES 75,000 WORDS OF MEMORY FOR THE 1ST PROGRAM AS IT STANDS:
THIS REQUIREMENT ENDS WHEN THE FIRST PROGRAM STOPS; THE DATA IS
THEN BACK ON TAPE
NOTE...THE THIRD PROGRAM UTILISES THE MICROFILM PLOTTING PACKAGE
CALLED 'DIMFILM'. THUS IT CAN ONLY BE RUN ON COMPUTERS WHERE THIS
LIBRARY IS AVAILABLE.

[III]. STEERING.

THE FIRST PROGRAM READS THE INFORMATION FROM YOUR DATA CARDS.
IT THEN COMPARES THIS WITH INFORMATION FROM THE STEERING CARDS AND
MANIPULATES IT ACCORDING TO YOUR INSTRUCTIONS. THE STEERING CARDS ARE
PLACED BEFORE THE DATA CARDS AND ARE SEPARATED FROM THEM BY AN <ER>
THERE ARE IN ALL 135 STEERING CARDS, AND THEY ARE ARRANGED IN
'BLOCKS' THUS 0--
CORD. 30 CARDS FORMAT(F5.2,1X,F5.2,1X,F5.2,1X,F5.2)
THESE CARDS ARE USED WHEN THE GRID SEARCHING FACILITY OF THE PROGRAM
IS TO BE UTILISED. THIS FACILITY ENABLES THE USER TO SPECIFY
SUB AREAS OUT OF THE TOTAL AREA HIS DATA COVERS AND TO PLOT ONLY
THAT DATA WITHIN THESE SUB AREAS.
THE 'CORD' CARDS SPECIFY ONE GRID AREA YOU WISH SEARCHED PER CARD.
IE. EACH CARD DEFINES ONE SQUARE. X,X, FIRST, Y,Y, SECOND, USING 5
FIG URE NUMBERS, THE MIDDLE ONE OF WHICH IS A DECIMAL POINT.
EG. 01.10 10.01 01.10 10.01 DEFINES A GRID SQUARE.
SHOULD THIS FACILITY NOT BE REQUIRED THEN THE FIRST 30 CARDS CAN
BE BLANK (BLANK FILLED).

TITL E 17 CARDS FORMAT(5A10,A4)
ON THESE CARDS IN COLUMNS 1-54 YOU CAN WRITE WHAT YOU WANT TO CALL EACH
TYPE OF STRUCTURAL READING YOU HAVE MEASURED AND WILL BE PLOTTING.
NOTE.
THE ORDER IN WHICH THESE CARDS ARE STACKED MUST BE THE SAME AS THAT
IN WHICH YOU STACK ALL THE SUBSEQUENT STEERING CARDS.
IF YOU ARE NOT USING 17 DIFFERENT TYPES OF READING THEN MAKE UP THE
NUMBERS OF TITLE CARDS WITH BLANKS.
TO GET A NEAT TITLING, THE TITLE SHOULD BE CENTERED IN THE
54 SPACES AVAILABLE ON THE TITLE CARD.

RAMAX 17 CARDS FORMAT(4A5)
INSERT THE INFORMATION IF NECESSARY--
(AS SPECIFIED IN THE COMMENTS FOUND AT THE TOP OF THE SECOND
PROGRAM (STERIO2)) IN THE FIRST 20 COLUMNS.
NOTE.
THE PLAGERISING FACILITY OF DR. OWEN'S PROGRAM HAS NOT BEEN TESTED
BY F.W.M.H.
BLANK FILL UP TO 17 IF ALL THE CARDS ARE NOT USED.

SYM B 17 FORMAT(A2)
IN THE FIRST TWO COLUMNS OF THE CARD WRITE TWO LETTERS UNIQUELY
IDENTIFYING EACH TYPE OF READING YOU WANT PLOTTED. EG. BD FOR BEDDING,
FP FOR PLUNGE, LA FOR LINEATION 1, LB FOR LINEATION 2, ETC.

MAKE SURE YOU STACK THE CARDS SO THAT IF BD IS AT THE TOP, THEN THE
1ST 'TITLE CARD' REFERS TO THE BEDDING TOO,
AND SHOULD RAMAX BE USED THEN THE FIRST RAMAX CARD IS THE RIGHT ONE
FOR THIS AND SO ON FOR THE FOLLOWING STEERING CARDS.
IF ALL 17 CARDS ARE NOT IN USE BLANK FILL.

EDOCN 17 CARDS FORMAT (II)

THE CORRECT NUMBER FOR THE TYPE OF READING, EG. BEDDING,
LINEATION, TO BE COMPUTED MUST BE PLACED IN COLUMN 1 HERE.
FOR THE CORRECT NUMBER IE. (1, 2, 3, 4, OR 5), SEE SECTION
[III] BELOW.
<BLANK FILL IF NECESSARY>.

NR 17 CARDS FORMAT (II)

THIS REFERS TO THE NUMBER OF COMPOUND ROTATIONS TO BE MADE.
<SEE COMMENTS IN DR. OWEN'S PROGRAM (STERIO)>. IF ROTATION IS UNUSED
THEN 0 SHOULD BE PLACED IN THE 1ST COLUMN OF ALL 17 CARDS.
IF TNR+ IS NOT 0 AND ROTATIONS ARE TO BE MADE, THEN THE
+DECR, DIPR, ANR+ CARDS MUST BE FILLED, SEE BELOW.

NPC 17 CARDS FORMAT (II)

THIS REFERS TO THE PERCENTAGE AREA OF THE STEREONET YOU WISH TO
SAMPLE AT. IT IS DIRECTLY EQUIVALENT TO THE SIZE OF THE HOLE IN
THE POINT COUNTER WHEN CONTOURING MANUALLY. NORMALLY THIS IS BEST SET
AT 1 OR 2 PERCENT; THUS, 1 OR 2 SHOULD BE PLACED IN THE FIRST COLUMN
OF EACH NPC CARD. BLANK FILL TO MAKE UP NUMBERS.

PASS, DEVIDE 1 CARD FREE FORMAT

TWO VALUES ARE REQUIRED ON ONE CARD, EACH SEPARATED BY A ',' AND
STARTING IN COLUMN 1.
BEFORE THE ',' = PASS
PASS SHOULD BE SET TO 10 IF THE CO-ORDINATE GRID SEARCH
IS TO BE OPERATIVE. IF IT IS NOT REQUIRED ANY NO. EXCEPT 10 EG. 1,
SHOULD BE PLACED THERE INSTEAD.
AFTER THE ',' = DEVIDE
ON THE 'STRUCTURAL' DATA CARDS YOU MAY HAVE WRITTEN
AN 'N' OR 'S' IN THE 7TH COLUMN IF YOU WISH TO USE THE DATA
DIVIDING PROCESS DESCRIBED BELOW, SECTION [III]. THEN-
TO GATHER AND USE ONLY CARDS WITH ONE OR THE OTHER OF N/S, ON THEM
YOU SHOULD PLACE AFTER THE ',', 'O-
1= FOR USING ALL THE DATA. THE N/S FACILITY IS IGNORED THIS IS THE
DEFAULT NUMBER PLACED AFTER THE ','.
2= FOR GATHERING TOGETHER ALL THOSE CARDS WITH 'S' ON THEM,
AND USING THEM ONLY.
3= FOR GATHERING TOGETHER ALL THE CARDS WITH 'N' ON THEM,
AND USING THEM ONLY.

DECR, DIPR, ANR 3 CARDS FREE FORMAT

THES CARDS ARE USED WHEN A ROTATION OR COMPOUND ROTATION
IS TO BE MADE ON ONE OR ALL OF THE DATA SETS PRODUCED,
YOU CAN'T DO DIFFERENT rotations ON DIFFERENT DATA SETS,
BUT YOU CAN ROTATE ONE AND NOT THE REST OR TWO THE SAME
AND NOT THE REST.
SET THE NRM OF THE RELEVANT DATA SET(S) TO
(1-3) DEPENDING ON HOW MANY OF THE DECR, DIPR, ANR CARDS
YOU HAVE FILLED. THIS WILL CAUSE THE ROTATION OF
THAT DATA SET(S).

EACH CARD PRODUCES ONE ROTATION O-
DECR = 3 FIGURE NUMBER FOR THE COMPASS BEARING OF THE ROTATION
AXIS.
DIPR = 2 FIGURE NUMBER FOR THE INCLINATION OF THE AXIS FROM
THE HORIZONTAL.
ANR = UP TO 3 FIGURE NUMBER FOR THE AMOUNT OF ROTATION ABOUT
THE AXIS.

ALL THE NUMBERS ARE SEPERATED BY +, +.
ROTATION OCCURS ONLY IN THE CLOCKWISE SENSE, thus
360 DEGREES OF ROTATION ARE POSSIBLE.

IF THE CARDS ARE TO BE UNUSED, THEY SHOULD BE FILLED WITH 3+0+.
IE.0.0.0. IF THIS IS NOT DONE, THE PROGRAM WILL FAIL.

---------------------------------------------------------------------------------------------------------------------------------

NOTE.

DO NOT INCORPORATE INTO YOUR STRUCTURAL DATA (CARD S WHICH DO
NOT HAVE A MATCHING 'SYMB' ON THEM IN THE (STEERING CARDS). IF THIS HAPPENS THE FIRST PROGRAM WILL TREAT THESE CARDS AS
PUNCHING ERRORS AND TERMINATE THE PROGRAM.

THE STEERING CARD 'BLOCKS' MUST BE PLACED IN THIS ORDER: -
CARD (30)
TITLE (17)
RAMAX (17)
SYMB (17)
EDOCN (17)
NR (17)
NPC (17)
PASS,DEVIDE (1)
DECR,DIPR,ANR (3)

SHOULD YOU HAVE TOO FEW OR TOO MANY STEERING CARDS THEN THE PROGRAM WILL BE STOPPED AND A MESSAGE TO THE
EFFECT GIVEN.

'=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '=' '='

[III]. STRUCTURAL DATA.

DATA IS FED IN ON CARDS, ONE READING PER CARD.
THE CARDS ARE IN THE FORMATO-
(A5,1X,A1,1X,I3,1X,I2,1X,A2,1X,F5.2,1X,F5.2)

SUBDIVIDED.

STRUCTURAL READING= (8X,I3,1X,I2)
THE STRUCTURAL READINGS ARE PLACEDO-
DIP/STRIKE COLUMNS 9-11.
AMOUNT OF DIP COLUMNS 13 14.
N/S DIVIDER= (6X,A1)
SYMB= (15X:A2)
GRID REFERENCE CO-ORDINATES= X,Y,= (18X,F5.2,1X,F5.2)
THE FIRST 5 COLUMNS ARE FOR THE LOCALITY NO. FOR YOUR REFERENCE,
AND LIKESWISE COLUMNS 30 ONWARDS CAN BE USED FOR YOUR COMMENTS
OR SPECIMEN NO. FROM THAT LOCALITY.
COLUMNS 6,8,12,15,18,24, MUST BE LEFT BLANK.
THE FORM OF THE DATA: STRUCTURAL READINGS

4 TYPES OF READING ARE ACCEPTABLE.

(A) PLAGIARISED DATA - USED FOR POLE PLOTS WHERE THE DIP IS
EXPRESSED AS THE RADIAL DISTANCE FROM THE CENTRE, IN THE
RADIAL DIRECTION (SEE COMMENTS AT START OF PROGRAM STERIO).

(B) DATA IN THE FORM OF COMPASS DIRECTION AND DIP FROM HORIZONTAL
IN THAT DIRECTION, IE. AS DEC, DIP OF LINES OR POLES IN SPACE
EG. 090/45

(C) READINGS FOR PLANAR ELEMENTS MEASURED AS DIRECTION OF DIP
AND AMOUNT OF DIP IN THAT DIRECTION, IE. DEC=AZIMUTH OF DIP.
EG. 090/45, NOTE PLANES MUST BE TREATED DIFFERENTLY FROM LINES
WHEN PLOTTING THIS READING IS NOT THE SAME AS IN (B).

(D) READINGS ON PLANAR ELEMENTS USING THE STRIKE DIRECTION AS
THE COMPASS BEARING, ALWAYS ARRANGING IT SO THAT THE DIP IS 90
DEGREES CLOCKWISE FROM THE STRIKE DIRECTION QUOTED, IE. DEC=
STRIKE AND (AZIMUTH OF DIP = STRIKE + 90). EG. 000/45 =
(090/45, (C)) = (000E45 OR 180E45).

IN ORDER THAT THE COMPUTER UNDERSTAND WHICH OF THE 4 SORTS
OF DATA INPUT YOU ARE USING IT IS NECESSARY FOR THE USER
TO TELL IT. THIS IS DONE BY SETTING SPECIFIC VALUES INTO AN ARRAY
CALLED ED0CN, HOW THIS IS DONE IS EXPLAINED ABOVE. HOWEVER,
THE VALUES ARE O -
(A) ED0CN=4 (IF THE PLOT IS OF EQUAL AREA)
(B) ED0CN=5 (IF THE PLOT IS OF EQUAL ANGLE)
(C) ED0CN=2
(D) ED0CN=3

USER SET PARAMETERS

THE DIVIDER.

SHOULD THE USER HAVE SAMPLED TWO AREAS, THEN HE MAY WISH
TO COMPARE OVERALL READINGS FROM THE TWO. FOR THIS PURPOSE A
SYMBOL (N) IS DESIGNATED TO ALL READINGS FROM ONE AREA, AND
(S) TO ALL THE READINGS FROM THE OTHER. THIS N/S IS PLACED
IN COLUMN 7 OF THE DATA CARD. SHOULD THIS FACILITY NOT BE
REQUIRED THEN THE COLUMN MAY BE LEFT BLANK, OTHERWISE IT MUST
BEAR ONE OR THE OTHER.

THE SYMB (STRUCTURAL TYPE IDENTIFIER).
AS HAS BEEN EXPLAINED ABOVE (SECTION [III]) YOU MUST DECIDE ON TWO
LETTERS WHICH IDENTIFY EACH SORT OF STRUCTURAL READING YOU HAVE BEEN
MEASURING. THESE ARE THEN PLACED IN COLUMNS 16 17 OF THE DATA CARD FOR
THAT READING.

THE PROGRAM CAN HANDLE UP TO 17 DIFFERENT TYPES OF STRUCTURE AT ONCE.

THE GRIDS.

THE PROGRAM HAS THE ABILITY (WHEN PASS IS SET TO 10, SEE ABOVE)
TO SEARCH FOR READINGS INSIDE SPECIFIED AREAS WHICH YOU HAVE COVERED
(SEE 'CORD' ABOVE [III]).
YOU CAN LOOK AT 30 DIFFERENT GRID AREAS ON ONE RUN SHOULD YOU WISH TO,
THESE MAY BE SUB AREAS OF EACH OTHER OR SEPARATE AREAS. THEY MAY BE OF
ANY SIZE.
ALL DATA FROM EACH AREA ARE SORTED INTO THE DIFFERENT STRUCTURAL TYPES
SYMB), AND A STEREONET FOR EACH IS PRODUCED, TELLING YOU WHICH (1-30)
COORDINATE CARD WAS USED (SEE [III] ABOVE)
WARNING ***************

THE PROGRAM CAN TAKE UP TO 17 DIFFERENT SYMB'S, SHOULD THESE OCCUR
IN ALL 30 OF THE REQUESTED GRID SQUARES THEN, THE RESULT WILL
BE 510 STEREONETS IE. NO.0F SYMB = MAX NO OF STEREONETS.
THE GRID REFERENCE OF THE LOCALITY WHERE A PARTICULAR READING IS MADE IS PLACED IN COLUMNS 19-23=X, 25-29=Y. IF NO USE IS TO BE MADE OF THIS GRID SEARCH PROCESS LEAVE THE COLUMNS BLANK. THE 1ST 1-5 COLUMNS ARE FOR LOCALITY NOS. THE PROGRAM IgNORES THEM. SIMILARLY WITH COLUMNS 30 ONWARDS.

WHEN PUNCHING DATA ASK THE ADVISOR ABOUT PROGRAMMING A DRUM CARD. THIS WILL HELP ELIMINATE ERRORS AND SPEED PUNCHING. 

[IV]. CONSTRUCTION

---

ALL 4 OF THE POSSIBILITIES FOR CONSTRUCTION REFERRED TO IN SECTION [II] ABOVE WILL BE DEALT WITH. \( \text{E 0 R} \) = END OF RECORD. \( \text{E 0 F} \) = END OF FILE.

JC = JOB CONTROL CARDS.

P1 = PROGRAM SORT2.

P2 = PROGRAM STERI02.

P3 = PROGRAM DRAW.

D = DATA CARDS.

S = STEERING CARDS.

THUSO-

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ADDITIONAL FEATURES AVAILABLE IN THE LIBRARY.

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THERE ARE TWO MORE PROGRAMS AVAILABLE IN THE LIBRARY.

STERI03 ... PRODUCING A LINE PRINTER OUTPUT

STERI05 ... PRODUCING MICROFILM OUTPUT

THEY ARE USED TO PLOT AND ROTATE DATA OBTAINED FROM THREE RIGHT-ANGLE SECTIONS CUT THROUGH A ROCK. EG. 'C AXIS' FABRIC DATA.

ONLY ONE ROCK CAN BE DEALT WITH PER RUN OF THE PROGRAM.

THE PROGRAMS ARE USED THUS O-

NOTE -- (+++) CARD NEEDED FOR PAPER OUTPUT.

--- (***+) CARD NEEDED FOR MICROFILM OUTPUT.
THEN COMES THE DATA THUS 0-
FIRST THE STEERING CARDS.
STARTING IN COLUMN 1 0 -
3
0
(INTEGER (I4) THE NUMBER OF READINGS TO BE USED IN THIS
DATA SET) RIGHT ADJUSTED .
12111:TITLE OF DATA UP TO 55 CHARACTERS LONG
* THE NO. OF ROTATIONS THIS DATA SET IS TO UNDERGO.
A MAX OF 3. THUS *=0,1,2 OR 3.
THE NEXT CARD HAS THE ROTATION DATA IN 9F5.1 FORMAT
DECLINATION OF ROTATION AXIS / DIP OF ROTATION AXIS /
AMOUNT OF ROTATION (CLOCKWISE ONLY) / ETC. UP TO 3 TIMES.IE
230. 90. 124. 360. 30. 555. 042. 20. 80.
THE NEXT CARDS ARE THE DATA CARDS WITH THE DATA ON
THEM IN 5 SETS OF DIRECTION OF DIP / AMOUNT OF DIP.
FORMAT 10F4.0
WHEN THE DATA IS FINISHED FOR THIS FACE TYPE 1000 IN
THE NEXT SPACE MEANT FOR THE DIRECTION OF DIP.
(THE DATA IS RIGHT ADJUSTED IN THE FORMAT SPACE).
THE NEXT TWO FACES NEED THE SAME SORT OF DATA SET STARTING
AT THE CARD MARKED ............. ABOVE
NO MORE THAN THREE DATA SETS CAN BE HANDLED IN A RUN
IE. 1 ROCK A RUN.
THE OUTPUT IS 4 NETS, DATASET1,2,3; THEN NET 4 WHICH
IS THE SUM OF THE ROTATED FORMS OF 1+2+3; BUT IS ITSELF
UNROTATED ABOVE THE ROTATIONS THAT HAVE GONE INTO 1,2,3.
THE DECK IS TERMINATED BY AN £ 0 R> CARD, FOLLOWED BY
AN £ 0 F> CARD.
THE PROGRAMS ARE A FURTHER MODIFIED FORM OF DR OWENS
PROGRAM (MODIFIER,F.W.M.H)
MISCELLANEOUS.
THE PROGRAM IS AT PRESENT SET TO TAKE 5000 DATA CARDS.
INTX/INTY THESE SHOULD NOT BE TOUCHED
IWS.
THE PROGRAM WILL CALCULATE TO EQUAL AREA OR ANGLE OUTPUT.
IWS=1 FOR =AREA (PRESENT SETTING)
IWS=2 FOR =ANGLE
THE PROGRAM ON FILE AT BEDFORD IS IN Binary FORM-IE. COMPILED.
NOTE ...........

THE CIRCLE DESCRIBED AROUND THE CONTOURS IS NOT THE TRUE
PRIMITIVE. IT COINCIDES WITH THE PRIMITIVE NEAR THE CARDINAL POINTS, BUT ELSEWHERE LIES SLIGHTLY OUTSIDE IT.
DATA FALLING EXACTLY ON THE PRIMITIVE IN THESE AREAS CAN BE RECOGNISED AS THEY WILL BE EXACTLY MIRRORED ACROSS THE NET, AND THE CONTOURS WILL CONVERGE TO FORM A STEEP "CLIFF" ON THE OUTER SIDE.
DATA Plotting near the primitive will not be mirrored, and will not have such a steep "CLIFF".
WHERE ONLY SOME OF THE DATA FALL ON THE PRIMITIVE, THIS PORTION CAN BE DETECTED BY LOOKING FOR MIRRORED PEAKS AT THE EDGE, WHICH ARE AT THE SAME "HEIGHT" ACROSS THE NET. THIS EDGE EFFECT IS AN UNAVOIDABLE PART OF (DIMFILM) CONTOURING ROUTINES.

NOTE also... contour labels are suppressed if the contours are close; values for them may be determined by counting "Up slope" from the "lowest" which will always be (0.50).

WHEN PAPER OUTPUT IS REQUESTED, 10' DIAMETER NETS FILLED WITH ALPHA-NUMERICS ARE PRINTED ONTO THE LINE PRINTER. THEY HAVE A KEY PRINTED WITH THEM AND ARE THUS SELF EXPLANATORY.
IT IS RECOMMENDED THAT INITIALLY WHEN LOOKING AT YOUR NETS ONLY PAPER OUTPUT IS USED. THEN, WHEN YOU ARE SURE YOU WANT THEM, RUN THE JOB TO PRODUCE MICROFILM.

XXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXXX
PROGRAM SORT2 (DATA, OUTPUT, INPUT, SCRATCH, TAPE10=DATA,
- TAPE6=OUTPUT, TAPE5=INPUT, TAPE7=SCRATCH)

DATA SORTING PROGRAM

ALL OUTPUT SUPPRESSED. PROGRAM WRITES TO TAPE 7, EXCEPT WHEN
AN ERROR IS FOUND IN THE DATA INPUT.

F.W.M. HOPPER.

SET UP ARRAY DIMENSIONS.

INTEGER CDEC (5000), CDIP (5000), DCEC (5000), DCIP (5000), DADEC (5000),
- DADIP (5000).

INTEGER DEC, PO, TESN, A, GRID, PASS, B, C, D, E, DEVIDE

INTEGER SYMB (17), EDOCN (17), NR (17), NPC (17)

DIMENSION IDENT (5000), XCORD (5000), YCORD (5000), NS (5000), IDEN (5000)

DIMENSION TITLE (17, 5)

DIMENSION RAMA (17, 4)

COMMON IDENT, XCORD, YCORD, NS, IDEN, CDEC, CDIP, DCEC, DCIP, DADEC, DADIP

LEVEL2, IDENT, XCORD, YCORD, NS, IDEN, CDEC, CDIP, DCEC, DCIP, DADEC, DADIP

C

C READ IN CO-ORDINATE DATA. STEERING.

DO 1 I = 1, 30
READ (5, 2) (CORD (I, J), J = 1, 4)
WRITE (6, 1111)

1111 FORMAT (1HN)

IF (EOF (5) .NE. 0.0) GO TO 50

2 FORMAT (F5.2, 1X, F5.2, 1X, F5.2, 1X, F5.2)

1 CONTINUE

C

C D = THE SIZE OF SYMB ARRAY, = THE SIZE OF THE EDOCN ARRAY

D = 17

C READ IN THE REST OF THE STEERING DATA.

DO 3 I = 1, D
READ (5, 4) (TITLE (I, J), J = 1, 6)
IF (EOF (5) .NE. 0.0) GO TO 6050

4 FORMAT (5A10, A4)

3 CONTINUE

DO 34 I = 1, D
READ (5, 5) (RAMAX (I, J), J = 1, 4)
IF (EOF (5) .NE. 0.0) GO TO 6050

5 FORMAT (4A5)

34 CONTINUE

DO 35 I = 1, D
READ (5, 5) (SYMB (I))
IF (EOF (5) .NE. 0.0) GO TO 6050

6 FORMAT (A2)

35 CONTINUE

DO 36 I = 1, D
READ (5, 7) (EDOCN (I))
IF (EOF (5) .NE. 0.0) GO TO 6050

7 FORMAT (I1)

36 CONTINUE

DO 40 I = 1, D
READ (5, 41) (NR (I))
IF (EOF (5) .NE. 0.0) GO TO 6050

41 FORMAT (I1)

40 CONTINUE

DO 50 I = 1, D
READ (5, 51) (NPC (I))
IF (EOF (5) .NE. 0.0) GO TO 6050

51 FORMAT (I1)

50 CONTINUE

READ (5, 50) PASS, DEVIDE
IF (EOF (5) .NE. 0.0) GO TO 6050
DO 71 I=1,3
READ (5,10) (DECR(I),DIPR(I),AMR(I)).
71 CONTINUE
IF EOF(5).NE.0.0) GO TO 6050
C
C CHECK TO SEE IF NEXT CARD IS <E 0 R> . IF NOT
C STOP AND SAY SO . TOO MANY STEERING CARDS.
READ (5,9000) DUMMY
9000 FORMAT(A1)
IF EOF(5).NE.0.0) GO TO 7000
WRITE (5,8000)
8000 FORMAT('1 ',25X,' YOU HAVE TOO MANY STEERING CARDS , CHECK THEM ')
C TO STOP ENTIRE JOB. I CALL NON EXISTANT ROUTINE.
CALL ABNORMAL
STOP
C <E 0 R> WAS REACHED TOO SOON WHEN READING IN THE STEERING CARDS.
C WRITE THIS FACT OUT.
6050 WRITE (5,5000)
5000 FORMAT('1 ',25X,' YOU HAVE TOO FEW STEERING CARDS , CHECK THEM ')
C TO STOP ENTIRE JOB. I CALL NON EXISTANT ROUTINE.
CALL ABNORMAL
STOP
7000 CONTINUE
C
C FIXED PARAMETERS AND COUNTERS.
AN=0
TEN=510
JOR=N=0
INTX=1
INTY=2
IR=1
IWS=1
NSET=D
GRID=30
DEC=1000

C
C READ IN STRUCTURAL DATA.
D0 8 P0=1,5000
READ (10,9) (NS(P0),DADEC(P0),DADIP(P0),IDENT(P0),XCORD(P0),
-YCORD(P0))
9 FORMAT(6X,A1,1X,I3,1X,I2,1X,A2,1X,F5.2,1X,F5.2)
IF EOF(10).NE.0.0) GO TO 10
C CHECK FOR <EOF> AND COUNT NO OF DATA CARDS.
NSET=NSET+1
8 CONTINUE
10 CONTINUE
C
C SEARCH FOR GROSS PUNCHING ERRORS
CALL ERROR (IDENT,DADIP,DADEC,SYMB,P0,D)
C
C JUMP INTO LOOP IF GRID SEARCHING IS NOT REQUIRED.
IF <PASS.NE.10) GO TO 15
C
C GRID SEARCHING LOOP CYCLE =30
11 DO 12 A=1,GRID
AA=A
K=0
C SEARCH USING ALL DATA AND ONE GRID CARD.
D0 13 B=1,NSET
C IF CO-ORDINATE CARD BLANK, SKIP TO END OF LOOP.
  IF(CORD(A,2).EQ.-0.0) GO TO 12
  IF(XCORD(B).LE.CORD(A,1).OR.XCORD(B).GE.CORD(A,2)) GO TO 13
  IF(YCORD(B).LE.CORD(A,3).OR.YCORD(B).GE.CORD(A,4)) GO TO 13
  K=K+1
C CONDITIONS SATISFIED FILL ARRAYS.
  DCCE(K)=DADEC(B)
  DCIP(K)=DADIP(B)
  IDEN(K)=IDENT(B)
13 CONTINUE
C LEAP OVER THE NON CO-ORDINATE BASED ARRAY FILLER ROUTINE.
  GO TO 14
C
C15 K=0
C FILL ARRAYS.
  DO 16 C=1,NOSET
  K=K+1
  DCCE(K)=DADEC(C)
  DCIP(K)=DADIP(C)
  IDEN(K)=IDENT(C)
  CONTINUE
C
C START THE COMPARATIVE AND ASSIGNMENT ROUTINES.
14 DO 17 NUM=1,NOSET
  CALL SET(NUM,D,SYMB,IDEND)
C SETS UP THE IDEND TO TEST FOR A CERTAIN IDENTIFIER
C
C START TESTING
  CALL TEST(K,IDEND,IDENT,CDECE,CDIP,CDECE,CDIP,J,DEVIDE,NS)
C
C IF TEST HAS FOUND NOTHING GO TO NEXT CYCLE OF THE LOOP.
  IF(J.EQ.0) GO TO 17
C
C FIND THE CORRECT NCODE FOR THE SYMB FUNCTION
  CALL FINE(D,IDEND,SYMB,NUM,NCODE,EDOCN)
C
C MAKE SURE IDEND IS CLEARED OR IT WILL REPEAT THE LAST LOOP.
  IDEND=2H
C
C START THE COMPLEX PROCESS OF WRITING OUT PROCESSED DATA.
  IF(JOHN.GT.1) GO TO 20
C WRITE TOTAL NO OF DATA SETS TO BE EXPECTED USING CO-ORDINATES.
  IF(PASS.EQ.10) WRITE(7,21)TESN
C WRITE NORMAL (17) NO OF DATA SETS TO BE EXPECTED.
  IF(PASS.NE.10) WRITE(7,21)NOSET
21 FORMAT(I3)
20 JOHN=JOHN+1
C IF CO-ORDINATE CARDS WERE USED GO AND PRINT WHICH ONE.
  IF(AA.NE.0) GO TO 30
C NO CO-ORDINATES WERE USED PRINT 0.
  WRITE(7,27)E
C LEAP OVER CO-ORDINATE PRINTING BIT.
  GO TO 29
C PRINT THE NO OF THE CO-ORDINATE CARD IN USE.
30 WRITE(7,27)A
29 CONTINUE
27 FORMAT(I3)
C WRITE TOTAL NO OF DATA TO BE PRINTED.
  WRITE(7,19)J
FORMAT(14)
C PRINT THE TITLE CARD FOR THE NEXT PROGRAM WITH ALL PARAMETERS
C RELEVANT TO THIS CYCLE OF THE LOOPS.
WRITE(7,22) (INTX,INTY,MCODE,NPC(NUM),IWS,NR(NUM),(TITLE(NUM,I),
-I=1,6),(RAMAX(NUM,I),I=1,4))

FORMAT(911,5A10,A4,4A5)
C WRITE OUT THE TITLE CARD FOR THE NEXT PROGRAM WITH ALL PARAMETERS
C RELEVANT TO THIS CYCLE OF THE LOOPS.
WRITE(7,85) (DECR(II),DIPR(II),ANR(II),II=1,3)

FORMAT(6F5.1)
C WRITE OUT THE ROTATION DATA IF NR IS NOT 0
IF (NR(NUM).EQ.0) GO TO 84
WRITE(7,85) (DECR(II),DIPR(II),ANR(II),II=1,3)

FORMAT(1014)
23 CONTINUE
17 CONTINUE
12 CONTINUE
C WRITE TOTAL NO OF DATA CARDS.
WRITE(7,25)NDSET
C PREPARE TAPE (SCRATCH) FOR NEXT PROGRAM.
REWIND 7
REWIND 5
REWIND 10
STOP
END

C
C
C
C THIS ROUTINE CHECKS THE DATA FOR 'IMPOSSIBLE ERRORS', TELLS
C YOU WHAT THEY ARE AND STOPS THE PROGRAM.
SUBROUTINE ERROR(IDENT,DADIP,DADEC,SYMB,P0,D)
INTEGER DADIP(P0),DADEC(P0),SYMB(D),P0,D
DIMENSION IDENT(P0)
LOGICAL FOUND
NO=1
IP=P0-1

1112 FORMAT(IHZ)
WRITE(5,34)
34 FORMAT('1',25X,'ERRORS IN YOUR DATA (IF ANY) WILL BE LISTED BELOW'

DO 31 I=1,IP
FOUND=.FALSE.,
DO 32 J=1,D
IF (IDENT(I).NE.SYMB(J))GO TO 32
FOUND=.TRUE.
32 CONTINUE
IF (.NOT.FOUND.OR.DADEC(I).GT.360.OR.DADIP(I).GT.90) WRITE(6,33)
-I,IDENT(I),DADEC(I),DADIP(I)
IF (.NOT.FOUND.OR.DADEC(I).GT.360.OR.DADIP(I).GT.90) NO=3
33 FORMAT(25X,'CARD ',I4,5X,' IDENT= ',A2,5X,' DADEC= ',I3,5X,
-DADIP=' ,I2)
IF (NO.NE.3) WRITE (6,35)
FORMAT (5(/),25X,'WELL DONE - NO ERRORS WERE FOUND IN YOUR DATA.')
IF (NO.NE.3) RETURN
WRITE (6,100)
FORMAT (0',10X,'THE PROGRAM WAS TERMINATED BECAUSE YOUR DATA HAD THE ABOVE ERROR(S) IN IT. TERMINATION OCCURRED IN SUBROUTINE ERROR.
='5/10X,'1')
WRITE (6,1112)
CALL ABNORM
STOP
END
C
SUBROUTINE SET(NUM,D,SYMB,IDEND)
C SETS IDEND TO THE CORRECT IDENT
C
INTEGER SYMB(D),D
IDEND=SYMB(NUM)
RETURN
END
C
SUBROUTINE TEST(K,IDEND,IDEN,CDEC,CDIP,DCEC,DCIP,J,DEVIDE,NS)
C TESTS THE IDEND AGAINST THE IDENT AND STOPS IT IF CORRECT
C
INTEGER CDEC(K),CDIP(K),DCEC(K),DCIP(K),DEVIDE
DIMENSION IDEN(K),NS(5000)
LEVEL2,CDEC,CDIP,DCEC,DCIP,NS,IDEN
J=0
C SET DEVIDE TO ITS DEFAULT VALUE OF 1
IF (DEVIDE.EQ.-0.0) DEVIDE=1
C CREATE THE ARRAYS OF SORTED DATA.
DO 18 I=1,K
IF (DEVIDE.EQ.1) GO TO 1
IF (DEVIDE.EQ.2) GO TO 2
IF (NS(I).NE.1HN) GO TO 18
GO TO 1
2 IF (NS(I).NE.1HS) GO TO 18
1 CONTINUE
IF (IDEN(I).NE.IDEND) GO TO 18
J=J+1
CDEC(J)=DDEC(I)
CDIP(J)=DCIP(I)
18 CONTINUE
RETURN
END
C
SUBROUTINE FINDE(D,IDEND,SYMB,NUM,NCODE,EDOCN)
C FINDS THE NCODE CORRESPONDING TO THE IDEND
C
INTEGER SYMB(D),EDOCN(D),D
IF (IDEND.EQ.SYMB(NUM)) NCODE=EDOCN(NUM)
RETURN
END
IDENT ABNORM
SUBROUTINE ABNORM
THIS ROUTINE IS SPECIFIC TO THE ULCC.7600, AND WILL ABORT THE WHOLE JOB SHOULD IT BE CALLED ENTRY ABNORM
ABNORM BSS 1 .ENTRY/EXIT
MESSAGE STARS,,1
MESSAGE MESSAGE,,1
MESSAGEF STARS,,1
ABORT,NO DUMP
STARS DATA
MESSAGE DATA ERROR IN YOUR DATA, SEE OUTPUT FILE FOR DETAILS
END
PROGRAM STE102 (INPUT, OUTPUT, FILMOUT, TAPE5=INPUT, TAPE6=OUTPUT, -TAPE8=FILMOUT)

C PROGRAM SET FOR A 66 LINE/PAGE PRINTER BUT WITH ALL OUTPUT
C SUPPRESSED EXCEPT THAT WRITTEN TO TAPE 8 THIS GOES TO THE
C NEXT PROGRAM FOR FLOATING.

INTEGER A, TND

DIMENSION ALPH(27), STER(81), TITLE (55), S(3,3), DCR(3), ARAN(2)
DIMENSION SR(3,3), ST(3,3), DCR(3), DIPR(3), ANR(3)
DIMENSION ND(49,41), DADEC(10), DADIP(10), NND(49,41)

C DIMENSION COSINE ARRAYS

C DIMENSION QL(65,41), QM(65,41), QN(65,41)

DATA ALPH (1), ALPH (2), ALPH (3), ALPH (4), ALPH (5), ALPH (6), ALPH (7)
DATA (8), ALPH (9), ALPH (10), ALPH (11), ALPH (12), ALPH (13), ALPH (14)
DATA (15), ALPH (16), ALPH (17), ALPH (18), ALPH (19), ALPH (20)
DATA (21), ALPH (22), ALPH (23), ALPH (24), ALPH (25), ALPH (26)
DATA (27)/
DATA (1), IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR, IHR
DATA (2), ARAN (1), ARAN (2)/


C DATA IX1/31/, RX1/31.52/, IX2/65/, RX2/1.48/, RX3/64.52/, RX4/1.25/, IX3/33/

C FOR 6 LI/IN USE 23 23.62 49 1.38 48.62 1.6667 25
C 8 31 31.52 65 1.48 64.52 1.25 33

PI = 3.1415926536
DEGTOR = PI/180.0
INTXL=0
INTYL=0
IWSL=0
READ (5,487) NSET
WRITE (8,487) NSET

487 FORMAT(I3)

C DO 71 ISET=1,NSET
READ (5,344) A
IF (EOF (5), NE.0.0) GO TO 7

344 FORMAT(I3)
READ (5,343) TND
IF (EOF (5), NE.0.0) GO TO 7

343 FORMAT(I4)
READ (5,14) INTX, INTY, NCODE, NPC, IWS, NR, (TITLE(I), I=1,55)
RAMAX = IF (INTY .LT .2) INTY=2

14 FORMAT(6I1,55A1,4F5.2)
IF (INTY, L.T.2) INTY=2

C PRELIMINARIES FOR SET
C CANORM IS COSINE OF ANGLE, ALPH, OF CONE OF SAMPLING (I.E. SEMI-ANGLE)
C FOR 1 PERCENT AREA (OR SOLID ANGLE) CANORM=0.99
C CANORM=1.0-0.01*FLOAT(NPC)
SANCORM=SQR(1.0-CANORM)+CANORM
C FORM SIN(ALPHA+5) - SEE BELOW
C SAS=SANCORM*CS(36.0*DEGTO)
C RTGR=SANCORM*SR(36.0*DEGTO)
LTOT=0
C CALCULATE ARRAY LIMITS FOR ND
NIX=IX1/INTX
NIY=39/INTY
C ND IS ND(2*NIX+3,2*NIY+3)
C CENTRE IS (NIX+2,NIY+2)
C CALCULATE LOCAL GRID LIMITS
LGY=IFIX(FLOAT(NIY)*SANCORM+1.5)
4/19

LGX=IFIX(FLOAT(NIX) \#SANORM,1.5)
C CLEAR ND, NUMERICAL OUTPUT ARRAY
201 N2X=2\#NIX+3
   N2Y=2\#NIY+3
   DO 1030 I=1,N2X
   DO 1030 J=1,N2Y
1030 ND(I,J)=0
C BUILD DIRN COSINE ARRAYS - SKIP IF ALREADY DONE
   IF (INTX.EQ.INTX.AND.INTY.EQ.INTY) GO TO 1032
   IF (INTX.EQ.INTX.AND.INTY.EQ.INTY) GO TO 1033
   DO 1031 I=1,N2X
   DO 1031 J=1,N2Y
   X=FLOAT( (NIX+2-I) \#INTX ) /RX1
   Y=FLOAT( (J-NIY-2) \#INTY ) /39.37
   CALL ERECT(X,Y,IWS,DCL,DCM,DCN)
   QL(I,J)=DCL
   QM(I,J)=DCM
   QN(I,J)=DCN
1032 CONTINUE
1031 QN(I,J)=DCN
1031 WRITE(8,342)A
1032 WRITE(8,727)NR
5012 FORMAT(15H1INPUT DATA - , 55A1,3X,9HSTEERING ,511,3X,8HRA\NMAX = , 1F5.2/) \C IF DATA ARE TO BE ROTATED, CALCULATE ROTATION MATRIX, S.
   IF (NR.EQ.0) GO TO 1009
   READ (5,1107) (DECR(IR),DIPR(IR),ANR(IR),IR=1,3)
   1107 FORMAT(9F5.1)
   727 FORMAT(I1)
   DO 1100 I=1,3
   DO 1100 J=1,3
   S(I,J)=0.0
1100 S(I,J)=1.0
   DO 1102 IR=1,3
   CALL ROTDAT(DECR(IR),DIPR(IR),ANR(IR),SR)
   CALL P\NMAT(SR,S,ST)
   DO 1103 I=1,3
   DO 1103 J=1,3
   S(I,J)=ST(I,J)
1102 CONTINUE
1100 CONTINUE
1009 CONTINUE
C READ AND PROCESS DATA
1007 READ (5,1001) (DADEC(I),DADIP(I),I=1,5)
   1001 FORMAT(10F4.0)
   5013 FORMAT(5(F10.2,F7.2))
   D0 1003 I=1,5
   I2END=1
   ENDEC=DADEC(I)
   IF (ENDEC.EQ.1000.0) GO TO 1002
   IF (ENDEC.EQ.2000.0) GO TO 1007
   IF (DADIP(I).GE.0.0) GO TO 800
   DADIP(I)=-DADIP(I)
   ENDEC=ENDEC+180.0
   IF (ENDEC.GT.360.0) ENDEC=ENDEC-360.0
   800 CONTINUE
   LDE=LTOT+1
   ENDEC=DEGTOR\#ENDEC
   IF (NC\CODE.LT.4) GO TO 1500
C PLAGIARISED DATA
   X=DADIP(I) \#COS(ENDEC) /RAMAX
   Y=DADIP(I) \#SIN(ENDEC) /RAMAX
   IW52=NC\CODE-3
   CALL ERECT(X,Y,IW52,DCL,DCM,DCN)
   GO TO 2030
C OTHER CASES - SHOW EMPTY AREA WITH NO Grid - 1
1000 ENDIF.
ENDIF.

4001 IF (DCLA.GT.0.0) GO TO 1002
1001 ENDIF.

4030 IF (IC.NE.0) GO TO 3000
C ROTATE DATA IF REQD. TAKE OTHER END OF AXIS IF OUT OF LR HEMIS.
   DO 3001 IC=1,3
3001 DCR(IC)=S(INDIP*IC)+DCMA+5
DCMA=DCMA*(IC.2)
DCN=DCN*(IC.3)
   IF (DCN.GE.0.0) GO TO 3000
2003 DCNA=-DCNA
DCLA=-DCLA
DCMA=-DCMA
3000 CONTINUE
C FIND LOCAL ZERO NEAREST ND LOCN TO DATA POINT
   CALL PJECT(DCLA,DCMA,DCNA,IWS,XCGMP,YCQMP)
   ILZ=-IFIX(FLOAT(NIX)+XC0MP+0.5)
   JLZ=IFIX(FLOAT(NIY)+YCOMP+0.5)
C SCAN OVER LOCAL GRID
   ILL=ILZ-LGX
   ILU=ILZ+LGX
   JLL=JLZ-LGY
   JLU=JLZ+LGY
   IF (ILL.LT.1) ILL=1
   IF (JLL.LT.1) JLL=1
   IF (ILU.GT.N2X) ILU=N2X
   IF (JLU.GT. N2Y) JLU=N2Y
   DO 2001 IL=ILL,ILU
      DO 2001 JL=JLL,JLU
C FIND DIRN COSINES OF ARRAY LOCN - FIND CORRESP TRUE SPACE POINT
C AND CALL ERECT
   DCL=QL(IL,JL)
   DCM=QM(IL,JL)
   DCN=QN(IL,JL)
C IGNORE GRID POINTS IF MORE THAN 5 DEGREES BEYOND PRIMITIVE
   IF (DCN.LT.-0.087) GO TO 2001
C ADD ONE TO CELL IF APPROPRIATE
   IF (DCL+DCM+DCN.LT.CANORM) ND(IL,JL)=ND(IL,JL)+1
2001 CONTINUE
C CHECK WHETHER POINT LIES WITHIN ALPHA+5 OF PERIMETER - IF SO, REPEAT
C CALCULATION FOR OTHER END OF AXIS
   IF (DCNA.GT.SA5.OR.I2END.LT.0) GO TO 1003
   I2END=-1
   GO TO 2003
1003 CONTINUE
GO TO 1007
C WRITE TABLE OF CALCULATED VALUES
1002 WRITE(8,2014)(TITLE(I),I=1,55)
2014 FORMAT(IH ,55A1)
   WRITE(8,2113)INTY,INTX,NPC,LTOT,ARAN(IWS)
2113 FORMAT(IH ,' SAMPLING ',I3,' ACRROSS/ ',I3,' DOWN ',15X,' POLES PER - ',12,' PERCENT AREA ',5X,14,' DATA ',5X,5)$
2004 FORMAT(IH0,55A1//12H SAMPLING - ',I3,' ACROSS/ ',I3,' 5H DOWN',5X,14,' DATA ',5X,55A1)
15X: 10 HPOLES PER 12: 13H PERCENT AREA, 5X: 14.5H DATA, 5X: A6
5001 FORMAT(12HORATED BY ,F6.1, 13H ABOUT DEC = ,F6.1, 18H DIP = ,
16F6.1)
7000 FORMAT(13H ROTATION MATRIX ,3F10.4/(19X,3F10.4))
CALL CENT(ND, TND, NND,
3777 FORMAT(12H)
DO 3333 IX = 1, N2X
WRITE(B,3777) N2X
WRITE(B,3777) N2Y
3333 WRITE(B,2006) (NND(IX, IY), IY = 1, N2Y)
2006 FORMAT(11H, 41I3)
C ALPHANUMERIC OUTPUT
342 FORMAT(1H , ' USING CORDINATE CARD ', 13)
2013 FORMAT(60X,1HN)
IX=1
DO 2007 ISX=1, IX2
C CLEAR STER AND PUT IN LIMITING STARS
DO 2008 ISY=1, 81
2008 STER(ISY) = ALPH(1)
CSQ = (FLOAT(ISY) - RX2) - (RX3 - FLOAT(ISX))
IF (CSQ.LT.0.0) GO TO 2020
IDIFY= IFIX(RX4 - SQRT(CSQ) + 0.5)
IYL'R = 41-IDIFY
IYUR= 41+IDIFY
STER(IYLR)=STAR
STER(IYUR)=STAR
C IDENTIFY CORRESPONDING LOCNS OF STER AND ND
2020 IF (ISX.NE. (IX-NIX-2) * INTX+IX3) GO TO 2009
DO 2010 IY=1, N2Y
2010 CONTINUE
IX = IX + 1
IF (ISX.EQ. 1X3) STER(ISX) = ALPH(24)
IF (ISX.EQ.1X3) STER(81)=ALPH(6)
2011 FORMAT(20X,81A1)
2009 CONTINUE
2007 CONTINUE
902 FORMAT(60X,1HS/
910 FORMAT(25X,50H
910 FORMAT(25X,50HKEY A=1 B=2 C=3 D=4 E=5 F=11-12
g=13-14 h=15-16 i=17-18 j=19-20 k=21-22
l=23-24 m=25-26 n=27-28 o=29-30 p=31-32
q=33-34 r=35-36 s=37-38 t=39-40 u=41-42
v=43-44 w=45-46 x=47-48 y=49-50 z=50+
% , REPRESENTS A VALUE, GT.0,BUT.LT.1 PERCENT %)
C 910 FORMAT(25X,50HKEY A=1 B=2 C=3 D=4 E=5 F=6 G=7 H=8 I=9
C 119HJ=10 K=11 L=12 M=13//30X,30HN=14 O=15 P=16 Q=17 R=18 S=19
C 235HT=20 U=21 V=22 W=23 X=24 Y=25 Z=26+
C STORE PARAMETERS FOR COMPARISON WITH NEXT SET
INTX=INTX
INTY=INTY
IWSL=IWS
71 CONTINUE
C
READ (5,333) NDSET
333 FORMAT (14)
8107 FORMAT (20X,14X,13)
C
7 CONTINUE
REWIND 8
REWIND 5
STOP
END
SUBROUTINE PJECT(DCL,DCM,DCN,IWS,X,Y)
C
GIVEN DEC,DIP IN RADIANS CALCULATES PROJECTED POINT (X UP,Y RIGHT)
AS FRACTION OF PLOTTING CIRCLE RADIUS, FOR EQUAL AREA (IWS=1) OR
EQUAL ANGLE (IWS=2).
SIND=ABS(DCN)
IF(IWS.EQ.1) GO TO 1
R=SQRF((1.0-SIND)/(1.0+SIND))
GO TO 2
1 R=SQRF(1.0-SIND)
2 IF (DCM.EQ.0.0.AND.DCL.EQ.0.0)  GO TO 3
DEC=ATAN2(DCM,DCL)
GO TO 4
3 DEC=0.0
4 X=RYCOS(DEC)
Y=RYSIN(DEC)
RETURN
END
SUBROUTINE ERECT(X,Y,IWS,DCL,DCM,DCN)
C FROM COORDINATES X,Y (FRACTIONS OF PLOTTING CIRCLE RADIUS) CALCULATES
DIRECTION COSINES ASSUMING EQUAL AREA (IWS=1) OR EQUAL ANGLE (IWS=2)
PROJECTION.
R=SQRF(X*X+Y*Y)
IF (R.GE.1.41) RETURN
IF (R.LT.1.0E-9) GO TO 1
IF (IWS.EQ.1) DCN=1.0-R*R
IF (IWS.EQ.2) DCN=COS(2.0*ATAN2(R,1.0))
STH0R=SQRF(1.0-DCNYDCN)/R
DCL=XYSTH0R
DCM=Y'STH0R
RETURN
1 DCL=0.0
DCM=0.0
DCN=1.0
RETURN
END
SUBROUTINE ROTDAT(DECNR,DIPR,ANGROT, S )
C CALCULATES ROTATION MATRIX,S.
DIMENSION S(3,3),R(3,3),T(3,3)
DEGTOR=0.01745329252
DO 3000 1=1,3
DO 2999 J=1,3
2999 S(I,J)=0.0
3000 SCI, I )  =1.0
DO 3001 IRPT=1,5
1 ANG=-DEGTORYDIPR
IF (IRPT.NE.3) GO TO 3002
1=2
ANG=-DEGTORYDECR
IF(IRPT.EQ.5) ANG=DEGTORYDIPR
3002 ANG=-DEGTOR%ANGROT
DO 3003 K=1,3
DO 3005 L=1,3
3006 R(K,L)=0.0
3003 R(K,K)=1.0
R(I,I)=COS(ANG)
R(J,J)=R(I,I)
R(I,J)=-SIN(ANG)
R(J,I)=-R(I,J)
DO 3005 K=1,3
DO 3005 L=1,3
T(K,L)=0.0
DO 3005 M=1,3
3005 T(K,L)=T(K,L)+R(K,M)*S(M,L)
DO 3004 K=1,3
DO 3004 L=1,3
3004 S(K,L)=T(K,L)
3001 CONTINUE
RETURN
END

SUBROUTINE P3MAT(XM1,XM2,XMP)
C FORMS MATRIX PRODUCT (XMP) = (XM1) (XM2).
DIMENSION XM1(3,3),XM2(3,3),XMP(3,3)
DO 1 I=1,3
DO 1 J=1,3
XMP(I,J)=0.0
DO 1 K=1,3
1 XMP(I,J)=XM1(I,K)*XM2(K,J)+XMP(I,J)
RETURN
END

SUBROUTINE CENT(ND,TND,NND,N2X,N2Y)
C CENT CONVERTS ALL COMPONENTS OF ARRAY TO 0/0 OF TOTAL NO OF DATA
C FED INTO THE CONSTRUCTION OF THE ARRAY.
INTEGER TND
DIMENSION ND(49,41),NND(49,41)
DIMENSION ND(65,41),NND(65,41)
USE (49) FOR 6 LI/IN AND (65) FOR 8 LI/IN
DO 1 I=1,N2X
DO 2 K=1,N2Y
NND(I,K)=ND(I,K)*100)/TND
IF (NND(I,K).EQ.0).AND.(ND(I,K).NE.0)NND(I,K)=-1
2 CONTINUE
1 CONTINUE
RETURN
END

C C C
PROGRAM STER103 (INPUT, OUTPUT, FILMUT, TAPE5=INPUT, TAPE6=OUTPUT, TAPE8=FILMUT)
C PROGRAM SET FOR A 66 LINE/PAGE PRINTER, BUT WITH ALL OUTPUT
C SUPPRESSED EXCEPT THAT WRITTEN TO TAPE 8. THIS GOES TO THE
C NEXT PROGRAM FOR PLOTTING.
C THE PROGRAM IS MODIFIED TO RUN 3 SETS OF CRYSTALLOGRAPHIC DATA
C ROTATE THESE ONTO A SINGLE PLANE AND SUM THEM.
C THE OUTPUT IS 4 PLOTS, 1 FOR EACH SET AND THE SUM LAST.
C IE. THE PROGRAM HANDLES UNIVERSAL STAGE DATA
INTEGER A, TND, TND1
DIMENSION ALPH(27), STER(81), TITLE(65), S(3,3), DCR(3), ARAN(2)
DIMENSION SR(3,3), ST(3,3), DIPR(3), ANR(3)
DIMENSION ND(49,41), DADEC(10), DADIP(10), NND(49,41), NNND(49,41)
DIMENSION ND(65,41), DADEC(10), DADIP(10), NND(65,41), NNND(61,41)
C USE ND(49 FOR 6 LI/IN, ND(65 FOR 8 LI/IN
C DIMENSION COSINE ARRAYS
DIMENSION QL(55,41), OM(55,41), QN(55,41)
DATA ALPH(l), ALPH(2), ALPH(3), ALPH(4), ALPH(5), ALPH(6), ALPH(7),
1 ALPH(8), ALPH(9), ALPH(10), ALPH(11), ALPH(12), ALPH(13), ALPH(14),
2 ALPH(15), ALPH(16), ALPH(17), ALPH(18), ALPH(19), ALPH(20),
3 ALPH(21), ALPH(22), ALPH(23), ALPH(24), ALPH(25), ALPH(26), ALPH(27),
4 IH, 1HA, 1HB, 1HC, 1HD, 1HE, 1HF, 1HG, 1HH, 1HI, 1HJ, 1HK, 1HM, 1HN, 1HO,
51HP, 1HQ, 1HR, 1HS, 1HT, 1HU, 1HV, 1HW, 1HX, 1HY, 1HZ,
6ARAN(1), ARAN(2), SHEQ AR, SHEQ AN/
DATA IX1/23/, RX1/23.52/, 1X2/49/, RX2/1.38/, RX3/48.52/, RX4/1.6667/, IX3/25/
DATA IX1/31/, RX1/31.52/, IX2/55/, RX2/1.48/, RX3/64.52/, RX4/1.25/, IX3/33/
C FOR 6 LI/IN USE 23 23.52 49 1.38 48.52 1.6667 25
C 8 31 31.52 65 1.48 64.52 1.25 33
PI=3.1415926536
DEGTOR=PI/180.0
INTXL=0
INTYL=0
IWSL=0
TND1=0
III=0
READ (5,487) NSET
NSET1=NSET+1
WRITE (8,487) NSET1
487 FORMAT(13)
C C
C DO 71 ISET=1,NSET
READ (5,344) A
IF(EOF(5) .NE.0.0) GO TO 7
344 FORMAT(13)
READ (5,343) TND
IF(EOF(5) .NE.0.0) GO TO 7
343 FORMAT(14)
READ (5,14) INTX, INTY, NCODE, NPC, IWS, NR, (TITLE(I) , I=1,55),
1 RAMAX
IF(EOF(5) .NE.0.0) GO TO 7
14 FORMAT(S11,5F4.2)
IF(INTY.LT.2) INTY=2
C PRELIMINARIES FOR SET
C CNORM IS COSINE OF ANGLE, ALPHA, OF CONE OF SAMPLING (I.E. SEMI-ANGLE)
C FOR 1 PERCENT AREA (OR SOLID ANGLE) CNORM=0.99
C CANORM=1.0-0.01*FLOAT(NPC)
C SANDR=SQR(1.0-CANORM*CANORM)
C FORM SIN (ALPHA+5) - SEE BELOW
C SAS=SANDR*PCOS (35.0*DEGTOR) +CANORM*PSIN (35.0*DEGTOR)
LQ7=0
C CALCULATE ARRAY LIMITS FOR ND
   NIX=IX1/INTX
   NIY=39/INTY
C ND IS ND(2*NIY+3,2*NIX+3)
C CENTRE IS (NIX/2,NIY/2)
C CALCULATE LOCAL GRID LIMITS
   LGY=IFIX((FLOAT(NIY)+SANOR)+1.5)
   LGX=IFIX((FLOAT(NIX)+SANOR)+1.5)
C CLEAR ND, NUMERICAL OUTPUT ARRAY
201 N2X=2*NIX+3
   N2Y=2*NIY+3
   DO 1030 I=1,N2X
   DO 1030 J=1,N2Y
      IF (III.GT.O) GO TO 1030
      NNND(I,J)=0
1030 ND(I,J)=0
II=II+1
C BUILD DIRN COSINE ARRAYS - SKIP IF ALREADY DONE
   IF(IWS.NE.IWSL) GO TO 1033
   IF(INTX.EQ.INTXL.AND.INTY.EQ.INTYL) GO TO 1032
1033 DO 1031 I=1,N2X
   DO 1031 J=1,N2Y
      X=FLOAT((NIX+2-I)*INTX)/RX1
      Y=FLOAT((J-NIY-2)*INTY)/39.37
      CALL Erect(X,Y,IWS,DCL,DCM,DCN)
      QL(I,J)=DCL
      QM(I,J)=DCM
      QN(I,J)=DCN
1032 CONTINUE
C IF DATA ARE TO BE ROTATED, CALCULATE ROTATION MATRIX,S.
   IF(NR.EQ.0) GO TO 1009
   READ (5,1107) (DECIR(IR),DIPR(IR),ANR(IR),IR=1,3)
   1107 FORMAT(9F5.1)
   DO 1101 1=1,3
      DO 1101 J=1,3
         S(I,J)=0.0
      S(I,I)=1.0
   DO 1102 IR=1,3
      CALL RODAT(DECIR(IR),DIPR(IR),ANR(IR),SR)
      CALL PROMAT(SR,S,ST)
   DO 1103 1=1,3
      DO 1103 J=1,3
         S(I,J)=ST(I,J)
   1102 CONTINUE
1009 CONTINUE
C C READ AND PROCESS DATA
1007 READ (5,1001) (DADEC(I),DADIP(I),I=1,5)
1001 FORMAT(15F4.0)
   5013 FORMAT(5(F10.2,F7.2))
   DO 1003 I=1,5
      I2END=1
      ENDEC=DADEC(I)
      IF(ENDEC.EQ.1000.0) GO TO 1002
      IF(ENDEC.EQ.2000.0) GO TO 1007
      IF(DADIP(I).GE.0.0) GO TO 800
      DADIP(I)=-DADIP(I)
      ENDEC=ENDEC+180.0
   IF(ENDEC.GT.360.0) ENDEC=ENDEC-360.0
CONTINUE
L105=L105+1
ENDEC=DEGTOR+ENDEC
IF (NCODE.LT.4) GO TO 1500

C PLAGIARISED DATA
X=DADIP(I)*COS(ENDEC)/RAMAX
Y=DADIP(I)*SIN(ENDEC)/RAMAX
I1S2=NC0DE-3
CALL ERECT(X, Y, I1S2, DCLA, DCMA, DCNA)
GO TO 2030

C OTHER CASES - CHANGE TO CONFORM WITH NCODE = 1
1500 ENDEC=DEGTOR*PI
1502 CONTINUE
1503 ENDEC=ENDEC+PI
1504 ENDEC=ENDEC+0.5*PI
1505 ENDEC=ENDEC+2.0*PI
ENDIP=0.5*PI-ENDEC
GO TO 2030

C CALCULATE DIRECTION COSINES
DCLA=COS(ENDIP)*COS(ENDEC)
DCMA=COS(ENDIP)*SIN(ENDEC)
DCNA=SIN(ENDIP)

2030 IF (NR.EQ.0) GO TO 3000

C ROTATE DATA IF REQD. TAKE OTHER END OF AXIS IF OUT OF LR HEMIS.
DO 3001 IC=1,3
3001 DCR(IC)=S(IC,1)*DCLA+S(IC,2)*DCMA+S(IC,3)*DCNA
DCLA=DCR(1)
DCMA=DCR(2)
DCNA=DCR(3)
IF (DCNA.GE.0.0) GO TO 2030

3003 DCNA=-DCNA
DCLA=-DCLA
DCMA=-DCMA
3000 CONTINUE
C FIND LOCAL ZERO - NEAREST ND LOCN TO DATA POINT
CALL PJECT(DCLA,DCMA,DCNA,I1S2,XCOMP,YCOMP)
ILZ=-IFIX(FLOAT(NIX) *XCOMP+0.5)+NIX+2
JLZ=IFIX(FLOAT(NIY) *YCOMP+0.5)+NIY+2

C SCAN OVER LOCAL GRID
ILL=ILZ-LGX
ILU=ILZ+LUX
JLL=JLZ-LGY
JLU=JLZ+LUY
IF (ILL.LT.1) ILL=1
IF (JLL.LT.1) JLL=1
IF (ILL.GT.N2X) ILL=N2X
IF (JLL.GT.N2Y) JLL=N2Y
DO 2001 IL=ILL,ILU
DO 2001 JL=JLL,JLU

C FIND DIRN COSINES OF ARRAY LOCN - FIND CORRESP TRUE SPACE POINT
DCL=QL(IL,JL)
DCM=QM(IL,JL)
DCN=QN(IL,JL)
IF (DCN.LT.-0.087) GO TO 2001

2003 DCN=-DCN
DCL=-DCL
DCM=-DCM
2001 CONTINUE
C CHECK WHETHER POINT LIES WITHIN ALPHA+5 OF PERIMETER - IF SO, REPEAT
C CALCULATION FOR OTHER END OF AXIS
IF (DCN.GT.SA5.OR.I2END.LT.0) GO TO 1003
I2END=-1
GO TO 2003
1003 CONTINUE
GO TO 1007
C WRITE TABLE OF CALCULATED VALUES
1002 WRITE (8,2014) (TITLE (I), I=1,55)
2014 FORMAT (1H55A1)
WRITE (8, 2113) INTY, INTX, NPC, L, TOT, ARANDWS
2113 FORMAT (1H SAMPLING ACROSS / -13, DATA /5X, A6)
5001 FORMAT (12H ORATED BY , 8.1, 10H ABOUT DEC = , 8.1, 1F5.1)
7000 FORMAT (19H ROTATION MATRIX , 3F10.4/d9X, 3E10.4))
I > ^  (III. LE. 3 )
C call CENT (ND, TND, NND, N2X, N2Y)
IF (III.EQ.3) GO TO 7675
6767 CONTINUE
IF (III.EQ.3) III = 8
TND1 = TND1 + TND
IF (III.EQ.8) WRITE (8, 342) A
IF (III.EQ.8) NR = 0
IF (III.EQ.8) WRITE (8, 2113) INTY, INTX, NPC, TND1, ARANDWS
DO 7579 NAX = 1, N2X
DO 7578 NAY = 1, N2Y
NNND (NAX, NAY) = NNND (NAX, NAY) + ND (NAX, NAY)
7578 CONTINUE
7579 CONTINUE
IF (III.EQ.8) CALL CENT1 (NNND, TND1, NND, N2X, N2Y)
DO 7676 CONTINUE
C DO 2005 IX = 1, N2X
WRITE (8, 3777) N2X
WRITE (8, 3777) N2Y
3777 FORMAT (12)
DO 3333 IX = 1, N2X
WRITE (8, 3333) IX = 1, N2Y
3333 WRITE (8, 3333) (NNND (IX, IY), IY = 1, N2Y)
2006 FORMAT (1H10)
2015 FORMAT (1H1)
C ALPHANUMERIC OUTPUT
342 FORMAT (1H, ' USING COORDINATE CARD ', 13)
2013 FORMAT (1H, 60X, 1H N)
IX = 1
DO 2007 IX = 1, IX2
C CLEAR STAR AND PUT IN LIMITING STARS
2008 STER (ISY) = ALPH (1)
CSG = (FLOAT (ISX) - RX2) * (RX3 - FLOAT (ISX))
IF (CSQ.LT.0.0) GO TO 2020
IDIFY = IFIX (RX4/SQRT (CSQ) + 0.5)
IYLR = 41 - IDIFY
IYUR = 41 + IDIFY
STER (IYLR) = STAR
STER (IYUR) = STAR
C IDENTIFY CORRESPONDING LOCNS OF STER AND ND
2020 IF (ISX.NE.((IX-NIX-2)*INTX+IX3)) GO TO 2009
DO 2010 IY = 1, N2Y
ISY = 41 + (IY-NIY-2)*INTY
IF (ISY.LT.1.0R.OR.ISY.GT.8.1) GO TO 2010
C SET CORRECT ALPHANUMERIC VALUE IN STER
C ID = ND (IX, IY) + 1
C IF (ID.GT.27) ID = 27
ID = (NND (IX, IY) + 1)/2 + 1
IF (ID.GT.27) ID = 27
IF (ID.EQ.0) ID=1
STER(ISY) = ALPH(ID)
IF(NND(X,Y).EQ.-1) STER(ISY) = 1H.

2010 CONTINUE
IX=IX+1
IF(ISX.EQ.1X3) STER(1) = ALPH(24)
IF(ISX.EQ.1X3) STER(81) = ALPH(6)

2011 FORMAT(20X,B1A1)
2009 CONTINUE
2007 CONTINUE
902 FORMAT(60X,1HS//)
910 FORMAT(25X,%NUMERICS = THE PERCENTAGE INTERVAL THE LETTERS COVER:)
  1/25X * A=1-2 B=3-4 C=5-6 D=7-8 E=9-10 F=11-12 *
  1/25X * G=13-14 H=15-16 I=17-18 J=19-20 K=21-22 *
  1/25X * Q=33-34 R=35-36 S=37-38 T=39-40 U=41-42 *
  1/25X * V=43-44 W=45-46 X=47-48 Y=49-50 Z=50+ *
  1/25X *, REPRESENTS A VALUE.GT.0, BUT.LT.1 PERCENT *%

C 910 FORMAT(25X,50HKEY A=1 B=2 C=3 D=4 E=5 F=6 G=7 H=8 I=9 ,
C 119J=10 K=11 L=12 M=13/30X,N=14 O=15 P=16 Q=17 R=18 S=19 ,
C 255T=20 U=21 V=22 W=23 X=24 Y=25 Z=26+,
C STORE PARAMETERS FOR COMPARISON WITH NEXT SET
INTXL=INTX
INTYL=INTY
IW=SL=IWS
IF(IWS.EQ.3) GO TO 6767
71 CONTINUE
C
READ(5,333) NDSET
333 FORMAT(I4)
8107 FORMAT(20X,I4,5X,I3)
C
7 CONTINUE
REWIND 8
REWIND 5
STOP
END

SUBROUTINE PJECT(DCL,DCM,DCN,IWS,X,Y)
C GIVEN DEC,DIP IN RADIANS CALCULATES PROJECTED POINT (X UP,Y RIGHT)
C AS FRACTION OF PLOTTING CIRCLE RADIUS, FOR EQUAL AREA (IWS=1) OR
C EQUAL ANGLE (IWS=2).
SIND = ABS(DCN)
IF(IWS.EQ.1)  GO TO 1
R = SQRT(1.0-SIND)/(1.0+SIND)
GO TO 2
1 R = SQRT(1.0-SIND)
2 IF(DCM.EQ.O.0.AND.DCL.EQ.O.0) GO TO 3
DEC = ATAN2(DCM,DCL)
GO TO 4
3 DEC = 0
4 X = R * COS(DEC)
Y = R * SIN(DEC)
RETURN
END

SUBROUTINE ERECT(X,Y,IWS,DCL,DCM,DCN)
C FROM COORDINATES X,Y (FRACTIONS OF PLOTTING CIRCLE RADIUS) CALCULATES
C DIRECTION COSINES ASSUMING EQUAL AREA (IWS=1) OR EQUAL ANGLE (IWS=2)
C PROJECTION.
R = SQRT(X**2+Y**2)
IF(R.GE.1.41) RETURN
IF(R.LT.1.0E-9) GO TO 1
IF(IWS.EQ.1) DCM = 1.0-R**2
IF(IWS.EQ.2) DCM = COS(2.0*ATAN2(R,1.0))
STHOR = SQRT(1.0-DCM**2)/R
DCL=X*STHBR
DCM=Y*STHBR
RETURN

1
DCL=0.0
DCM=0.0
DCN=1.0
RETURN
END

SUBROUTINE RODAT(DECN,DIPN,ANGR,O,S)
  C CALCULATES ROTATION MATRIX,S.
  DIMENSION S(3,3),R(3,3),T(3,3)
  DEGR=0.01745329252
  DO 3000 I=1,3
  DO 2999 J=1,3
  2999 S(I,J)=0.0
  3000 S(I,I)=1.0
  DO 3001 IRPT=1,5
  1 =
  ANG=-DEGR*DIPN
  IF (IRPT.NE.3) GO TO 3002
  1 =2
  ANG=-DEGR*ANGR
  3002 J=3
  IF (IRPT.EQ.1.OR.IRPT.EQ.5) J=2
  IF (IRPT.EQ.1) ANG=-DEGR*DECR
  IF (IRPT.EQ.5) ANG=DEGR*DECR
  IF (IRPT.EQ.4) ANG=DEGR*DIPN
  DO 3003 K=1,3
  DO 3004 L=1,3
  3004 S(K,L)=T(K,L)
  3001 CONTINUE
  RETURN
END

SUBROUTINE PROMAT(XM1,XM2,XMP)
  C FORMS MATRIX PRODUCT (XMP) = (XM1)(XM2),
  DIMENSION XM1(3,3),XM2(3,3),XMP(3,3)
  DO 1  I=1,3
  DO 1  J=1,3
  1  XMP(I,J)=XM1(I,K)*XM2(K,J)+XMP(I,J)
  RETURN
END

SUBROUTINE CENT(ND,TND,NND,N2X,N2Y)
  C CENT CONVERTS ALL COMPONENTS OF ARRAY TO 0/0 OF TOTAL NO OF DATA
  FORMED INTO THE CONSTRUCTION OF THE ARRAY.
  INTEGER TND
  DIMENSION ND(49,41),NND(49,41)

C
C DIMENSION ND(65,41), NND(65,41)
C USE(49) FOR 6 LI/IN AND (65) FOR 8 LI/IN
DO 1 I=1,N2X
DO 2 K=1,N2Y
NND(I,K) = (ND(I,K) ^ 100) / TND
IF(NND(I,K) .EQ. 0) .AND. (ND(I,K) .NE. 0) NND(I,K) = -1
2 CONTINUE
1 CONTINUE
RETURN
END

C SUBROUTINE CENT1(NNND, TND1, NND, N2X, N2Y)
C THIS PERCENTAGES THE SUMED MATRIX NND WHICH IS
C RUN 1+2+3= THE OUTPUT NET NO 4
INTEGER TND1
C DIMENSION NNND(65,41), NND(65,41)
DIMENSION NNND(49,41), NND(49,41)
C USE (49) FOR 6 LI/IN, AND (65) FOR 8 LI/IN.
DO 1 I=1,N2X
DO 2 K=1,N2Y
NND(I,K) = (NNND(I,K) ^ 100) / TND1
IF(NND(I,K) .EQ. 0) .AND. (NNND(I,K) .NE. 0) NND(I,K) = -1
2 CONTINUE
1 CONTINUE
RETURN
END
PROGRAM DRAW(INPUT, OUTPUT, TAPE3=INPUT, TAPE4=OUTPUT)
C THE PURPOSE OF THIS PROGRAM IS TO CONSTRUCT A PAGE CONTAINING A
C CIRCLE, IN WHICH IS FITTED A TWO DIMENSIONAL ARRAY (ND), HOLDING
C THE DATA FROM PROGRAM STERIO.
C THIS ARRAY IS THEN CONTOURED AT PRESET VALUES.
C THESE VALUES CAUSE CONTOURS AT 0-50 PERCENT INTERVALS FROM 0-50
C PERCENT, WITH INCREASING CONTOUR DENSITY WITH ASCENT.
C THE PROGRAM READS FROM A TAPE CREATED BY STERIO, AND USES INFORMATION
C FROM THIS, AS WELL AS ARRAYS OF DATA WITHIN ITSELF TO CONSTRUCT TITLING,
C AND A KEY FOR THE PAGE, WHICH IS DRAWN BY THE PROGRAM ON FILM. THUS
C THE PROGRAM MUST RUN AT ALL WITHOUT MODIFICATION; MUST RUN WITH STERIO,
C WHICH IN TURN IN ITS PRESENT FORM MUST RUN WITH SORT2. .........
C SET UP DIMENSIONS
INTEGER X(3) , TITLE (6)  , STRR(10) , N(1) , NND(49,41)
INTEGER S(1) , E(1) , W(1) , SP(1) , PS(1) , NRS(2) , N(1) , N(1)
REAL N D (41,49), CVALS(27)
DATA N D/41,49/, CVALS(27)
DIMENSION NND(49,41)
C COMMON BLOCK WITH A DIMFILM SUBROUTINE
C CONTROLLING THE SIZE OF CONTOUR PLOTTING
C DATA STATEMENTS CONTAINING LABELING THAT
C DO NOT NEED TO CHANGE INSIDE THE PROGRAMME.
CALL DATA(NRR, 'THIS PLOT HAS UNDERGONE - ',3)
CALL DATA(NRS, '- ROTATIONS - ',2)
DATA N/IHN/
DATA S/IHS/
DATA E/2H-E/
DATA W/2HW-/
DATA SP/IHT/
DATA PS/IH./
C ARRAY HOLDING CONTOUR LEVEL VALUES.
DATA CVALS/-0.5,0.9,2.9,4.9,6.9,8.9,
-10.9,-12.9,14.9,16.9,18.9,20.9,
-22.9,-24.9,26.9,28.9,30.9,32.9,
-34.9,-36.9,38.9,40.9,42.3,44.9,
-46.9,-48.9,50.9/
C DATA FOR THE KEY.
CALL DATA(GUFA, 'THE CONTOURS REPRESENT THE FOLLOWING PERCENTAGES '
-5)
CALL DATA(GUFB, '0.0 <= 1-290 , 0.9 <= 3-400 , 0.49 <= 5-600 , 0.69 <= 7-800 ;
0.89 <= 9-100 , 10.9 <= 11-120 , 12.9 ',9)
CALL DATA(GUFC, '12.9 <= 13-160 , 14.9 <= 15-160 , 16.9 <= 17-180 , 18.9 <=
19-200 <= 21-220 <= 22.9 ',8)
CALL DATA(GUFD, '22.9 <= 23-240 <= 24.9 <= 25-260 <= 26.9 <= 27-280 <= 28.9 <=
29-300 <= 30.9 <= 31-320 <= 32.9 ',8)
CALL DATA(GUFE, '32.9 <= 33-340 <= 34.9 <= 35-360 <= 36.9 <= 37-380 <= 38.9 <=
39-400 <= 40.9 <= 41-420 <= 42.9 ',8)
CALL DATA(GUFF, '42.9 <= 43-440 <= 44.9 <= 45-460 <= 46.9 <= 47-480 <= 48.9 <=
49-500 <= 50.9 <= 510 ',8)
CALL DATA(GUFG, 'NOTE A VALUE OF 0.5-0.9 <= .LT.10 AND .GT.00 ',5)
C ARRAY CONTROLLING LEVELS OF CONTOUR INTENSITY.
DATA ICVALS/10,11,12,13,14,15,16,17,18,19,20,21,22,23,24,25,26,27,28,29,30/
C READ IN NUMBER OF DATA SETS TO BE DEALT WITH.
READ (3,1) NSET
1 FORMAT(I3)
C SET UP FOR PLOTTING ON DIMFILM.
CALL CAM35MM
A(1)=1.0
A(2)=0.0
CALL CINE
CALL BOUNDS(0.0,100.0,0.0,150.0)
CALL SYMTXT(1.66666666)
DO 2 I=1,NSET
  C START READING DATA FROM PREVIOUS PROGRAMME.
  README(3,3) (X(J),J=1,3)
  IF (EOF(3),NE.0) GO TO 4
  FORMAT(1X,2A10,A7)
  README(3,5) NR
  FORMAT(A1)
  README(3,6) (TITLE(J),J=1,6)
  FORMAT(1X,5A10,A5)
  README(3,7) (STRR(J),J=1,10)
  FORMAT(1X,9A10,A8)
  READ(3,8) N2X
  READ(3,8) N2Y
  FORMAT(12)
  N3X=N2X
  N3Y=N2Y
  C READ IN MATRIX OF DATA TO BE CONTOURED.
  DO 9 IX=1,N2X
  READ(3,10) (NND(IX,IY),IY=1,N2Y)
  FORMAT(1X,4113)
  9 CONTINUE
  C CONVERT IT TO A REAL ARRAY, STAND IT ON END,
  C AND TURN IT UPSIDE DOWN. SO THAT IT FITS THE CINE
  C MODE OF THE FILM AND IS STILL CORRECT N,S,E, AND W.
  N2X1=N2X+1
  DO 11 P=1,N2X
  DO 12 K=1,N2Y
  NND(K,P)=NND(N2X1-P,K)
  C EQUATION TO GATHER 0-1 PERCENT LEVEL.
  IF (NND(K,P).EQ.-1) NND(K,P)=0.51
  12 CONTINUE
  11 CONTINUE
  C FIND LARGEST CIRCLE THAT CAN BE FITTED INTO A
  C MATRIX N3X,N3Y. NOTE, N3X CAN BE <=,>=, N3Y
  C ALL VALUES OUTSIDE ARE THEN SET TO EXTREME VALUES TO
  C FORCE INTERPOLATION DURING CONTOURING TO REMAIN
  C WITHIN THE CIRCLE.
  CALL ZIRCLE(ND,N3Y,N3X,-10.0)
  CALL OUTLINE
  C START LABELING
  CALL OFFTOXY(35.5,134.0)
  CALL SYMTXT(X,27)
  CALL OFFTOXY(31.0,138.0)
  CALL SYMTXT(NRR,25)
  CALL OFFTOXY(56.25,138.0)
  CALL SYMTXT(NR,1)
  CALL OFFTOXY(57.5,138.0)
  CALL SYMTXT(NRS,11)
  CALL WINDOW(21.5,77.5,142.0,148.0)
  CALL WINDFRM
  CALL ENDC
  CALL OFFTOXY(22.0,144.5)
  CALL HVYTYPE
  CALL UNLIN
  CALL SYMTXT(TITLE,55)
  CALL NOUNLIN
  CALL NRMTYPE
  CALL OFFTOXY(1.0,130.0)
  CALL SYMTXT(STRR,98)
  CALL OFFTOXY(50.0,124.5)
  CALL SYMTXT(N,1)
  CALL OFFTOXY(50.0,123.0)
DRAW KEY BOX AND KEY.
CALL WINDOW(10.0,90.0,2.0,24.0)
CALL WINDOW(0.6,99.4,27.1,122.8)
C
TURN ON INTERPOLATION, A CUBIC EQUATION.
CALL CINTERP(10)
C
SET CONTOURING LABEL SIZE.
CALL CLABHT(0.5,0)
C
CONTOUR.
CALL CONT1(ND,N3Y,N3X,27,CVALS,ICVALS)
C
TURN EVERYTHING OFF.
CALL NOCINT
CALL ENDWIND
CALL NEWFRAM
2
CONTINUE
4
CONTINUE
CALL ENDFILM
STOP
END
SUBROUTINE DATA(ARRAY,STRING,LENGTH)
C
PUTS DATA STRINGS INTO 10H WORD BLOCKS.
IMPLICIT INTEGER(A-Z)
DIMENSION ARRAY(LENGTH), STRING(LENGTH)
IF (LENGTH.GT.0) GO TO 10
CALL DISPLA('SILLY WARNING - SILLY ARRAY SIZE: ', LENGTH)
```

RETURN

10 DO 20 I=1,LENTH
   IF (STRING(I).EQ.0) RETURN
20   ARRAY(I)=STRING(I)
   IF (STRING(LENTH+1).NE.0) CALL DISPLA
      '- ( "Warning: Attempt to Exceed Bounds", Lenth+1)
   C      NOTE DISPLA PRINTS OUT ONTO THE DAY FILE.
   RETURN
END

SUBROUTINE CIRCLE (ARRAY,NX,NY,SPECVAL)
   C GENERAL CIRCLE FITTING ROUTINE, ALTERS VALUES OUTSIDE
       C CIRCLE TO SPECIFIED VALUE.
   C IMPLICIT INTEGER (A-Z)
   REAL ARRAY(NX,NY)
   REAL XYRATIO,SPECVAL,R
   C FIND RADIUS IN TERMS OF Y
   R=NY/2.0
   IR=R+0.51
   C FIND RADIUS IN TERMS OF X
   IRX=NX/2.0 + 0.51
   IRSQ=R^2 + 0.51
   C FIND RATIO OF AXIS SCALES
   XYRATIO=FLOAT(NX)/FLOAT(NY)
   C CALC ODD/EVEN CENTERING ADJUSTMENTS
   IXCENT=0
   IF ((IRX+IRX).EQ.NX) IXCENT=1
   IYCENT=0
   IF ((IR+IR).EQ.NY) IYCENT=1
   DO 800 IY = 1 ,IR
      C FIND Y DISPLACEMENT FROM CENTRE
      ITABY=IR-IY
      C CALC X DISPLACEMENT FROM CENTRE
      ITABX=SQRT(FLOAT(IRSQ-(ITABY*ITABY)))*XYRATIO
      C FILL IN LEFT HAND SPECIAL VALUE
      ITAB=IRX-ITABX
      C JUMP IF CIRCLE REACHES LEFT BOUND
      IF (ITAB.LT.1) GO TO 350
      DO 300 IX = 1 ,ITAB
         C ABOVE EQUATOR
         ARRAY(IX,IR-ITABY)=SPECVAL
      C BELOW EQUATOR
      ARRAY(IX,IR+ITABY+IYCENT)=SPECVAL
   300 CONTINUE
   C FILL IN RIGHT HAND SPECIAL VALUES
   ITAB=IRX+ITABX+IXCENT
   C JUMP IF CIRCLE REACHES RIGHT BOUND
   IF (ITAB.GT.NX) GO TO 800
   DO 400 IX = ITAB,NX
      C ABOVE EQUATOR
      ARRAY(IX,IR-ITABY)=SPECVAL
   C BELOW EQUATOR
      ARRAY(IX,IR+ITABY+IYCENT)=SPECVAL
   400 CONTINUE
800 CONTINUE
RETURN
END
```
DOC SOURCE

FILEPROC = PROC (INREF CHARPUT) VOID;

BEGIN
  [ION] FILEPROC ARRAY;
  FOR I TO N DO ARRAY[I] = FILEPROC;
END;

REAL x, y; INT DIR, AMT;
PRINT ('FRAMING.....');
PRINT ((NEWLINE, "FRAME00=.CD257.16.ALGOL2", NEWLINE));
WHILE READ ((NEWLINE, 8 * SPACE, DIR, AMT, SPACE, C, X, Y));
  C = 'NF', B = 'SF';
  REAL A = 'NF', B = 'SF';
  PRINT ((NEWLINE, "FRAME00-.CD257.16.ALGOL2", NEWLINE));
  WHILE READ ((NEWLINE, 8 * SPACE, DIR, AMT, SPACE, C, X, Y));
  C = 'NF', B = 'SF';
  REAL A = 'NF', B = 'SF';
  PRINT ((NEWLINE, "FRAME00-.CD257.16.ALGOL2", NEWLINE));
  IF C = A OR C = B
    THEN OUTF (STANDOUT, 6X2D', 3D', 20.2D, 20.2DL, AMT, DIR, X, Y)
  END;
  PRINT ((NEWLINE, "FRAME00-.CD257.16.ALGOL2", NEWLINE))
END;
FINISH.
THE PLOTST EXTENSION, BLOCK - D -

This extension to the PLOTST library was written by -
F. W. M. Hopper and C. J. Cookson 1978

The library also utilises the ULCC. DIMFILM and MICROFILM routines, and the SYMAP and SYMVU packages, as available at ULCC.

Block -D-

This part of the PLOTST library is a set of fortran programs which take grid referenced foliation readings, measured as azimuth of dip and amount of dip. From them it constructs a three dimensional block diagram of the average foliation surface across the area from which the readings were taken.

The philosophy behind the package.

The data is taken, read in as binary (the default) or in a user specified format. (where the information is presented - dip, azimuth, x coordinate, y coordinate, x is along the bottom axis of the map with 0.0 in the bottom left corner.)

From this information, the apparent dip along the E-W grid lines are calculated and written off to a file for further use. Likewise for the apparent dips measured along the N-S grid lines.

Two files are therefore made (written in binary) with the information on them as - apparent dip (+ or -, depending on whether the dip was dominantly to the e=+,-w=-,n=+,-s=- the values ranging from +90.0 to -90.0 degrees), x coordinate and y coordinate, (one set of readings per line).

Next these two data files must be processed by a program that prepares them for SYMAP. This is necessary because SYMAP will only accept 1000 data points (so the data must be 'shrunk' if there are too many). This is done by fitting the data onto a coarser grid and averaging points that happen to fall on the same intersections.

SYMAP also expects the 0.0 position to be in the top left corner, and the x (first) coordinate to be the number of rows down from this. The program sorts all this out; the data are arranged in SYMAP format and to this data a set of default SYMAP steering instructions are added.

The user may override the program, requesting the data be 'shrunk' to less than 1000, he may also supply his own SYMAP instructions at this point.

Then, at this stage in the progress of running the library, there are still two data files; one of easterly apparent dips and one of northerly apparent dips, only now these files are in a form acceptable to SYMAP. (see ULCC. publications).
THE SYMAP PACKAGE IS NOW USED TO INTERPOLATE APPARENT DIP VALUES FOR GRID POSITIONS FOR WHICH NO INFORMATION WAS SUPPLIED. THIS IS DONE IN TURN FOR EACH DATA FILE.

IT IS IMPORTANT TO UNDERSTAND HOW SYMAP WORKS. IT MOVES TO EACH GRID INTERSECTION AND IF THERE IS NO VALUE FOR THAT INTERSECTION, IT LOOKS ROUND TO THE NEAREST 3 OR 4 FILLED INTERSECTIONS, AVERAGES THEIR VALUES AND STORES THE AVERAGE AS THE VALUE FOR THAT INTERSECTION. IT USES THIS VALUE IF NECESSARY FOR FUTURE AVERAGINGS.

SYMAP THUS TENDS TO MODEL A SOAP FILM (LYING) OVER A SET OF PEGS ARRANGED AT ABSOLUTE HEIGHTS DEPENDING ON THE VALUES OF THE 'READINGS' AT INTERSECTIONS WITH VALUES; IT CLINGS TO THE GIVEN VALUES IN THE SMOOTHEST POSSIBLE WAY, NEVER MOVING TO A VALUE ABOVE OR BELOW THEM, 'BETWEEN PEGS'.

WITH DATA IN THIS PROGRAM LIBRARY IT IS BEING ASKED TO INTERPOLATE BETWEEN APPARENT DIPS WHICH VARY BETWEEN 90.0 AND -90.0. THIS IT DOES FAITHFULLY, HOWEVER, THE NATURE OF AN APPARENT DIP MEANS THAT AS THE VALUE GIVEN ON ONE DATA FILE APPROACHES 90 THE CORRESPONDING VALUE ON THE OTHER APPROACHES 0.0. THIS IS FINE FOR SINGLE OR SIMPLE STRUCTURES WITH A LARGE WAVELENGTH RELATIVE TO THE MAP AREA, BUT IF THE STRUCTURES ARE SMALL SCALE AND THERE ARE A LOT OF THEM ACROSS THE MAP OR IF THERE IS AN INTERFERENCE PATTERN, THEN WE HAVE THIS SITUATION. CONSIDER, A SITUATION WHERE TWO ORTHOGONAL UPRIGHT FOLD TRAINS INTERFERE, THEN, THE DIP-AZIMUTH DATA WILL BE SUCH THAT WHEN THE APPARENT DIPS ARE CALCULATED ON A ROW OF READINGS WITH HIGH E-W COMPONENTS TO THEIR DIPS, (AND THEREFORE LOW N-S COMPONENT ON THE OTHER FILE), THERE MIGHT BE A POINT AT WHICH THERE IS A LOW E-W COMPONENT AND A CORRESPONDING HIGH N-S COMPONENT.

ON INTERPOLATION THE EFFECT OF THE ONE LOW (NEAR 0.0) VALUE ON THE FILE FULL OF HIGH VALUES WILL BE NEGLIGIBLE, BUT THE EFFECT OF THE ONE HIGH VALUE ON THE TAPE FULL OF LOW VALUES WILL BE MARKED.


THE PROBLEM NOW IS TO RECOMBINE THE APPARENT DIPS TO GENERATE THE ACTUAL SURFACE WHICH THEY REPRESENT. THIS PRESENTS A PROBLEM BECAUSE THE PROCESS OF INTERPOLATION, DESCRIBED ABOVE MEANS THAT THE DATA CANNOT NECESSARILY BE MERGED TO PRODUCE A UNIQUE OUTPUT; THE OUTPUT WILL VARY AS TO WHICH FILE IS DOMINANTLY USED TO CONSTRUCT THE OUTPUT. SEVERAL (5) PROGRAMS WERE WRITTEN TO RECOMBINE THE APPARENT DIP FILES, SO THAT THEY MAY BE LOOKED AT IN MANY POSSIBLE WAYS, THUS ALLOWING A FULLER UNDERSTANDING AS TO THE GEOMETRY OF THE STRUCTURES IN THE AREA. WHERE THE FORM OF THE FOLIATION SURFACE IS SIMPLE, ALL THESE PROGRAMS WILL GIVE THE SAME RESULT; IT IS ONLY WHERE THE INFORMATION ON THE TWO DATA FILES IS AFFECTED BY THE INTERPOLATION
AFFECT, THAT ANY DIFFERENCE IN OUTPUT IS SEEN.
WHAT THESE PROGRAMS DO IS TO SET ONE CORNER OF THE AREA TO A HEIGHT
OF 0.0. THEN STARTING FROM THIS THEY WORK OUT THE FIRST EDGE
OF THE AREA USING THE DATA FROM ONE FILE. (THE ALGORITHM IS-
ONE KNOWS THE SPACING BETWEEN GRID POINTS AND THE APPARENT DIP
AT THE POINT, SO USING PYTHAGORAS THE RELATIVE HEIGHT OF THE NEXT
POINT MAY BE CALCULATED).
NEXT THIS IS DONE USING DATA FROM THE OTHER FILE FOR ALL POINTS
ON THE AREA RUNNING AT RIGHT ANGLES TO THE FIRST EDGE. (THE CORRECT
E-W,N-S FILE USED IN EACH CASE).
ONE PROGRAM DOES A SIMILAR PROCESS ON THE NE DIAGONAL; HAVING
FIRST WORKED OUT TRUE DIP FROM BOTH APPARENT DIP FILES.
THE PRODUCT OF RUNNING THESE PROGRAMS IS ONE NEW DATA FILE THAT IS
TWO DIMENSIONAL MATRIX (129X129) OF ABSOLUTE HEIGHT VALUES.
THE OUTPUT IS IN A FORM EXACTLY LIKE THAT SYMAP WOULD PRODUCE.
THE NEW DATA FILE MUST NOW BE TURNED INTO A VISUAL PICTURE
THIS CAN BE DONE IN TWO WAYS.
1) BY PASSING IT TO SYMVU AND GETTING A THREE D BLOCK DIAGRAM ON
MICROFILM. IF THIS IS DONE THE USER MUST SUPPLY THE
SYMVU STEERING ELECTIVES.
2) BY PASSING IT TO YET ANOTHER PROGRAM IN THE LIBRARY (WHICH
WILL ACCEPT ANY SYMAP TYPE OUTPUT AND IS THEREFORE OF GENERAL
USE AS A GOOD ALTERNATIVE TO USUAL LINE PRINTED SYMAP OUTPUT).
THIS PROGRAM USES THE ULCC DIMFILM ROUTINES AND DRAWS A NORMAL
GEOGRAPHIC CONTOUR MAP OF THE SURFACE ONTO MICROFILM.

PROGRAM SPECIFICATIONS

THE APPARENT DIP GENERATOR-

SURFACE (DATAIN,NORDIP,EASDIP,INPUT,OUTPUT)
DATAIN- DATA FILE, EACH RECORD CONTAINS DIP, AZIMUTH, X, Y.
FORMAT TO READ DATA LOOKED FOR ON INPUT, IF NOT FOUND
IT DEFAULTS TO A BINARY READ.
NORDIP
EASDIP- BINARY OUTPUT FILES OF APPARENT DIPS (Respectively)
EACH RECORD CONTAINS X, Y, VALUE (REAL NOS)
FILESREWOUND AFTER WRITING.
INPUT - IF NO FORMAT DIRECTIVE IS TO BE SUPPLIED, SUBSTITUTE
A DUMMY FILE FOR INPUT, OR PUT TWO <EOR> CARDS TOGETHER
ON THE INPUT STREAM. IF A FORMAT IS TO BE SUPPLIED, A
CARD - ( FORMAT ) - SHOULD BE PLACED BETWEEN THEM.
NO PROGRAM EXPECTS REAL NOS.
OUTPUT- ERROR MESSAGES AND INFORMATION AS TO NUMBER OF RECORDS
READ ON DATAFILE, PASSED TO USER.

THE SYMAP PREPARING PROGRAM, THIS IS A GENERAL PURPOSE PROGRAM
AND MAY BE USED TO PREPARE ANY GEOGRAPHIC DATA FOR SYMAP IN
ISOLATION FROM THE REST OF THE LIBRARY, SO A USER FORMAT IS
SUPPLIED HERE TOO.
SYMPREP (TAPE1,TAPE2,INPUT,OUTPUT)
TAPE1 - SOURCE FILE FOR DATA, EACH RECORD CONTAINS X, Y, VALUE
FORMAT LOOKED FOR ON INPUT, BINARY DEFAULT.
SOURCE NOT REWOUND BEFORE OR AFTER READING.
TAPE1 SHOULD BE SUBSTITUTED FOR NORDIP OR EASDIP
WHEN RUNNING THE PACKAGE.
TAPE2 - SYMAP DATA FILE PRODUCED BY PROGRAM.
THE FILE SHOULD BE SUBSTITUTED FOR A DIFFERENT FILE NAME AS THE PROGRAM MUST BE RUN TWICE ONCE FOR EACH APPARENT DIP FILE.

TAPE2 IS REWOUND AFTER WRITING.

INPUT- READS DIRECTIVES FROM THE INPUT STREAM, MAY BE SUBSTITUTED FOR A DUMMY FILE IF DEFAULT PARAMETERS ARE TO BE USED, (AS WITH INPUT FOR SURFACE).

AVAILABLE DIRECTIVES-

1) LIMIT - DEFAULTS TO 1000, BUT MAY BE SET BETWEEN 1 AND 1000, CONTROLS THE NO. OF RECORDS PUT ONTO TAPE2.

2) FORMAT - DEFAULTS TO BINARY, BUT A CARD FORMAT (......), WILL CAUSE READING IN THAT FORMAT, REAL NOS.

3) F-MAP - A DEFAULT SET OF F-MAP DIRECTIVES ARE INCLUDED IN THE PROGRAM, THESE MAY BE CHANGED BY THE USER. SHOULD ONE BE CHANGED THEY MUST ALL BE SET.

THE USER SHOULD SUPPLY CARDS-

; ELECTIVES (SEE ULCC. SYMAP PUBLICATIONS)

99999 =REQUIRED TERMINATOR

OUTPUT- MESSAGES AND ERRORS ROUTED TO HELP USER.

THE INTERPOLATOR

SYMAP, SEE ULCC. PUBLICATIONS.

IF SYMAP IS CALLED WITH A DUMMY FILE AS THE SECOND FILE, THE MAP OUTPUT IS DIVERTED, THIS SAVES PAPER AS IT IS NOT NEEDED FOR THE RUNNING OF THE PROGRAM. THE SYMAP OUTPUT PRODUCED IS TAPE8. THIS IS IN SYMAP BINARY FORMAT.

TAPE 8 SHOULD BE REWOUND AND COPIED TO A LOCAL FILE THEN RETURNED (THE TAPE8 FILE NAME WILL BE REQUIRED LATER) SUGGESTED LOCAL FILE NAMES ARE INDIP, AND IEDIP, RESPECTIVELY THEY NEED NOT BE REWOUND.

RECOMBINATION OF THE TWO APPARENT DIP FILES.

THERE ARE 5 WAYS OF DOING THIS.

DIAG- WORKS ON N-E DIAGONAL WITH 0.0 IN BOTTOM LEFT.

NDEEPTh-WORKS SOUTH TO NORTH 0.0 IN BOTTOM LEFT.

SDEEPTh-WORKS NORTH TO SOUTH 0.0 IN TOP RIGHT.

WDEEPTh-WORKS EAST TO WEST 0.0 IN TOP RIGHT.

EDEPTH-WORKS WEST TO EAST 0.0 IN BOTTOM LEFT.

(INDIP, IEDIP, TAPE8, TAPE?, OUTPUT)

INDIP - IEDIP- APPARENT DIP FILES IN SYMAP FORMAT (SEE ULCC. PUB)

TAPE8- FINAL MATRIX FILE OF RECONSTRUCTED SURFACE; IN SYMAP TYPE BINARY FORMAT (REWOUND AFTER WRITING)

TAPE7- FILE FOR LOCAL USE WITHIN PROGRAM, SHOULD NOT BE TAMPERED WITH.

OUTPUT- MESSAGES TO USER, AS TO - NATURE OF THE RESULTS OF THE CALCULATIONS AND AS TO POSSIBLE PROBLEMS.

THE DRAWING OF THE BLOCK DIAGRAM

SYMMU, SEE ULCC. PUBLICATIONS (MICROFILM MUST BE ATTACHED AND LIBRARIED, SEE ULCC.PUB)

INPUT- READS TAPE8 IN SYMAP FORMAT, BUT ALSO REQUIRES A SET OF STEERING ELECTIVES SUPPLIED BY THE USER, THESE ARE PUT ON THE INPUT STREAM. THE USER IS REFERED TO THE ULCC. BULLETINS, BUT A SAMPLE SET IS SUPPLIED BELOW.

OUTPUT- TO MICROFILM + INFORMATION TO USER.
THE GEOGRAPHIC CONTOURING PROGRAM (AN ALTERNATIVE TO SYMVU)
THIS DRILLS A NORMAL CONTOUR MAP. IT IS A GENERAL PURPOSE
PROGRAM AND WILL HANDLE ANY SYMAP TYPE MATRIX. DIMFILM AND
MICROFILM MUST BE ATTACHED IN A LIBRARY FORM FOR OUTPUT TO
BE PRODUCED ON MICROFILM (SEE ULCC. PUBLICATIONS).

GRID (TAPE5, PARM, OUTPUT)
TAPE5- USED AS INPUT (SUBSTITUTE FOR SYMAP TAPEB, OR TAPEB
PRODUCED BY ONE OF THE '5' PROGRAMS).
TAPE5 ISREWOUND BEFORE AND AFTER READING.

PARM- THE WORD 'NOLABEL' OR 'LABEL'; SUBSTITUTED HERE WILL TURN
LABELING OF THE CONTOURS ON OR OFF RESPECTIVELY, THE DEFAULT
IS FOR LABELS TO BE PRODUCED.

OUTPUT-MICROFILM AND INFORMATION TO THE USER AS TO PROGRESS OF AND
ERRORS IN EXECUTION.

THERE IS AN ALTERNATIVE GRID PROGRAM WHICH ALLOWS THE USER TO
SPECIFY THE NUMBER AND LEVEL OF THE CONTOURS

GRID2 (TAPEB, INPUT, PARM, OUTPUT)
TAPEB- USED AS INPUT (THIS IS SYMAP TAPEB, OR TAPEB
PRODUCED BY ONE OF THE '5' PROGRAMS).
TAPEB IS NOT REWOUND IN THE PROGRAM

INPUT- THIS IS DEFAULTED TO READ FROM THE INPUT STREAM, ON
IT THERE SHOULD BE TWO CARDS: THE FIRST, STATING
AS AN INTEGER NUMBER (N) HOW MANY CONTOURS ARE TO BE
DRAWN, WHERE (N) < OR = 10.
THE SECOND CARD SHOULD CONTAIN (N) CONTOUR LEVELS, REAL
NOS. SEPERATED BY (,)S.
IF TAPEB HAS MORE THAN ONE MAP ON IT THESE CARDS MUST
BE SUPPLIED EACH TIME GRID2 IS CALLED, THE SETS OF CARDS
SEPERATED BY (EOR) CARDS.

PARM- THE WORD 'NOLABEL' OR 'LABEL', SUBSTITUTED HERE WILL TURN
LABELING OF THE CONTOURS ON OR OFF RESPECTIVELY, THE DEFAULT
IS FOR LABELS TO BE PRODUCED.

OUTPUT-MICROFILM AND INFORMATION TO THE USER AS TO PROGRESS OF AND
ERRORS IN EXECUTION.

TO RUN THE LIBRARY ONE MUST SET UP THE JOB CONTROL CARDS
AS IN THE SPECIMEN DECK BELOW.
NOTE. IT IS ASSUMED THAT ONLY THE FIRST LOT OF DATA IS NOT
IN BINARY FORMAT THUS ONLY ONE FORMAT IS SUPPLIED.
THE SPECIMEN SYMVU ELECTIVES WILL PRODUCE A SATISFACTORY
OUTPUT, THESE ARE AT THE BOTTOM OF THE JOB CONTROL CARDS.

JOB ........ J6.T59.M7600. ** YOUR NAME **
ATTACH.PLOTST.ID=LIBBED.
ATTACH.SVU,SYMVU,ID=PUBLIC.
ATTACH.SYMAP,ID=PUBLIC.
ATTACH.DIMFILM,ID=PUBLIC.
ATTACH.MFILM,MICROFILM,ID=PUBLIC.
LIBRARY,PLOTST,DIMFILM,MFILM.
SURFACE. GENERATES NORDIP AND EASDIP
SYMPREP,NORDIP,SYMORT,INPUT. GENERATES SYMNORT
SYMPREP,EASDIP,SYMEAST,INPUT. GENERATES SYMEAST
SYMAP,SYMORT,EBUSH. GENERATES TAPEB
REWIND,TAPEB.
COPY,TAPEB,INDIP.
RETURN, TAPE8.
SYMAP, SYMEAST, RUBBISH. Generates TAPE8
COPY, TAPE8, IEDIP.
RETURN, TAPE8.
Rewind, TAPE8.
WDEPTH. Generates TAPE8 (Use one of the '5' programs here)
GRID, TAPE8, LABEL.
REWIND, TAPE8.
LIBRARY, SVU, MFILM.
SYMVU.
<EOR>
(YOUR FORMAT SPECIFICS)
YOUR DATA IN THE ABOVE FORMAT
.
.
.
.
.<EOR>
LIMIT 1000
<EOR>
LIMIT 1000
<EOR>
A TITLE TO YOUR PLOT CAN BE PUT IN HERE ------------------------
129 129 2 4 2 4 1 6
55.0 45.0 6.0 5.0 -300.0 200.0
.<EOR>
.<EOR>
---------- GOOD LUCK ----------
PROGRAM SURFACE(DATAIN, NORDIP, EASDIP, INPUT, OUTPUT,
+ TAPE1=NORDIP, TAPE2=EASDIP, TAPE4=INPUT, TAPE5=DATAIN)

PROGRAM WRITTEN BY C. COOKSON AND F. W. M. HOPPER 1978.

A PROGRAM TO READ IN DIP AND AZIMUTH DATA, CALCULATE FROM THEM THEIR
NORTHERLY AND EASTERLY COMPONENTS, AND WRITE OUT THOSE COMPONENTS
(ALONG WITH THEIR X- AND Y-COORDINATES) TO THE FILES EASDIP AND
NORDIP.

FORMATS - THE DATA INPUT FILE WILL BY DEFAULT BE CALLED DATAIN,
AND IS THE FIRST FILE ON THE PROGRAM CARD. THE DATA
ON THAT FILE MUST CONTAIN 4 VALUES ON EACH RECORD (LINE), IN
THE ORDER - DIP, AZIMUTH, X-COORD, Y-COORD.
THese ARE ALL REAL NUMBERS.
A SINGLE RECORD READ FROM THE FILE 'INPUT' WILL BE USED
AS THE FORMAT SPECIFIER TO BE APPLIED TO THE INFORMATION ON
THE FILE 'DATAIN'. IF 'INPUT' IS EMPTY THE PROGRAM WILL
DEFAULT TO ASSUMING A FREE-FORMAT READ.
('INPUT' MAY BE MADE TO APPEAR EMPTY BY SUBSTITUTING THE NAME
OF A DUMMY FILE IN THE 4TH PARAMETER POSITION ON THE PROGRAM
CALL CARD).
READING WILL CONTINUE UNTIL END-OF-FILE IS ENCOUNTERED.

THE OUTPUT FILES EASDIP AND NORDIP WILL BE WRITTEN IN BINARY
FORMAT WITH THREE VALUES PER RECORD, IN THE ORDER
X, Y, EASTERLYDIP X, Y, NORTHERLYDIP
ON THE APPROPRIATE FILE.
BOTH OUTPUT FILES ARE REWOUND AT START AND FINISH OF THE
PROGRAM so THE DATA INPUT FILE IS NOT REPOSITIONED BEFORE OR
AFTER READING.
N.B. since NO ARRAYS ARE USED IN THIS PROGRAM (NO, NOT ONE) THERE
IS NO LIMIT ON THE SIZE OF THE DATA SET TO WHICH IT MAY BE
APPLIED - UNTIL YOU RUN OUT OF DISC SPACE, OF COURSE.

INTEGER FNAME, FORM(8)
LOGICAL FIXFORM

----- STATEMENT FUNCTIONS ------

SIND ( ARG ) = SIN ( ARG + 0.0175 )
COSD ( ARG ) = COS ( ARG + 0.0175 )
TAND ( ARG ) = TAN ( ARG + 0.0175 )
ATAND ( ARG ) = ATAN ( ARG ) * 57.143

MPTS = 0
PRINT 1
1 FORMAT('0',T22,17(''), 'PROGRAM SURFACE ',17(''))
FIXFORM = .T.

SEE IF THERE IS A FORMAT IN INPUT FOR US.
READ(4,7) FORM
FIXFORM = EOF(4).NE.0
AND ECHO IT IF THERE WAS.
IF (FIXFORM) PRINT 8, FORM
AND TELL HIM IF THERE WAS NOT
IF (.NOT.FIXFORM) PRINT 9
9 FORMAT('0',T22, 'DEFAULT FREE-FORMAT DATA ASSUMED. ')
GO.

7 FORMAT(8A10)
8 FORMAT('0',T22, 'INPUT FORMAT READ AS - ',8A10)

NPTS WILL COUNT THE NO OF POINTS...
READ (5, FORM) DIP, AZIMUTH, X, Y
IF (.NOT. FIXFORM) READ (5, *) DIP, AZIMUTH, X, Y
SEE WHERE TO STOP
READ (5, *) DIP, AZIMUTH, X, Y

DO THE SUMS ------FIRST NORTHERLY

DIPNORT = ATAND ( TAND ( DIP ) * COSD ( AZIMUTH ) )

THEN EASTERLY

DIPEAST = ATAND ( TAND ( DIP ) * SIND ( AZIMUTH ) )

OK.

NOW WRITE THEM TO OUT FILES...
WRITE (1) X, Y, DIPNORT
WRITE (2) X, Y, DIPEAST

AND ROUND FOR THE NEXT

GO TO 10

PRINT 4, NPTS

REWIND 1       REWIND 2

WELL - WE MAY AS WELL PRINT SOMETHING
STOP 'OK'

END
PROGRAM SYMMPREP (TAPE1, TAPE2, INPUT, OUTPUT, TAPE5=INPUT)

C
C WRITTEN BY C.J. COOKSON 1978.
C
C TO FUDGE A LARGE DATA SET (>1000 POINTS) INTO AN INPUT
C FORMAT ACCEPTABLE TO SYMAP.
C DATA MUST BE AS RECORDS OF THREE VALUES - X CO-ORD, Y CO-ORD,
C VALUE; ON THE FILE SUBSTITUTED FOR TAPE1.
C THE FORMAT OF THE DATA SHOULD BY DEFAULT BE BINARY, BUT IF REQUIRED
C A DIRECTIVE ON FILE INPUT MAY SPECIFY A FORTRAN FORMAT STATEMENT WHICH
C WILL BE USED INSTEAD FOR READING THE DATA.
C (SEE DOCUMENTATION)
C THE X AND Y CO-ORDS ARE EXPECTED IN *CONVENTIONAL* GEOGRAPHICAL
C RELATIONSHIP - I.E. X IS THE FIRST VALUE REPRESENTING HORIZONTAL
C DISPLACEMENT FROM THE LEFT EDGE OF THE AREA, Y IS VERTICAL
C DISPLACEMENT FROM THE BOTTOM EDGE. (SYMAP, OF COURSE, EXPECTS
C SOMETHING DIFFERENT ; THIS IS DEALT WITH BY THE PROGRAM.)

COMMON / FREDFl / DATA
COMMON / FREDFl2 / FLAGS
REAL DATA(191,191)
INTEGER FLAGS(191,191), FMAPDIR(73,8), INP(8), FMT(8)
INTEGER UNRECOG, FMAPNO
LOGICAL ENDFIL, FINISHD, FIXFORM
LEVEL 2, DATA, FLAGS

ENDFIL(K) = EOF(K) .NE. 0.0
REWIND 2

FIXFORM = .F. FMAPNO = 0 LIMIT = 0 UNRECOG = 0
CHECK FOR INPUT DIRECTIVES

2000 READ(5,2001) INP
2001 FORMAT(8A10)
C SO WE WANT A DUMMY RECORD IF HE LETS
C EVERYTHING DEFAULT
C F-MAP 0
2002 IF ((INP(1) .AND. MASK(30)) .NE. 5) GO TO 2010
C F-MAP - HAVE WE BEEN HERE BEFORE?
C IF (FMAPNO .NE. 0) CALL SYSTEM(52, '1 INIT DUPLICATE F-MAP DIRECTIVE FOUND')
C COPY ACROSS TOP LINE
DO 2003 I=1,8
   FMAPDIR(1,I) = INP(I)
2003 CONTINUE
FMAPNO = 1
C AND KEEP READING THE REST OF THEM
2005 READ(5,2001) (FMAPDIR(FMAPNO+1,J),J=1,8)
2006 IF (ENDFIL(5)) GO TO 2990
C WATCH FOR THE END OF F-MAPS
C IF ((FMAPDIR(FMAPNO+1,1) .AND. MASK(6)) .EQ. 1L) GO TO 2007
C SEE IF HE'S GIVEN US A 9999
C IF ((FMAPDIR(FMAPNO+1,1) .AND. MASK(6)) .EQ. 1L9) GO TO 2008
C YES (PROBABLY)
C FMAPNO = FMAPNO+1 IF (FMAPNO.GT.73) GO TO 2009
2007 GOTO 2002
C ELSE COPY THIS LINE BACK TO INP -
2008 DO 2006 I=1,8
   INP(I) = FMAPDIR(FMAPNO+1,I)
2009 CONTINUE
C
2006 CONTINUE
C    GO TO 2002
C    AND GO BACK TO THE TOP TO TEST IT AGAIN....
C    BUT FOR LEGITIMATE F-MAP DIR,
C 2007 FMAPNO = FMAPNO+1
C 2009 IF(FMAPNO.GT.73)CALL SYSTEM(52, '1*** TOO MANY FMAP DIRECTIVES - PROGRAM ABANDONED ***')
C    ELSE GET THE NEXT
C    GO TO 2005
C 2010 IF((INP(1).AND.MASK(30)).NE.BLIMIT) GO TO 2020
C    YES - SEEN BEFORE :
C    IF(LIMIT.NE.0)CALL SYSTEM(52, '1*** DUPLICATE LIMIT DIRECTIVE FOUND - PROGRAM ABANDONED ***')
C    OK - CONVERT TO INTEGER
C    REMOVE THE TEXT WORD -
C    INP(1) = INP(1).AND.(.NOT.MASK(30))
C    LEFT JUSTIFY THE REST OF THE LINE -
C    CALL SHUNT(INP,8)
C    WHICH SHOULD PUT THE INTEGER CHARs IN WORD 1 - SO ITS NOW OK TO INTIFY IT
C    IFAIL = 1 . LIMIT = INTIFY(INP(1),IFAIL)
C    IF(IFAIL.EQ.1) .OR. LIMIT.LT.1 .OR. LIMIT.GT.1000,
C      CALL SYSTEM(52, '1*** BAD LIMIT DIRECTIVE - PROGRAM ABANDONED ***')
C    ELSE OK
C    GO TO 2000
C 2020 IF((INP(1).AND.MASK(36)).NE.6LFORMAT)GO TO 2030
C    YES - HAD ONE ALREADY :
C      IF(FIXFORM)CALL SYSTEM(52, '1*** DUPLICATE FORMAT DIRECTIVE FOUND - PROGRAM ABANDONED ***')
C      FIXFORM = .T.
C      COPY ACROSS
C      FMT(1) = INP(1).AND.(.NOT.MASK(36)) .OR. 6L
C      DO 2021 I=2,8
C        FMT(I) = INP(I)
C      2021 CONTINUE
C    ELSE OK
C    GO TO 2000
C 2030 CONTINUE
C    HERE FOR UNRECOGNISED DIRECTIVE
C    PRINT 2989,INP
C    UNRECOG = UNRECOG+1
C    IF(UNRECOG.GT.50) CALL SYSTEM(52, '1*** TOO MANY UNRECOGNISED DIRECTIVES - PROGRAM ABANDONED ***')
C    GO TO 2000
C 2989 FORMAT('0*** UNRECOGNISED DIRECTIVE - IGNORED 0 ',8A10)
C    HERE AT END OF I/P FILE
C 2990 PRINT 2992
C 2992 FORMAT('0',T22,17('*'), ' PROGRAM SYMPREP ',17('*'))
C    IF(FIXFORM)PRINT 2993,FMT
C 2993 FORMAT('0*** FIXED FORMAT SPECIFIED AS 0 ',8A10)
C    IF(LIMIT.NE.0)PRINT 2994, LIMIT
C    IF(LIMIT.EQ.0) PRINT 2995
C 2994 FORMAT('0*** MAXIMUM NO OF POINTS REQUIRED IS ',I4)
C 2995 FORMAT('0 DEFAULT MAXIMUM NO OF POINTS (1000) USED')
C    IF(LIMIT.EQ.0) LIMIT=1000
C    DONE
C    START READS...
C    IF(.NOT.FIXFORM) READ (1) XMAX,YMAX,ZMAX
IF (FIXFORM) READ (1, FMT) XMAX, YMAX, ZMAX
IF (ENDFIL (1)) CALL SYSTEM (52, '*** EMPTY SOURCE FILE ***')
NPTS = 1
XMIN = XMAX YMIN = YMAX ZMIN = ZMAX

AND GO ON TO EOF
1 IF (.NOT. FIXFORM) READ (I) X, Y, Z
IF (FIXFORM) READ (1, FMT) X, Y, Z
IF (ENDFIL (1)) GO TO 5
NPTS = NPTS + 1
IF (X.GT.XMAX) XMAX = X IF (X.LT.XMIN) XMIN = X
IF (Y.GT.YMAX) YMAX = Y IF (Y.LT.YMIN) YMIN = Y
IF (Z.GT.ZMAX) ZMAX = Z IF (Z.LT.ZMIN) ZMIN = Z
GO TO 1

5 PRINT 7, NPTS, XMAX, XMIN, YMAX, YMIN, ZMAX, ZMIN
7 FORMAT (0, ' ', 18, ' DATA FIELDS ON SOURCE FILE', /
+ 0, ' ', 11, ' X CO-ORDINATE RANGE - ', 2F20.10, /,
+ 0, ' ', 11, ' Y CO-ORDINATE RANGE - ', 2F20.10, /,
+ 0, ' ', 8, 'IN P U T DATA VALUE RANGE - ', 2F20.10)

C CHECK FOR NON-SILLY RANGES
C IF (XMAX.EQ.XMIN) CALL SYSTEM (52, '*** ZERO X CO-ORDINATE RANGE ***' +)
C IF (YMAX.EQ.YMIN) CALL SYSTEM (52, '*** ZERO Y CO-ORDINATE RANGE ***' +)
C NOW START THE MATRIX FITTING.
C WE WANT TO LOOP ROUND SQUEEZZING THE WHOLE DATA SET INTO A MATRIX
C WHICH STARTS SMALL AND GETS BIGGER ON EACH PASS. THE BIGGER IT GETS,
C THE MORE DEFINED DATA POINTS WILL SURVIVE THE SQUEEZE.
C LO = SQRT (FLOAT (LIMIT)) IHI = 191 - ISTEP = 20 FINISHD = .F.
900 DO 1000 IMAT = LO, IHI, ISTEP

D0 8 I=1,IMAT
D0 8 J=1,IMAT
FLAGS(I,J) = 0
8 CONTINUE

XFACT = (FLOAT (IMAT) -1) / (XMAX-XMIN)
YFACT = (FLOAT (IMAT) -1) / (YMAX-YMIN)

NOW PLUG THE DATA VALS INTO OUR ARRAY,
KEEPING TRACK OF THE NUMBER OF MULTIPLE
ASSIGNMENTS TO ANY POINT.
NMULTS = 0
REWIND 1
NASS = 0
10 READ (I) X, Y, Z
IF (ENDFIL (1)) GO TO 20

C CONVERT CO-ORDS
IX = IFIX((X-XMIN) * XFACT) + 1
IY = IFIX((Y-YMIN) * YFACT) + 1
IF (FLAGS (IX, IY).EQ.0) GO TO 12

C NOT THE FIRST VALUE AT THIS POINT - AVERAGE
DATA(IX, IY) = ( DATA(IX, IY) + Z ) / 2.0
NMULTS = NMULTS + 1
GO TO 10
12 DATA(IX, IY) = Z
FLAGS(IX, IY) = 1
NASS = NASS + 1
GO TO 10

20 CONTINUE

IF (FINISH) PRINT 101, IMAT, IMAT, NASS, NMULTS
101 FORMAT ('D DATA POINTS AT MATRIX SIZE 'I4,' BY 'I4,' 'I5,' POINTS REMAIN
+NE ON AFTER 'I5,' REPEATED ASSIGNED')
C IF (NASS.GT.LIMIT) GO TO 1010
C OR DO WE NEED TO CONTINUE...
1000 CONTINUE
C IF WE GET HERE, THEN OUR
C MATRIX IS NOT BIG ENOUGH...

1005 FORMAT ('D WARNING: LESS THAN OPTIMAL DATA PACKING AT
+ LIMIT OF THIS PROGRAM';/'+
+ ACUITY IF APPLIED DIRECTLY';/'+
+ TO SYMAP',/';
1006 FORMAT ('D IH = 'I5,' L0 = 'F10.4,' NASS = 'I5,' NPTS = 'I5)
IMAT = IH GO TO 1030
C PRINT 1005
C PRINT 1006
1010 L0 = IMAT - ISTEP
IH = IMAT
FINISH = ISTEP.EQ.1
IF (.NOT.FINISH) ISTEP = ISTEP / 4
GO TO 900
C AND GO ROUND AGAIN
C 1030 WRITE (2,30)
30 FORMAT ('D DATA POINTS X')
C WRITE MATRIX CO-ORDS IN SYMAP FORMAT IF
C THE MATRIX POINTS CONTAIN VALUES

DO 40 IX = 1, IMAT
X = ( FLOAT(IX-1) / XFACT ) + XMIN
DO 40 IY = 1, IMAT
C A VALUE HERE
Y = YMAX - ( ( FLOAT(IY-1) / YFACT ) + YMIN )
WRITE (2,50) Y, X
40 CONTINUE
50 FORMAT (10x,2F10.4)
C WRITE (2,55)
55 FORMAT ('99999')
C END THE B-DATA
C WRITE (2,60)
C NOW REPEAT THE EXERCISE FOR THE VALUES...
GET MODIFIED ZMAX, ZMIN ON THE WAY

\[
\text{ZMAX} = \text{ZMIN} = \left( \frac{\text{ZMAX} + \text{ZMIN}}{2} \right) / 2.0
\]

DO 70 IX = 1, IMAT
   DO 70 IY = 1, IMAT
      IF (FLAGS(IX, IY) .EQ. 0) GO TO 70
      IF (DATA(IX, IY) .GT. ZMAX) ZMAX = DATA(IX, IY)
      IF (DATA(IX, IY) .LT. ZMIN) ZMIN = DATA(IX, IY)
   WRITE(2, 75) DATA(IX, IY)
CONTINUE

WRITE (2, 75) ZMAX, ZMIN

AND WRITE IT OUT

WRITE (2, 75) DATA(IX, IY)

WRITE (2, 80) FORMAT ('9999999')

WRITE (2, 82) (FMAPDIR(IJ), J = 1, FMAPNO)

WRITE (2, 85) FORMAT ('9999999999')

HERE FOR DEFAULTS

WRITE (2, 100) FORMAT ('F-MAP X , /', + '1 21.6 13.0 , /', + '3 10.0 , /', + '17 , /', + '21 , /', + '22 , /', + '24 , /', + '999999')

WRITE (2, 85) FORMAT ('F-MAP X ', /, + '0 END OF SYMPREP.', ' /')

AND TELL HIM THE GOOD NEWS

END FILE 2
REWIND 2

PRINT 102

END
PROGRAM WDEPTH (INDIP, JEDIP, TAPE8, TAPE7, OUTPUT, TAPE1=INDIP, TAPE2=JEDIP, TAPE6=OUTPUT)  

PROGRAM WRITTEN BY F.W.M. HOPPER 1978

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THIS PROGRAM RECOMBINES TWO INTERPOLATED APPARENT DIP FILES AND GENERATES THE THEORETICAL FOLLIATION SURFACE FROM THEM, TO BE PASSED TO SYMVU.
THE ONLY OTHER OUTPUT IS INFORMATION AS TO THE SIZE OF THE MATRIX AND THE MAXIMUM / MINIMUM VALUES GENERATED, CALCULATED FROM THE EAST TO THE WEST WITH 0.0 IN THE TOP RIGHT CORNER.

ARRAYS.
=======

ROW - CONTAINS CURRENT ROW DIP VALUES.
COL - CONTAINS LAST COLUMN FROM INDIP
DROP - CONTAINS DEPTHS OF THE SURFACE FROM AN ORDNANCE DATUM LEVEL IN THE CURRENT ROW.
SCALES - CONTAINS THE CONTROL VALUES FOR SYMVU.
DUM - A DUMMY ARRAY NOTHING USED FROM IT.

VARIABLES.
==========

NROW - NUMBER OF ROWS (= SCALES(1)).
NCOL - NUMBER OF COLUMNS (= SCALES(2)).
ROWDIST - DISTANCE BETWEEN THE ROWS (= SCALES(3)).
COLDIST - DISTANCE BETWEEN THE COLUMNS (= SCALES(4)).
MIN - MINIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(5)).
MAX - MAXIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(6)).
SCALES 7-24 ARE REDUNDANT BUT MUST BE PRESENT.

DIMENSION SCALES(24), DUM(200), COL(200), ROW(200), DROP(200)

THE ARRAY SCALES WILL FORWARD THE 24 VALUES NEEDED FOR THE SYMVU ROUTINE.

EQUIVALENCE (NROW, SCALES(1)), (NCOL, SCALES(2))

STATEMENT FUNCTIONS  -------------------------------------

TAND (ARG) = TAN (ARG * 0.0175)

REWIND THE TAPE TO BE USED AS INPUT, WHICH ARE 'INDIP' (INTERPOLATED NORTHERLY DIPS) AND 'JEDIP' (INTERPOLATED EASTERLY DIPS), AND HAVE BEEN COPIED FROM TAPE 8.

REWIND 1
REWIND 2
READ (1) (SCALES(K), K=1,24)
'DISTROW' AND 'DISTCOL' ARE THE DISTANCES BETWEEN THE GRID POSITIONS OF THE INTERPOLATED MATRIX.

DISTROW = 1.0 / SCALES(3)
DISTCOL = 1.0 / SCALES(4)
READ 1ST CARD OF (2) WITH SCALES ON.
READ (2)
READ FROM TAPE (1) THE LAST COLUMN FROM EACH ROW.
NN=NCOL-1
WRITE (6,133)
FORMATT(5X, '...............
      4X, ' ---- PROGRAM WDEPTH ----
      4X, '...............
      DO 88 I=1,NROW
      READ(1) (DUM(K),K=1,NROW)
      CONTINUE

WORK OUT THE DROP AT EACH POINT ALONG THE LAST COL OF THE MATRIX
FROM THE INTERPOLATED NORTHERLY DIPS, ASSUMING THE DROP IN THE TOP
RIGHT CORNER TO BE 0.0 AND WRITE THEM TO COL AS AN INTERMEDIATE
STORE.

      COL(I)=0.0
      DO 451 K=2,NROW
         R=ROW(K)
         ROW(K) = ABS (ROW(K))
         COL(K)=COL(K-1) + DISTROW * (SIGN (TAND (ROW(K)), R ))
      451 CONTINUE

DROPHI=DROPLO=COL(1)
NOW WORK OUT THE DROP AT ALL OTHER GRID POINTS USING THE
INTERPOLATED EASTERLY DIPS, EXCEPT THE LAST WORD ON THE ROW,
WHICH WE ALREADY KNOW FROM WHAT WE HAVE IN COL.

      DO 452 L=1,NROW
      READ(2) (ROW(K),K=1,NCOL)
      DROP(NCOL)=COL(L)
      452 CONTINUE

      DO 74 I=1,NCOL
         IF (DROP(I) .GE. DROPHI) DROPHI=DROP(I) IF (DROP(I) .LT. DROPLO)
            DROPLO=DROP(I)
      74 CONTINUE

WRITE VALUES TO TAPE 7 AS AN INTERMEDIATE STOR.
      WRITE(7) (DROP(K),K=1,NCOL)
      ENDFILE 7

WE NOW HAVE THE MINIMUM AND MAXIMUM DROP VALUES AND THESE ARE
PUT IN SCALES(5) AND SCALES(6) -(24) RESPECTIVELY.

      DO 454 K=6,24
      SCALES(K) = DROPHI
      SCALES(5) = DROPLO
      CONTINUE
      WRITE(8) SCALES
      DO 700 J=1,NROW
      READ(7) (DROP(J),J=1,NCOL)
      WRITE(8) (DROPHI,J=1,NCOL)
700 CONTINUE
ENDFILE 8
REWIND 8
WRITE (6,777) NROW,NCOL,(SCALES(K),K=0,5)
777 FORMAT (5X,'ROWS',5X,15,/,5X,'COLUMNS',5X,15,/,-
5X,'ROWSPACING',5X,22.11,/,5X,'COLUMNSPACING',5X,22.11,/,-
5X,'LOWEST VALUE',5X,22.11,/,5X,'HIGHEST VALUE',5X,22.11)
C FIND OUT IF THERE ARE GOING TO BE ANY SCALING
C PROBLEMS WHEN SYMVU GETS THE FILE.
IF (DROPHI.EQ.0) GO TO 771
IF ((DROPHI+DROPLQ)/DROPHI.GT.2/3) PRINT (6,778)
771 CONTINUE
778 FORMAT (20X,'WARNING',/+/5X,'YOU HAVE A SCALING PROBLEM YOUR SYMVU PLOT MAY'/+
5X,'LOOK FLAT. YOU HAVE EITHER A VERY HIGH (HIGHEST '/+
5X,'VALUE) OR A VERY LOW (LOWEST VALUE). SEE ABOVE'/+
5X,'IF YOUR WHOLE DATA IS NOT EVENLY SPREAD BETWEEN THEM'/+
5X,'DURING SCALING THE VARIATION IN THE SURFACE MAY BE'/+
5X,'LOST.'/+
5X,'THIS MAY BE OVERCOME BY CAREFUL SELECTION OF THE '/+
5X,'APPROPRIATE SYMVU ELECTIVES - SEE MANUAL - OR '/+
5X,'YOU MAY HAVE TO OPERATE ON YOUR MATRIX. TO DO THIS'/+
5X,'-FIRST EXAMINE IT BY LOOKING AT THE TAPEB PRODUCED'/+
5X,'BY THIS PROGRAM, WITH PROGRAM GRID IN THIS LIBRARY'/+
5X,'SEE DOCUMENTATION-. THIS WILL SHOW YOU THE LOCATION'/+
5X,'OF THE TROUBLE. WITH THIS KNOWLEDGE YOU CAN START TO'/+
5X,'PRUNE OUT THE OFFENDING READINGS. GET HELP DOING THIS'/+
+/5X,'STOP
END
PROGRAM EDEPTH (INDIP, IEDIP, TAPE1, TAPE2, OUTPUT, TAPE1=INDIP, TAPE2=IEDIP, TAPE3=OUTPUT)

PROGRAM WRITTEN BY F.W.M. HOPPER 1978

THIS PROGRAM RECOMBINES TWO INTERPOLATED APPARENT DIP FILES AND GENERATES THE THEORETICAL FOLLIATION SURFACE FROM THEM TO BE PASSED TO SYMVU.

THE ONLY OTHER OUTPUT IS INFORMATION AS TO THE SIZE OF THE MATRIX AND THE MAXIMUM / MINIMUM VALUES GENERATED.

CALCULATED FROM THE WEST TO THE EAST WITH 0.0 IN THE BOTTOM LEFT CORNER.

ARRAYS.

======

ROW - CONTAINS CURRENT ROW DIP VALUES.
COLL - CONTAINS FIRST COLUMN FROM INDIP
DROP - CONTAINS DEPTHS OF THE SURFACE FROM AN ORDNANCE DATUM LEVEL IN THE CURRENT ROW.
SCALES - CONTAINS THE CONTROL VALUES FOR SYMVU.

VARIABLES.

=========

NROW - NUMBER OF ROWS ( = SCALES(1) ).
NCOL - NUMBER OF COLUMNS ( = SCALES(2) ).
ROWDIST - DISTANCE BETWEEN THE ROWS ( = SCALES(3) ).
COLDIST - DISTANCE BETWEEN THE COLUMNS ( = SCALES(4) ).
MIN - MINIMUM VALUE OF INTERPOLATED DIP VALUES ( = SCALES(5) ).
MAX - MAXIMUM VALUE OF INTERPOLATED DIP VALUES ( = SCALES(6) ).
SCALES 7-24 ARE REDUNDANT BUT MUST BE PRESENT.

DIMENSIONS SCALES(24), COL(200), ROW(200), DROP(200)

THE ARRAY SCALES WILL FORWARD THE 24 VALUES NEEDED FOR THE SYMVU ROUTINE.

EQUIVALENCE (NROW, SCALES(1)), (NCOL, SCALES(2))

STATEMENT FUNCTIONS ===============

TAND ( ARG ) = TAN ( ARG * 0.0175 )

REWIND THE TAPES TO BE USED AS INPUT, WHICH ARE 'INDIP' (INTERPOLATED NORTHERLY DIPS) AND 'IEDIP' (INTERPOLATED EASTERLY DIPS), AND HAVE BEEN COPIED FROM TAPE 8.

REWIND 1
REWIND 2
READ(1) (SCALES(K), K=1,24)
'DISTROW' AND 'DISTCOL' ARE THE DISTANCES BETWEEN THE GRID POSITIONS OF THE INTERPOLATED MATRIX.

DISTROW = 1.0 / SCALES(3)
DISTCOL = 1.0 / SCALES(4)

READ 1ST CARD OF (2) WITH SCALES ON.
READ(2)

READ FROM TAPE(1) THE FIRST COLUMN FROM EACH ROW.
WRITE(6,133)
133 FORMAT(5X, '------------------------------', '///', +5X, '--- PROGRAM EDEPTH ---', '///',):
D O  88 1=1,N R O W
READ(1) ROW(I)
CONTINUE

WORK OUT THE DROP AT EACH POINT ALONG THE FIRST COL OF THE MATRIX
FROM THE INTERPOLATED NORTHERLY DIPS, ASSUMING THE DROP IN THE BOTTOM
LEFT CORNER TO BE 0.0 AND WRITE THEM TO COL AS AN INTERMEDIATE
STORE.

COL(NROW)=0.0
DO 451 J=2,NROW
  K  =  NROW - J + 1
  R=ROW(K) ** (-1.0)
  ROW(K) = ABS(ROW(K))
  COL(K) = COL(K+1) + DISTROW*(SIGN(TAND(ROW(K)),R))
451 CONTINUE

DRGPHI=DRGPLO=COL(1)
NOW WORK OUT THE DROP AT ALL OTHER GRID POINTS USING THE
INTERPOLATED EASTERNLY DIPS, EXCEPT THE FIRST WORD ON THE ROW,
WHICH WE ALREADY KNOW FROM WHAT WE HAVE IN COL.

DO 452 L=1,NROW
READ(2) (ROW(K),K=1,NCOL)
DROP(1)=COL(L)
  DO 453 K=2,NCOL
    R=ROW(K) ** (-1.0)
    ROW(K) = ABS(ROW(K))
    DROP(K)=DROP(K-1)+DISTCOL*(SIGN(TAND(ROW(K)),R))
453 CONTINUE

DO 74 I=1,NCOL
  IF (DROP(I).GT.DRGPHI) DROPHI=DROP(I) IF (DROP(I).LT.DRPLO)
  -DRPLO=DROP(I)
  CONTINUE

WRITE VALUES TO TAPE 7 AS AN INTERMEDIATE STORE.
WRITE(7) (DROP(K),K=1,NCOL)
452 CONTINUE
ENDFILE 7

SCALES(5) = DROPHI
SCALES(6) = DRPLO
REWIND 7
WRITE(8) NROW,NCOL,(SCALES(K),K=3,24)
DO 700 1=1,NROW
READ(7) (DROP(J),J=1,NCOL)
WRITE(8) (DROP(J),J=1,NCOL)
700 CONTINUE

WE NOW HAVE THE MINIMUM AND MAXIMUM DROP VALUES AND THESE ARE
PUT IN SCALES(5) AND SCALES(6) -(24) RESPECTIVELY.
CONTINUE
ENDFILE 8
REWIND 8
WRITE(6,777) NROW,NCOL,(SCALES(K),K=3,6)
777 FORMAT(5X,'ROWS',5X,15,I5,/,5X,'COLUMNS',5X,I5,/, -5X,'ROWSSPACING',5X,G22.11,/,5X,'COLUMNSSPACING',5X,G22.11,/, -5X,'LOWEST VALUE',5X,G22.11,/,5X,'HIGHEST VALUE',5X,G22.11)
C FIND OUT IF THERE ARE GOING TO BE ANY SCALING
C PROBLEMS WHEN SYMVU GETS THE FILE.
IF((DROPHI.EQ.0.0).GO TO 771
IF((DROPHI+DROPLO)/DRGPHI.GT.2/3) PRINT (6,778)
771 CONTINUE
778 FORMAT(20X,'WARNING',/ +,5X,'YOU HAVE A SCALING PROBLEM YOUR SYMVU PLOT MAY'/ +,5X,'LOOK FLAT. YOU HAVE EITHER A VERY HIGH (HIGHEST '/ +,5X,'VALUE) OR A VERY LOW (LOWEST VALUE), SEE ABOVE'/ +,5X,'IF YOUR WHOLE DATA IS NOT EVENLY SPREAD BETWEEN THEM,'/ +,5X,'DURING SCALING THE VARIATION IN THE SURFACE MAY BE'/ +,5X,'LOST.'/ +,5X,'THIS MAY BE OVERCOME BY CAREFUL SELECTION OF THE '/ +,5X,'APPROPRIATE SYMVU ELECTIVES - SEE MANUAL', OR '/ +,5X,'YOU MAY HAVE TO OPERATE ON YOUR MATRIX. TO DO THIS'/ +,5X,'--FIRST EXAMINE IT, BY LOOKING AT THE TAPES PRODUCED'/ +,5X,'BY THIS PROGRAM, WITH PROGRAM GRID IN THIS LIBRARY'/ +,5X,'SEE DOCUMENTATION-. THIS WILL SHOW YOU THE LOCATION'/ +,5X,'OF THE TROUBLE. WITH THIS KNOWLEDGE YOU CAN START TO'/ +,5X,'PRUNE OUT THE OFFENDING READINGS. GET HELP DOING THIS '/ +,5X,'STOP
END
PROGRAM SDEPTH (INDIP, IEDIP, TAPE7, TAPE8, OUTPUT, TAPE1=INDIP, 
1TAPE2=IEDIP, TAPE6=OUTPUT)

PROGRAM WRITTEN BY F.W.M. HOPPER 1978.

THIS PROGRAM RECOMBINES TWO INTERPOLATED APPARENT DIP 
FILES AND GENERATES THE THEORETICAL FOLLIATION SURFACE 
FROM THEM, TO BE PASSED TO SYMVU.
THE ONLY OTHER OUTPUT IS INFORMATION AS TO THE SIZE OF THE 
MATRIX AND THE MAXIMUM / MINIMUM VALUES GENERATED.
CALCULATED FROM NORTH TO SOUTH WITH 0.0 IN THE TOP RIGHT 
CORNER.

ARRAYS.
======
ROW - CONTAINS CURRENT ROW DIP VALUES.
DROP - CONTAINS DEPTHS OF THE SURFACE FROM AN 
ORDNANCE DATUM LEVEL IN THE CURRENT ROW.
SCALES - CONTAINS THE CONTROL VALUES FOR SYMVU.

VARIABLES.
========
NROW - NUMBER OF ROWS ( = SCALES(1) ).
NCOL - NUMBER OF COLUMNS ( = SCALES(2) ).
DISTROW - DISTANCE BETWEEN THE ROWS.
DISTCOL - DISTANCE BETWEEN THE COLUMNS.
MAX - MAXIMUM VALUE OF INTERPOLATED DIP VALUES.
MIN - MINIMUM VALUE OF INTERPOLATED DIP VALUES.

DIMENSION SCALES(24), ROW(200), DROP(200)

THE ARRAY SCALES WILL FORWARD THE 24 VALUES NEEDED FOR THE 
SYMVU ROUTINE.

EQUIVALENCE (NROW, SCALES(1)), (NCOL, SCALES(2))

STATEMENT FUNCTIONS

TAND (ARG) = TAN(ARG * 0.0175)

REWIND THE TAPES TO BE USED AS INPUT, WHICH ARE 'INDIP' 
(INTERPOLATED NORTHERLY DIPS) AND 'IEDIP' (INTERPOLATED 
EASTERLY DIPS), AND HAVE BEEN COPIED FROM TAPE 8, SYMAP OUTPUT.

REWIND 1
REWIND 2

THE FIRST 24 WORDS CONTAIN THE CONTROL INFORMATION FROM SYMAP.

SCALES(1) - (NROW) - NUMBER OF ROWS.
SCALES(2) - (NCOL) - NUMBER OF COLUMNS.
SCALES(3) - NUMBER OF ROWS TO AN INCH.
SCALES(4) - NUMBER OF COLUMNS TO AN INCH.
SCALES(5) - MINIMUM VALUE OF INTERPOLATED DIP VALUES.
SCALES(6)-(24) - MAXIMUM VALUE OF INTERPOLATED DIP VALUES.

READ(1) (SCALES(K), K=1,24)

'DISTROW' AND 'DISTCOL' ARE THE DISTANCES BETWEEN THE GRID
POSITIONS OF THE INTERPOLATED MATRIX.

DISTROW = 1.0 / SCALES(3)
DISTCOL = 1.0 / SCALES(4)
READ(2)

HAVING READ OF THE SCALES ON 2 READ THE FIRST ROW.
READ(2) (ROW(I), I=1, NCOL)

WORK OUT THE DROP AT EACH POINT ALONG THE TOP ROW OF THE MATRIX
FROM THE INTERPOLATED EASTERLY DIPS, ASSUMING THE DROP IN THE TOP
RIGHT CORNER TO BE 0.0 AND WRITE THEM TO TAPE 7 AS AN
INTERMEDIATE STOR.

DROP(NCOL)=0.0
WRITE(6,133)
133 FORMAT(5X, '-------------------------',:///,
       +5X, '-- PROGRAM SDEPTH --',:///,
       +5X, '-------------------------'),)
DO 451 J=2, NCOL
K=NCOL-J+1
R=ROW(K)
ROW(K)=ABS(ROW(K))
DROP(K)=DROP(K+1)+DISTCOL*SIGN(TAND(R0W(K)),R)
451 CONTINUE
WRITE(7) (DROP(K), K=1, NCOL)

INITIALISE 'DROPHI' AND 'DROPLO'

DROPHI=DROPLO=DROP(1)
DO 79 I=2, NCOL
IF (DROP(I).GT.DROPHI) DROPHI=DROP(I) IF (DROP(I).LT.DROPLO) DROPLO=

79 CONTINUE

NOW WORK OUT THE DROP AT ALL OTHER GRID POINTS USING THE
INTERPOLATED NORTHERLY DIPS, AND WRITE THESE TO TAPE 7 TOO.

READ(1)
DO 452 L=2, NROW
READ(1) (ROW(K), K=1, NCOL)

452 CONTINUE

DROPHI=DROPLO=DROP(1)
DO 73 J=2, NCOL

73 CONTINUE

IF (DROP(I).GT.DROPHI) DROPHI=DROP(I) IF (DROP(I).LT.DROPLO) DROPLO=
CONTINUE

WRITE(7) (DROP(K),K=1,NCOL)

ENDFILE 7
ENDFILE 7

NOW HAVE THE MINIMUM AND MAXIMUM DROP VALUES AND THESE ARE
PUT IN SCALES(5) AND SCALES(6)-(24) RESPECTIVELY.

DO 454 K=6,24
SCALES(K) = DROPHI
CONTINUE
SCALES(5) = DROPLO
REWIND 7
WRITE(8) NROW,NCOL,(SCALES(K),K=3,24)
DO 700 I=1,NROW
READ(7) (DROP(J),J=1,NCOL)
WRITE(8) (DROP(J),J=1,NCOL)
CONTINUE
ENDFILE 8
REWIND 8
WRITE(6,777) NROW,NCOL,(SCALES(K),K=3,6)

777 FORMAT(5X, 'ROWS ',5X,15,/,5X,'COLUMNS',5X,15,/,5X,'ROWSPACING',5X,G22.11,/,5X,'COLUMNSPACING',5X,G22.11,/,5X,'LOWEST VALUE',5X,G22.11,/,5X,'HIGHEST VALUE',5X,G22.11)

C FIND OUT IF THERE ARE GOING TO BE ANY SCALING
C PROBLEMS WHEN SYMVU GETS THE FILE.
IF (DROPHI.EQ.0.0) GO TO 771
IF ((DROPHI+DROPLO)/DROPHI.GT.2/3) PRINT (6,778)

771 CONTINUE

778 FORMAT(20X, '********** WARNING **********')
/+5X,'YOU HAVE A SCALING PROBLEM YOUR SYMVU PLOT MAY'/
/+5X,'LOOK FLAT. YOU HAVE EITHER A VERY HIGH (HIGHEST '/
/+5X,'VALUE) OR A VERY LOW (LOWEST VALUE); SEE ABOVE.' /
/+5X,'IF YOUR WHOLE DATA IS NOT EVENLY SPREAD BETWEEN THEM,' /
/+5X,'DURING SCALING THE VARIATION IN THE SURFACE MAY BE'/
/+5X,'LOST.' /
/+5X,'THIS MAY BE OVERCOME BY CAREFUL SELECTION OF THE '/
/+5X,'APPROPRIATE SYMVU ELECTIVES - SEE MANUAL-,' OR '/
/+5X,'YOU MAY HAVE TO OPERATE ON YOUR MATRIX. TO DO THIS'/
/+5X,'FIRST EXAMINE IT, BY LOOKING AT THE TAPE PRODUCED'/
/+5X,'BY THIS PROGRAM, WITH PROGRAM GRID IN THIS LIBRARY'/
/+5X,'SEE DOCUMENTATION-. THIS WILL SHOW YOU THE LOCATION'/
/+5X,'OF THE TROUBLE. WITH THIS KNOWLEDGE YOU CAN START TO'/
/+5X,'PRUNE OUT THE OFFENDING READINGS. GET HELP DOING THIS.'/
/+5X,'********** WARNING **********')
STOP
END
PROGRAM NDEPTH (INDIP, IEDIP, TAPE8, TAPE7, OUTPUT, TAPE1=INDIP, 
- TAPE2=IEDIP, TAPE6=OUTPUT)

PROGRAM WRITTEN BY F.W.M. HOPPER 1978

THIS PROGRAM RECOMBINES TWO INTERPOLATED APPARENT DIP 
FILES AND GENERATES THE THEORETICAL FOLIATION SURFACE 
FROM THEM, TO BE PASSED TO SYMVU.

THE ONLY OTHER OUTPUT IS INFORMATION AS TO THE SIZE OF THE 
MATRIX AND THE MAXIMUM / MINIMUM VALUES GENERATED.

CALCULATION DONE FROM SOUTH TO NORTH WITH 0.0 IN THE BOTTOM 
LEFT CORNER.

ARRAYS.

ROW - CONTAINS CURRENT ROW DIP VALUES.
DROP - CONTAINS DEPTHS OF THE SURFACE FROM AN 
ORDNANCE DATUM LEVEL IN THE CURRENT ROW.
SCALES - CONTAINS THE CONTROL VALUES FOR SYMVU.

VARIABLES.

NROW - NUMBER OF ROWS (= SCALES(1)).
NCOL - NUMBER OF COLUMNS (= SCALES(2)).
ROWDIST - DISTANCE BETWEEN THE ROWS (= SCALES(3)).
COLDIST - DISTANCE BETWEEN THE COLUMNS (= SCALES(4)).
MIN - MINIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(5)).
MAX - MAXIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(6)).

SCALES 7-24 ARE REDUNDANT BUT MUST BE PRESENT.

DIMENSION SCALES(24), ROW(200), DROP(200)

THE ARRAY SCALES WILL FORWARD THE 24 VALUES NEEDED FOR THE 
SYMVU ROUTINE.

EQUIVALENCE (NROW, SCALES(1)), (NCOL, SCALES(2))

STATEMENT FUNCTIONS -----------------------------

TAND (ARG) = TAN (ARG * 0.0175)

REWIND THE TAPES TO BE USED AS INPUT, WHICH ARE 'INDIP'
(INTERPOLATED NORTHERLY DIPS) AND 'IEDIP' (INTERPOLATED 
EASTERLY DIPS), AND HAVE BEEN COPIED FROM TAPE 8.

REWIND 1
REWIND 2
READ(1) (SCALES(K), K=1, 24)

'DISTROW' AND 'DISTCOL' ARE THE DISTANCES BETWEEN THE GRID 
POSITIONS OF THE INTERPOLATED MATRICES.

DISTROW = 1.0 / SCALES(3)
DISTCOL = 1.0 / SCALES(4)

READ ALL CARDS ON 2 EXCEPT THE LAST.
WRITE (6, 133)

133 FORMAT (5X, '---------------------------------------------------------------------------', / / / , 
+5X, ' ------ PROGRAM NDEPTH ------ ', / / / , 
+5X, '---------------------------------------------------------------------------'
)
DO 37 K=1, NROW
READ (2)
37 CONTINUE
READ (2) (ROW(I), I=1, NCOL)

WORK OUT THE DROP AT EACH POINT ALONG THE BOTTOM ROW OF THE MATRIX
FROM THE INTERPOLATED EASTERLY DIPS, ASSUMING THE DROP IN THE BOTTOM
LEFT CORNER TO BE 0.0 AND WRITE THEM TO TAPE 7 AS AN INTERMEDIATE
STORE.

DROP(1)=0.0
DO 451 K=2, NCOL
   R=ROW(K) * (-1.0)
   ROW(K) = ABS(ROW(K))
   DROP(K)=DROP(K-1) + DISTCOL * (SIGN (TAND (ROW(K)), R))
451 CONTINUE
WRITE (7) (DROP(K), K=1, NCOL)

DROPHI=DROPLO=DROP(1)
DO 73 I=2, NCOL
   IF (DROP(I) .GT. DROPHI) DROPHI=DROP(I)
   IF (DROP(I) .LT. DROPLO) DROPLO=-DROP(I)
73 CONTINUE

NOW WORK OUT THE DROP AT ALL OTHER GRID POINTS USING THE
INTERPOLATED NORTHERLY DIPS.

READ TO END OF 1 THEN WORK BACKWARDS THROUGH IT.
DO 93 I=1, NROW
   READ (1)
93 CONTINUE
DO 452 L =2, NROW
   BACKSPACE 1
   BACKSPACE 1
   READ (1) (ROW(K), K=1, NCOL)
   DO 453 K=1, NCOL
      R=ROW(K) * (-1.0)
      ROW(K) = ABS(ROW(K))
      DROP(K)=DROP(K) + DISTROW * (SIGN (TAND (ROW(K)), R))
453 CONTINUE

DO 74 I=1, NCOL
   IF (DROP(I) .GT. DROPHI) DROPHI=DROP(I)
   IF (DROP(I) .LT. DROPLO) DROPLO=-DROP(I)
74 CONTINUE
WRITE (7) (DROP(K), K=1, NCOL)
452 CONTINUE
ENDFILE 7

WE NOW HAVE THE MINIMUM AND MAXIMUM DROP VALUES AND THESE ARE
PUT IN SCALES(5) AND SCALES(6)-(24) RESPECTIVELY.

DO 454 K=6, 24
   SCALES(K) = DROPHI
454 CONTINUE
SCALES(5) = DROPLO
BACKSPACE 7
WRITE (8) SCALES
DO 700 I=1,NROW
READ(7) (DROP(J),J=1,NCOL)
WRITE(8) (DROP(J),J=1,NCOL)
BACKSPACE 7
BACKSPACE 7
700 CONTINUE
ENDFILE 8
REWIND 8
WRITE(6,777) NROW,NCOL,(SCALES(K),K=3,5)
777 FORMAT(5X,'ROWS',5X,15,F15.15,/,5X,'COLUMNS',5X,15,F15.15,
-5X,'ROWSPACING',5X,G22.12,/,5X,'COLUMNSPACING',5X,G22.12,
-5X,'LOWEST VALUE',5X,G22.12,/,5X,'HIGHEST VALUE',5X,G22.12)
C FIND OUT IF THERE ARE GOING TO BE ANY SCALING
C PROBLEMS WHEN SYMVU GETS THE FILE.
IF(DROPHI.EQ.0.0)GO TO 771
IF(( DROPHI+DROPL0)/DROPHI.GT.2/3) PRINT (6,778)
771 CONTINUE
778 FORMAT (20X,'WARNING',5X,'YOU HAVE A SCALING PROBLEM. YOUR SYMVU PLOT MAY'
+5X,'LOOK FLAT. YOU HAVE EITHER A VERY HIGH (HIGHEST'
+5X,'VALUE) OR A VERY LOW (LOWEST VALUE). SEE ABOVE.'
+5X,'IF YOUR WHOLE DATA IS NOT EVENLY SPREAD BETWEEN THEM,'
+5X,'DURING SCALING THE VARIATION IN THE SURFACE MAY BE'
+5X,'LOST.'
+5X,'THIS MAY BE OVERCOME BY CAREFUL SELECTION OF THE'
+5X,'APPROPRIATE SYMVU ELECTIVES - SEE MANUAL- OR '/'
+5X,'YOU MAY HAVE TO OPERATE ON YOUR MATRIX. TO DO THIS'
+5X,'FIRST EXAMINE IT, BY LOOKING AT THE TAPEB PRODUCED'
+5X,'SEE DOCUMENTATION- THIS WILL SHOW YOU THE LOCATION'
+5X,'OF THE TROUBLE. WITH THIS KNOWLEDGE YOU CAN START TO'
+5X,'PRUNE OUT THE OFFENDING READINGS. GET HELP DOING THIS.'
+5X,'WARNING',5X,'YOU HAVE A SCALING PROBLEM. YOUR SYMVU PLOT MAY'
STOP
END
PROGRAM DIEPTH (INDIP, IEDIP, TAPE8, TAPE7, OUTPUT, TAPE1=INDIP, -TAPE2=IEDIP, TAPE6=OUTPUT)

PROGRAM WRITTEN BY F.W.M. HOPPER 1978

THIS PROGRAM RECOMBINES TWO INTERPOLATED APPARENT DIP FILES AND GENERATES THE THEORETICAL FOLIATION SURFACE FROM THEM TO BE PASSED TO SYMVU.
THE ONLY OTHER OUTPUT IS INFORMATION AS TO THE SIZE OF THE MATRIX AND THE MAXIMUM / MINIMUM VALUES GENERATED.
THIS IS THE BEST PROGRAM TO USE AS IT COMBINES THE VALUE FOR THE POINT FROM EACH TAPE TO GENERATE THE SURFACE. THIS ELIMINATES THE REMAINDER EFFECT WHEN THE FOLIATION SURFACE IS AN INTERFERENCE PATTERN.
THE 0.0 POINT IS IN THE BOTTOM LEFT CORNER.

ARRAYS.

ROW - CONTAINS CURRENT ROW DIP VALUES.
DROP - CONTAINS DEPTHS OF THE SURFACE FROM AN ORDNANCE DATUM LEVEL IN THE CURRENT ROW.
SCALES - CONTAINS THE CONTROL VALUES FOR SYMVU.
EDGE - HOLDS THE VALUES FOR THE LEFT EDGE OF THE MATRIX.
COL - THIS HOLDS THE ROW FROM TAPE 1 AND FOR A TIME THE LEFT EDGE VALUES.
DROPP - AN INTERMEDIATESTORE OF DROPS.

VARIABLES.

NROW - NUMBER OF ROWS (= SCALES(1)).
NCOL - NUMBER OF COLUMNS (= SCALES(2)).
ROWDIST - DISTANCE BETWEEN THE ROWS (= SCALES(3)).
COLDIST - DISTANCE BETWEEN THE COLUMNS (= SCALES(4)).
MIN - MINIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(5)).
MAX - MAXIMUM VALUE OF INTERPOLATED DIP VALUES (= SCALES(6)).
SCALES 7-24 ARE REDUNDANT BUT MUST BE PRESENT.

DIMENSION SCALES(24), ROW(200), DROP(200), EDGE(200), COL(200), -DROPP(200)

THE ARRAY SCALES WILL FORWARD THE 24 VALUES NEEDED FOR THE SYMVU ROUTINE.

EQUIVALENCE (NROW, SCALES(1)), (NCOL, SCALES(2))

STATEMENT FUNCTIONS  -----------------------------
TAND (ARG) = TAN (ARG x 0.0175)
ATAND (ARG) = ATAN (ARG) x 57.143
SIND (ARG) = SIN (ARG x 0.0175)
COSD (ARG) = COS (ARG x 0.0175)

REWIND THE TAPES TO BE USED AS INPUT, WHICH ARE 'INDIP' (INTERPOLATED NORTHERLY DIPS) AND 'IEDIP' (INTERPOLATED EASTERLY DIPS), AND HAVE BEEN COPYED FROM TAPE 8.

REWIND 1
REWIND 2
READ (1) (SCALES(K), K=1,24)
'DISTROW' AND 'DISTCOL' ARE THE DISTANCES BETWEEN THE GRID
C POSITIONS OF THE INTERPOLATED MATRIX.
C
DISTROW = 1.0 / SCALES(3)
DISTCOL = 1.0 / SCALES(4)
DIAG = SORT ( DISTROW * DISTROW + DISTCOL * DISTCOL )
ANG = ATAND (DISTCOL / DISTROW)
C
READ ALL CARDS ON 2 EXCEPT THE LAST.
WRITE(6,133)
133 FORMAT (5X, ' ', / / / / , +5X, 'PROGRAM DIEPTH --- ', / / / / , +5X, ' ', / / / / , +5X, ' ',)
DO 37 K = 1, NROW
READ(2)
37 CONTINUE
C READ LAST ROW FROM TAPE 2
READ(2) (ROW(I), I = 1, NCOL)
C
C WORK OUT THE DROP AT EACH POINT ALONG THE BOTTOM ROW OF THE MATRIX
C FROM THE INTERPOLATED EASTERN DIPS, ASSUMING THE DROP IN THE BOTTOM
C LEFT CORNER TO BE 0.0 AND WRITE THEM TO TAPE 7 AS AN INTERMEDIATE
C STORE.
C
DROP(1) = 0.0
DO 451 K = 2, NCOL
   R = ROW(K) * (-1.0)
   ROW(K) = ABS(ROW(K))
   DROP(K) = DROP(K - 1) + DISTCOL * (SIGN(TAND(ROW(K)), R))
451 CONTINUE
WRITE(7) (DROP(K), K = 1, NCOL)
C
C DROPHI = DROPLO = DROP(1)
DO 73 I = 2, NCOL
   IF (DROP(I) .GT. DROPHI) DROPHI = DROP(I)
   IF (DROP(I) .LT. DROPLO) DROPLO = -DROP(I)
73 CONTINUE
C
GET TO SECOND LAST ROW OF 2
BACKSPACE 2
BACKSPACE 2
READ(2) (ROW(I), I = 1, NCOL)
C
C NOW GET THE VALUES FOR THE LEFT EDGE FROM 1, READ THEM TO COL
C AND STORE THE VALUES IN EDGE.
C
DO 333 I = 1, NROW
   J = NROW - I + 1
   READ(1) COL(J)
333 CONTINUE
C EDGE(I) = DROP(1)
DO 334 I = 2, NROW
   R = COL(I) * (-1.0)
   COL(I) = ABS(COL(I))
   EDGE(I) = EDGE(I - 1) + DISTROW * (SIGN(TAND(COL(I)), R))
334 CONTINUE
C
NOW GET THE VALUES FOR THE SECOND LAST ROW OF 1 INTO COL.
BACKSPACE 1
BACKSPACE 1
READ(1) (COL(K), K = 1, NCOL)
C
NOW WORK OUT ALL THE OTHER GRID POINTS USING THE
INFORMATION FROM BOTH TAPES TO FIND OUT THE DROP
ALONG THE DIAGONAL.

DO 337 J = 2, NROW
IF (J.EQ.2) GO TO 338
BACKSPACE 1
BACKSPACE 1
READ (1) (COL (L), L = 1, NCOL)
BACKSPACE 2
BACKSPACE 2
READ (2) (ROW (L), L = 1, NCOL)
CONTINUE

DO 336 K = 2, NCOL
A = ABS (COL (K))
B = ABS (ROW (K))
FIE = ATAND (TAND (B) / TAND (A))
DIP = ATAND (TAND (B) * (1 / COSD (FIE)))
IF (COL (K), GE, 0.0 AND ROW (K), GE, 0.0) AZIMUTH = 90.0 - FIE
IF (COL (K), LE, 0.0 AND ROW (K), LT, 0.0) AZIMUTH = 270.0 + FIE
IF (COL (K), LT, 0.0 AND ROW (K), LE, 0.0) AZIMUTH = 90.0 + FIE
IF (COL (K), LT, 0.0 AND ROW (K), LT, 0.0) AZIMUTH = 270.0 - FIE

THIS HAS GIVEN US THE AZIMUTH AND TRUE DIP OF THE PLANE AT
THIS POINT AS DEFINED BY THE TWO APPARENT DIPS.

FROM THE AZIMUTH THE STRIKE WILL BE FOUND ON THE EAST END
OF THE AXIS.

IF (AZIMUTH, LT, 90.0) STRIKE = AZIMUTH + 90.0
IF (AZIMUTH, GE, 90.0 AND AZIMUTH, LT, 270.0) STRIKE = AZIMUTH - 90.0
IF (AZIMUTH, GE, 270.0) STRIKE = AZIMUTH - 270.0

NOW FIND THE TWO LIMITING AZIMUTH DIRECTIONS WHICH DEFINE
WEATHER THE COMPONENT OF DIP IS DOMINENTLY TO THE NE. OR SW.
AND HENCE WEATHER THE FINAL DIP WILL BE (+) OR (-).

XNTH = 270.0 + ANG
XSTH = 90.0 + ANG
NOW WORK OUT THE ANGLE BETWEEN THE STRIKE IN THE E. QUADRANT
AND THE DIRECTION OF THE DIAGNAL TO THE NE.

GAP = ABS (STRIKE - ANG)

NOW FIND THE APPARENT DIP ALONG THE NE. DIAGNAL.

DIDIP = ATAND (TAND (DIP) * SIND (GAP)) * (-1.0)

IF THE AZIMUTH WAS IN THE RIGHT DIRECTION ( TO THE SW. )
MAKE THE DIDIP (+).

IF (AZIMUTH, LE, XNTH, AND, AZIMUTH, GT, XSTH)
DIDIP = DIDIP * (-1.0)

NOW WORK OUT THE DROPS FROM THESE WITH REFERENCE TO
THE LAST DROPS FOUND. STORE THE DROPS IN DROPP UNTIL
THE END THEN REPLACE THE OLD DROPS WITH THE VALUES FROM DROPP.

DROPP(K) = DROPP(K-1) + DIAG * (TAND(DIDIP))

CONTINUE

DO 75 I = 2, NCOL
    DROPP(I) = EDGE(J)
    CONTINUE

DO 74 I = 1, NCOL
    IF(DROP(I) .GT. DROPHI) DROPHI = DROP(I)
    IF(DROP(I) .LT. DROPLO) DROPLO = DROP(I)
    CONTINUE

NOW WRITE OUT THE VALUES ONTO 7 AND CONTINUE

WRITE(7) (DROP(K), K = 1, NCOL)

ENDFILE 7

WE NOW HAVE THE MINIMUM AND MAXIMUM DROP VALUES AND THESE ARE PUT IN SCALES(5) AND SCALES(5:24) RESPECTIVELY.

DO 454 K = 5, 24
    SCALES(K) = DROPHI
    CONTINUE

DO 771 IF(DROPHI .EQ. 0.0) GO TO 771
    IF((DROPHI + DROPLO)/DROPHI .GT. 2/3) PRINT(6, 778)
    CONTINUE

778 FORMAT(20X, 'WARNING', /, +/-5X, 'YOU HAVE A SCALING PROBLEM YOUR SYMVU PLOT MAY'/, ++++++5X, 'YOU HAVE EITHER A VERY HIGH (HIGHEST'/, ++++++5X, 'VALUE) OR A VERY LOW (LOWEST VALUE); SEE ABOVE, '/, ++++++5X, 'IF YOUR WHOLE DATA IS NOT EVENLY SPREAD BETWEEN THEM,'/ , ++++++5X, 'DURING SCALING THE VARIATION IN THE SURFACE MAY BE'/, ++++++5X, 'LOST,'/ , ++++++5X, 'THIS MAY BE OVERCOME BY CAREFUL SELECTION OF THE'/, ++++++5X, 'APPROPRIATE SYMVU ELECTIVES - SEE MANUAL- OR '/, ++++++5X, 'YOU MAY HAVE TO OPERATE ON YOUR MATRIX. TO DO THIS'/, ++++++5X, 'FIRST EXAMINE IT, BY LOOKING AT THE TAPE PRODUCED'/, ++++++5X, 'BY THIS PROGRAM, WITH PROGRAM GRID IN THIS LIBRARY'/)
SEE DOCUMENTATION-. THIS WILL SHOW YOU THE LOCATION /
OF THE TROUBLE. WITH THIS KNOWLEDGE YOU CAN START TO /
PRUNE OUT THE OFFENDING READINGS. GET HELP DOING THIS /
STOP
END
PROGRAM GRID (TAPES,PARM,OUTPUT, -TAPE6=OUTPUT)

PROGRAM WRITTEN BY F.W.M.HOPPER 1978.

THE PROGRAM IS WRITTEN SO THAT A PARAMETER SUPPLIED
IN THE CALL TO THE PROGRAM MAY BE USED IN THE PROGRAM.
THIS PARAMETER (PARM) ENABLES LABELING OF THE CONTOURS
TO BE TURNED OFF.
THE DEFAULT IS FOR LABELS TO BE PRINTED.
THE WORD *LABEL* USED AS SHOWN BELOW ALSO PRODUCES
LABELS
THE WORD *NOLABEL* STOPS LABELS APPEARING.

THE PROGRAM SHOULD BE CALLED WITH LIB6000,ID=LIBBED
ATTACHED, IF LABELS ARE TO BE TURNED OFF.
THE PROGRAM CALL SHOULD HAVE AS ITS SECOND PARAMETER
(IF INPUT IS SUBSTITUTED, OTHERWISE THE ONLY PARAMETER.)
ONE OF THESE WORDS. EG D-
GRID(NOLABEL)
ANY UNKNOWN PARAMETER WILL DEFAULT TO PRODUCTION OF LABELS.

THE PROGRAM TAKES A MATRIX ARRAY FROM OUR FILLIN
OR SYMAP AND CONTOURS IT. THIS REPLACES TRADITIONAL
SYMAP OUTPUT (LINE PRINTED).
THE OUTPUT MUST RELATE TO THE USERS ORIGINAL CO-ORDINATES
THUS *** D-

IF THE GRID WAS CONSTRUCTED D-
Y
I
I
I
I

A

-X

THE GEOGRAPHIC WAY .

WHERE A IS A VALUE AT REF X,Y

THEN IT MUST BE TRANSLATED BY ALTERING THE X,Y ASSIGNMENTS
WITHOUT ALTERING THE POSITION OF THE DATA.
THIS CAN BEST BE DONE BY WRITEING OUT THE MATRIX ,THEN
READING IT IN AGAIN.
MA=THE MATRIX
XM=MAX X VALUE
YM=MAX Y VALUE

DO 1 I=1,YM
IK=YM+1-I
WRITE(8) (MA(IK,J),J=1,XM)
1 CONTINUE
REWIND 8
DO 2 I=1,YM
READ(8) (MA(I,J),J=1,XM)
2 CONTINUE
NDX=YM
NDY=XM

MA=THE NEW MATRIX WHICH IS THUSO-
IN THE PROGRAM, O-
THE FIRST WORD READ (NDX)
IS -NUMBER OF ROWS IN THE GRID
THE SECOND O- (NDY)
IS -NUMBER OF COLUMNS IN THE GRID
THE THIRD O- (NDC)
IS -SCALING HOW MANY ROWS IN EACH UNIT
THE FOURTH O- (NDD)
IS -SCALING HOW MANY COLUMNS IN EACH UNIT

ND IS THE ARRAY TO BE PLOTTED IN THE SYMAP FORM AS DESCRIBED ABOVE.

THE PROGRAM IS WRITTEN TO UTILISE THE ULCC, DIMFILM ROUTINES.
THIS ROUTINE ACTUALLY DOES THE PLOTTING.

INTEGER ICVALS(18), FNAME

SET UP

REAL ND(NDY,NDX),CVALS(18),MIN,MAX
REAL NDA,NDB,NDC,NDD
COMMON/DIMCONS/A(18)
DATA ICVALS/5,7,9,11,13,15,17,19,21,22,23,24,25,26,27,28,29,30/
NDX1=NDX
NDY1=NDY
NTP=0

READ IN MATRIX ND CHECK FOR ERRORS INFORM IF ANY.
NDX1= THE NUMBER OF COLUMNS IN THE MATRIX TO BE CREATED, AND THE
NUMBER OF LINES IN THE MATRIX BEING READ IN.
NDY1= THE NUMBER OF LINES IN THE MATRIX TO BE CREATED, AND THE
NUMBER OF COLUMNS IN THE MATRIX BEING READ IN.

DO 13 JJ=1,NDX1
J=NDX1+1—JJ
READ(5) (ND(I,J) ,I=1,NDY1)
IF(EOF) GO TO 50
13 NTP=NTP+1
GO TO 100
50 WRITE(6,55) NDX1,NDY1,NTP
55 FORMAT ( ' ERROR ' , / , ' MATRIX DIMENSIONS FROM FILE HEADER WERE'/
+ 'I5' ' LINES BY 'I4' COLUMNS ' ,/ , ' END-OF-FILE ENCOUNTERED AF'/
+TER READ OF LINE 'I4', ' ERROR ' ,/)
GO TO 56
100 IF(EOF) GO TO 150
WRITE(6,58) NDX1,NDY1
58 FORMAT ( ' WARNING ' , / , ' MATRIX SIZE FROM FILE HEADER WAS ' ,/
+ 'I4' ' LINES BY 'I4' COLUMNS ' ,/ , ' BUT MORE DATA THAN THIS EX'/
+ISTS ON YOUR INPUT FILE. CONTOUR MAP IS THEREFORE BASED ON ONLY P'/
+ART OF YOUR INFORMATION',/ , ' WARNING ' ,/)
56 RETURN
150 CONTINUE

FIND OUT THE MAX AND MIN VALUES OF THE DATA

MIN=MAX=ND(1,1)
DO 10 I=1,NDY1
DO 10 J=1,NDX1
IF(MIN.GT.ND(I,J)) MIN=ND(I,J)
IF(MAX.LT.ND(I,J)) MAX=ND(I,J)
10 CONTINUE

FIND OUT THE RANGE THE DATA COVERS

RANGE=MAX-MIN

DIVIDE THE RANGE INTO 17 STEPS

STEP=RANGE/17.0

ASSIGN TO CVALS EACH OF THE STEPPED CONTOUR VALUES
FOR THE COMPLETE RANGE CONTOURING WILL OCCUR AT
THese VALUES.

CVALS(1)=MIN
DO 12 I=2,18
CVALS(I) = CVALS(I-1) + STEP
12 CONTINUE
C START THE DIMFILM PROCESS AND ACTUALLY PLOT THE THING.
C
CALL CAM35MM
A(1)=1.0
A(2)=0.0
IF (NDA.GT.NDB) CALL CINE
CALL BOUNDS(0.0,NDB,0.0,NDA)
CALL CHECK(-2)
CALL MARGIN(0.0)
CALL WINDOW(0.0,NDB,0.0,NDA)
CALL WINDFRM
CALL CINTERP(10)
IPAR = FNAME(4,LPARM)
IF (IPAR.EQ.4,LPARM.OR.IPAR.EQ.5,LLABEL) GO TO 1000
IF (IPAR.EQ.7,LLLABEL) GO TO 1001
CALL REMARK("*UNRECOGNISED PARAMETER *")
CALL REMARK(IPAR)
GO TO 1000
1001 CALL NOCLAB
1000 CONTINUE
CALL CONTR1(ND,NDY,NDX,18,CVALS,ICVALS)
CALL ENDFILM
RETURN
END
C SUBROUTINE CALCMP(X,Y,I,J)
C DATA NFRAM/0/
C IF (J.LT.0) GO TO 1
C IF (J.EQ.2) NFRAM=NFRAM+1
C RETURN
C1 IF (J.EQ.-2) I=NFRAM
C X=Y=0.0
C ENTRY STARTF
C ENTRY FINISHF
C RETURN
C END
THE STRUCTURAL AND METAMORPHIC DEVELOPMENT
OF THE BERGEN ARC/BERGSDALEN FORELAND
GNEISSES OF OSTERØY, S.W. NORWAY

by

Frederick William Metcalfe HOPPER

VOLUME 2

Thesis submitted for the Degree of
Doctor of Philosophy

Bedford College (University of London) February 1980
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VOLUME 2

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Figure 158 .................................... Back Pocket

Microfiche ..................................... Back Pocket
Plate 1
NORTHERN GNEISSES

D "Early", S "Early" and its accompanying banding. Note how it is sheared and crosscut by two younger, sub-horizontal pegmatitic veins.

Plate 2
NORTHERN GNEISSES

D1, tight intrafolial F1 isoclines. Usually the limbs of these folds cannot be seen as they are sheared out. In this case the exposure is formed from M shaped parasitic folds in a larger nose region.
Plate 3

NORTHERN GNEISSES

D1, an early D1 pegmatitic vein which crosscuts the composite layering at a high angle yet is slightly folded by later D1 deformation.

Plate 4

NORTHERN GNEISSES

D1, S "Early" folded by an F1 fold. Note that by the hand a zone of shear has developed in limb of a small paracitic fold. An early D1 pegmatite occupies this zone of displacement. The larger veins are younger still.
Post D1 veins which crosscut the main composite foliation at a high angle and are then folded in D2. Note it is crosscut by a younger vein running sub parallel to the foliation.

Two generations of post D1 and pre D2 veining. Note they crosscut the earlier composite foliation yet display a younger schistosity (S3) internally, indicating subsequent deformation. They are also garnetiferous.
Plate 7
NORTHERN GNEISSES

D2, F1 isoclines and D1 basic bodies folded round the nose of a relatively open F2 fold.

Plate 8
NORTHERN GNEISSES

D2, A quartzite exposure showing S2 folded round in an F2 fold with the start of a very weak axial planar schistosity (S3) developing. Note that the F2 folds are tighter in this exposure than in plate 7. The obvious fold is 1.5 metres high.
Plate 9

NORTHERN GNEISSES

D2, L2 exposed on an S1 foliation plane, outcrop 1.5 metres long.

Plate 10

NORTHERN GNEISSES

D2, F1 folds refolded in an F2 nose. Note axial planar pegmatitic vein in the F2 fold and the fact that these folds are tighter than in plates 7 or 8.
Plate 11
NORTHERN GNEISSES

D2, Close up of a refolded F1 fold from plate 10. Note the new schistosity (S3) running parallel to the late axial plane pegmatitic vein which crosscuts the F1 fold and the typical "limb region" pegmatitic veins either side of the F1 closure.

Plate 12
NORTHERN GNEISSES

D2, a large F2 closure folding S1 and S2 and a set of cross-cutting pegmatites which intersect the composite foliation at a low angle. These pegmatites are attributed to the pre D2 migmatitic episode.
Plate 13

NORTHERN GNEISSES

D2, late D2 pegmatitic vein cutting a paracitic F2 fold and the composite foliation it deforms, at a high angle. Note this vein is itself slightly folded in the D2 flattening.

Plate 14

NORTHERN GNEISSES

D2-D3, a post F2 amphibolite cut by a late pegmatite which has boudinaged in D2b and is subsequently folded by F3.
Plate 15

NORTHERN GNEISSES

D3, a large scale, open F3 fold affecting S3b from the highest ground on Osterfy. These folds are the only large scale folds visible in the Northern Gneisses.

Plate 16

NORTHERN GNEISSES

D3, Open F3 folds producing a strong large scale mullion rodding, L3. The regularity of L3 proves the cylindrical nature of the folds in this area.
Plate 17

NORTHERN GNEISSES

D3, A profile view of F3, showing the open geometry which is typical over most of Osterøy. Note these folds are only seen where S3b is well developed.

Plate 18

NORTHERN GNEISSES

D4, A shear zone formed in the middle limb of an F4 fold, in which the new S5 schistosity is well developed.
Plate 19

NORTHERN GNEISSES

D4, A large scale example of a D4 shear zone from the north of the island. Note how the layering is contorted on entering the zone yet regular in the middle, only now vertical rather than horizontal.

Plate 20

NORTHERN GNEISSES

D4, Profile view of a D4 shear zone, demonstrating how they form in the middle limbs of poorly developed F4 folds.
Plate 21

NORTHERN GNEISSES

A D₄ quartz vein in a D₄ shear zone. The quartz veins are sub-axial planar to the F₄ folds.

Plate 22

NORTHERN GNEISSES

D₄, Nose region of an F₄ fold. Note that the wavelength of the paracitic folds varies with the thickness of the layer folded and that the folds in one layer do not force a fold in the next layer provided there is a separation of one fold wavelength between the layers.
D "Early", D1 & D2, An F1 fold folding S "Early". This is then crosscut by a pegmatitic vein which is then folded by F2. Note the F2 fold has a sheared limb which started as one limb of the F1 fold. The shear zone is a band of secondary mylonite. Note also that there is a secondary trondhjemite vein intruded horizontally, crosscutting the S2 schistosity.

D "Early" & D1, the general shape of an F1 fold folding S "early" showing a thinned limb & a pegmatitic vein which crosscuts S "Early" at a shallow angle (The leucocratic layers defining S "Early" may also once have been intrusive veins, some of these are now boudins and the whole texture of the gneiss in the F1 limbs is mylonitic).
D "Early" - DI, S "Early" folded round an F1 fold. Note the fold is crosscut by an axial planar pegmatitic vein and that S "Early" itself contains pegmatitic veins, some of which are boudinaged. There is also a small F3 fold represented in the top right corner and the rocks are generally mylonitic, the mylonite produced during DI.

DI, Intrafolial, isoclinal F1 folds just depicted by subtle changes in the composition of S "Early". Note the spindle shape of the quartz grains on the weathered surface. These are the product of S2 crosscutting the rodded shape fabric L1 produced by the interaction of DI with S "Early". This rock is a blastomylonite.
Plate 27

SOUTHERN GNEISSES

DI, The L1 lineation as exposed on planes of S1 schistosity.

Plate 28

SOUTHERN GNEISSES

DI, K-feldspar porphyroclasts in the gneisses. Seen in three dimensions it can be seen that these define an L tectonite rodding structure, the expression of an extreme prolate strain ellipsoid. The lineation this defines is L1.
Plate 29
SOUTHERN GNEISSES

D1-D3, The L1 lineation, here refolded by an F3 fold, is seen to be a mullion structure as well as a shape fabric, plates 27 & 28.

Plate 30
SOUTHERN GNEISSES

D1-D2, Dissociated F1 fold noses shredded into an amphibolite by the D2 shearing. The leucocratic portion formed strings and veins in the amphibolite prior to F1. Note: compare this plate with that of a similar mylonite described by Higgins 1971.
Plate 31

SOUTHERN GNEISSES

D1-D3, Type 3 interference pattern (Ramsay 1967), produced by the interplay of F1 refolded by F3.

Plate 32

SOUTHERN GNEISSES

D3, A fine scale banding between amphibolitic and quartzofeldspathic material displaying highly non-cylindrical F3 folds very well. Note the axial directions vary by as much as 120° over 10 cm. distances along the hinges.
D3, An S1 foliation plane folded by F3, showing the non-cylindrical nature of these folds, the relationship of the paracitic folds to the main fold and to each other.

D3, The truncated nose of a non-cylindrical F3 fold which could be misinterpreted as a Type 1 interference pattern (Ramsay 1967). Note L1 running from top right to bottom left, is folded round the nose.
D3. A broad S1/S2 foliation plane on which small F3 buckles are seen in an en echelon pattern. Note that each fold is nucleating quite separately, spaced out in the layer, and that all the folds are non-cylindrical.

D3. The S3 crenulation cleavage developed in a mica schist, the only lithology in which it is seen. This is axial planar to the F3 folds and its outcrop on the S1/S2 foliation planes in the noses of the F3 folds defines L3.
Plate 37
SOUTHERN GNEISSES

D3, Hydraulic fracture network filled with calcite. It is developed in a D1 mylonite band with an excellent F1 fold nose exposed in the middle.

Plate 38
SOUTHERN GNEISSES
AND BERGEN ARCS

D4, A view across the Northern part of the Bergen Arcs from Osterøy showing the large scale broad synformal F4 fold, which affects the dominant foliation planes as seen from the dip slopes of the distant hills.
D5, A view showing the scale and importance of the late jointing affecting the North Osterøy gneisses.

D5, A North-South joint, showing the mineralisation developed along these structures. Note the calcite veins, general epidotization and sharp planar walls. It is their alteration that causes them to weather out as such marked gullies.
Plate 41

NORTHERN GNEISSES
& SOUTHERN GNEISSES

A view from the slopes of Hogafjell south across the Northern and Southern Gneisses, showing the best and most obvious exposure of the Tysse thrust, the contact between them.

Plate 42

Least deformed specimen in a sequence (Pl.42-45) of mylonitized augen gneiss showing reduction in porphyroclast size with increasing deformation.
Second specimen in the (Pl.42-45) sequence of mylonitized augen gneiss.

Third specimen in the (Pl.42-45) sequence of mylonitized augen gneiss.
Most deformed specimen in the sequence (Pl. 42-45) mylonitized augen gneiss showing reduction in porphyroclast size with increasing deformation.

Quartz laths or deformation lamellae subgrains which define a new axial plane schistosity to the folds in the mica layers.
Hydraulic fractures filled with rock flour produced during the escape of the fluid phase. Note angular fragments of rock and crystals set in an isotropic matrix; and sharp sides to the bands of mylonitic material.

Fragments of rock in a breccia cut through by calcite veins, formed by hydraulic fracture, set in a matrix of hydraulic cataclasite with a vast range of fragment size.
Part of a late breccia zone from a joint. Note the voids rimmed by radially grown drusy mosaics of quartz crystals. The presence of the voids in the rock indicates the pp. H₂O exceeded or equaled the load pressure.

A hydraulic cataclasite, note that there is no foliation while the rock is annealed, with angular crystal and rock fragments set in an isotropic dusty matrix.
A good example of a K-feldspar overgrowing its surrounding matrix in an "amoeboid" manner. The poikilolitic margins of the porphyroblast prove secondary growth of the K-feldspar. The feldspar is not perthitic indicating it developed late in the metamorphism.

A braid perthite with poikilolitic inclusions of quartz, proving that this K-feldspar is a secondary feature in the rock, which crystallised over pre-existing matrix minerals.
A poikilolitic hornblende crystal; the inclusions of quartz and sphene are not portions of a single crystal and the hornblende must therefore replace a pre-existent mineralogy which was finer grained than that presently seen.

Quartz segregation ribbons with secondary \((D_4)\) deformation lamellae at an angle to them. Note the ribbons lie sub-parallel to the foliation and have sharp margins which cut across matrix crystals — they are therefore a secondary development.
A zoned garnet containing inclusions of quartz and graphite. A large distorted knot of graphite shows where pressure solution (of matrix) has occurred as the schistosity was flattened round the garnet.

A rotated garnet crystal, enveloped by matrix with a strong later schistosity.
Undulose extinction in a plagioclase crystal. Note the slight bending of the twin lamellae.

A microcline porphyroclast with a rim of coarser crystalline matrix in its pressure shadow. Note the patch perthite texture.
Microperthitic microcline. The fact that this microcline is not coarsely perthitic is taken to indicate that it grew later in the metamorphism than those which are.

In the bottom of the plate relict biotite crystals are seen overgrown by the leucocratic minerals; this area forms the margin of a small "tectonic pip." The top portion of the plate shows later recrystallised biotites which overgrow the leucocratic matrix.
Embayed biotite - biotite cut, overgrown and embayed by the leucocratic portions of the rock, leaving interstitial relics of the biotite.

Epitaxy - rimming of allenite by clinozoicite and epidote. In proximity to the metamict allenite the epitaxial epidotes are also metamict.
A relict plagioclase crystal associated with an amphibole knot - now overgrown by quartz and strongly normally zoned (Na rich rims).

Myrmekite corrosion. The rims of K-feldspar crystals are invaded by myrmekite (these may be stripped off during deformation, leaving isolated fragments of myrmekite in the matrix, as in Plate 95).
A large recrystallized K-feldspar which has enclosed crystals of the matrix at its margins. The margin is crenulate with "pseudo-podia" growing out from it into the matrix. Note this K-feldspar is not particularly perthitic, indicating that it was formed late in the metamorphism.

A small scale shear zone in an amphibolite. Note the sharp reduction in crystal size in the zone and the presence of "strings" of quartz in them. These were segregation "ribbons" which are deformed during shear to form sinuous "strings".
A K-feldspar strongly corroded by myrmekite round the margins. This myrmekite rimming increases through the M3 retrogression. Note characteristic "cauliflower" growth form of the myrmekite centred at numerous sites round the margin.

A K-feldspar (top) and a late Na-feldspar (bottom) showing the difference between the two. The K-feldspar has an amoeboid growth form and shows myrmekite invasion, while the Na-feldspar is full of chlorite and epidote inclusions. Both crystals are porphyroblasts.
A late Na-feldspar (albite) porphyroblast showing typical inclusion of unorientated mica, chlorite and epidote. Such crystals are typically, a late product of the M3 metamorphism.

A late albite porphyroblast full of epidote and chlorite inclusions, overgrowing biotite and hornblende in an amphibolite.
Plate 71

An example of typical un-mylonitized acid gneiss from the Northern Gneiss Unit.

Plate 72

Example of moderately mylonitized acid gneiss from the Southern Gneiss Unit. Note the F5 fold has no accompanying schistosity.
Plate 73

Highly mylonitized gneiss from the Southern Gneiss Unit showing characteristic banding and development of K-feldspar augen. This rock would be called a blastomylonite.

Plate 74

Highly mylonitized gneiss from the Southern Gneiss Unit, showing the characteristic strong linear fabric seen on the schistosity surfaces (Ll).
A "pip" of unmodified gneiss set in a younger mylonitized matrix. The "pip" contains an old texture and banding that shows no causal relationship to the younger texture and banding developed in the matrix surrounding it.

A K-feldspar broken up by stresses which have produced undulose extinction, sub grain formation and rotation of some of the sub grains. This is an example of the start of the process of deformation of K-feldspar porphyroblasts that is completed in Plates 80 and 81.
A crushed and boudinned sphene crystal, now much replaced by FeO and spread out in the plane of the schistosity.

An example of rod perthite. Once this was a large crystal that was deformed in a manner similar to that shown in Plate 76 so that the sub grains were rotated. Subsequently however, the sub grains have regrown to form 3 larger sub grains in slightly differing orientations. Eventually these will reunite to form a new crystal equal or larger in size than the original. Note that the crystals are poikilolitic.
Plate 79   x 15

An example of a K-feldspar crystal showing undulose extinction. This is the first stage of sub grain formation in a porphyroclast and precedes the stage shown in Plate 76.

Plate 80   x 10

An augen of K-feldspar composed of numerous sub grains all in differing orientations. Note that the augen is beginning to be spread into the plane of the schistosity and that the K-feldspar crystals are still much larger than the crystals comprising the matrix. This Plate represents a more advanced stage of deformation of what was originally a single crystal than that shown in Plate 76.
An aggregate of K-feldspar crystals forming a very thin augen spread in the plane of the dominant schistosity. This lens shape is a further advancement of the processes described in Plates 79, 76, and 80 respectively and represents the remnants of a very strained porphyroclastic crystal of K-feldspar.

A deformed K-feldspar porphyroclast such as that described in Plate 80 or 81 which has subsequently started to recrystallize. The large black portion in the middle of the augen is the re-crystallising portion and is overgrowing the surrounding sub grains as seen from the crosscutting relationships it has with them. Eventually the crystals will grow to be as large or larger than the original K-feldspar as shown in Plate 83.
A perthitic microcline which has recrystallized from an aggregate of sub grains (such as in Plate 82), to form a new crystal which is of the same size as that which occupied the site prior to the deformation producing the strains seen in the matrix (augen structure round the crystal - Plate 84). The fact that this is a recrystallization product is deduced from the lack of internal strain features and the manner in which the crystal crosscuts and overgrows its marginal neighbours. Note the large hydraulic fractures across the crystal.

A p.p.l view of the crystal shown in Plate 83, showing how the schistosity enveloping the porphyroblast is in fact overgrown, proving secondary recrystallization.
A K-feldspar porphyroblast which has overgrown its matrix during recrystallization. Note inclusions of individual matrix crystals, the crenulate margin and the extension "pseudopodia" of the K-feldspar crystals into the matrix. Such a growth form is said to be ameboid and is characteristic of a porphyroblastic crystal.

A K-feldspar porphyroblast which has enlarged during re-crystallization, overgrowing its matrix "sheath". Note the "ameoboid" outlines and poikilolitic nature. The large quartz grain to the right forms part of the old strain shadow.
An "amoeboid" margin to a K-feldspar porphyroblast, showing the process of inclusion of matrix grains during recrystallization. Note these grains are discrete crystals, not like a crystal intergrowth (symplectite texture).

Undulose extinction in quartz (This is a precursor to sub grain formation).
Quartz laths defining a strong schistosity which runs at a shallow angle diagonally across the plate. The laths were produced from older strained (Plate 88) quartz grains, by annealing and recrystallization processes.

Example of matrix from the mylonitic gneissic rocks. Note the irregular grain size, irregular crystal morphology and the lack of good 120° junctions (Though these can be seen to be beginning to form in places).
Recrystallized matrix of a mylonitic gneiss, compared to Plate 90 this matrix is now relatively strain free, the crystal size is larger, the crystals are more regular and 120° junctions are common. An annealed matrix.

Plate 92

Mylonitic banding, typical of the Southern Gneiss Unit rocks.
Plate 93

Gneiss in which mylonitization is extreme, from a D4 mylonitic band. Note virtually all the K-feldspar augen are destroyed.

Plate 94

Typical hand specimen of mylonitized gneiss from the Southern Gneiss Unit.
Thin bands of mylonite forming in a biotite rich layer adjacent to a porphyroblast. Note sharp margins and grain size contrast between the more leucocratic minerals inside and outside the zone of shear.

Myrmekite shredded from the rim of a K-feldspar augen. This myrmekite cannot have formed in situ because it is surrounded by minerals other than feldspar. Note it is overgrown by a late apatite crystal.
Plate 97  x 10

An albite porphyroblast overgrowing a fine grained micaceous groundmass. Note it is poikilolithically including relics of the micas.

Plate 98  x 15

A shear band cutting across the dominant schistosity at a high angle. Such bands, which reveal extreme local deformation, are not disruptive. They must therefore have formed by plastic deformation. This band is a zone of simple shear similar to those described by Ramsay and Graham 1970.
A fold train from Osterød. Note the progressive tightening of the folds from left to right and the accompanying development of a mylonitic banding as the folds get tighter.

Plate 100 x 10

A relict garnet porphyroblast now overgrown by feldspar. Note that the garnet is fractured with numerous fracture planes prior to the feldspar growth. This fracturing is interpreted as a cataclastic texture resulting from very high strain rates. Note too the later inclined fractures which cut the feldspar.
Quartzite showing crenulated crystal margins which are the result of annealing and recrystallisation of what was a finer grained and more highly strained rock. The large surface areas of the crystals are possible because the adjacent crystals are of the same species and the surfaces are therefore of lower energy than they would otherwise be.

A late albite porphyroblast overgrowing a fine grained matrix but aligned with the long axis parallel to the dominant schistosity (S5). Note it is poikilitic, full of inclusions from the matrix (which it has included by a process similar to that for Pl.78) and epidote and chlorite laths. Note too its amoeboid crenulated outlines.
A perthite showing compound texture. Note that this crystal shows none of the crenulated marginal textures which indicate that it overgrew the present matrix; indeed the margins are sharp. The augen is therefore a porphyroclast and is interpreted as a relic of the M3a paragenesis.
Ball bearing box model, modelling a plane through a crystal(s). In this case the crystals are relatively unstrained and the model represents an annealed rock which has been slightly restrained.
Ball bearing box model, modelling a plane through a crystal(s). In this case the model represents a section through a highly strained rock and shows the high energy situation while Pl.104 shows the low energy situation.
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Structural data added as Microfiche in vol. 2.
Fig 158.

**Representative Cross**

Key as for Map 1
THE GEOLOGY OF NORTH OSTERØY, NORWAY