STRUCTURAL STUDIES IN
THE JOTUNHEIM AREA,
SOUTHERN NORWAY.

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Thesis presented for the Degree of DOCTOR OF PHILOSOPHY of the UNIVERSITY OF EDINBURGH in the FACULTY OF SCIENCE.

1967.
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I. INTRODUCTION.

The area studied in this thesis is in the Caledonides of southern Norway (see location and geological map in pocket). The Caledonian mountain chain forms the western part of the Scandinavian Peninsula. Its eastern margin is defined by nappes overthrust onto autochthonous Cambro-Ordovician sediments which in turn rest on the Pre-cambrian rocks of the Baltic Shield. The western border of the Caledonides is not exposed and rocks found at the west coast of Norway belong to the central part of the mountain belt (Strand, 1961).

In the southern Norwegian Caledonides there are many problems warranting investigation and the author carried out detailed mapping to investigate the following:

i. Elucidation of the structural sequence in the area mapped. Comparison of these sequences regionally in an attempt to make regional structural correlations. Particular emphasis has been placed on the sequence of nappe emplacement.

ii. Investigation of the nature of the conformable contact between rocks of the Basal Gneiss Area (the large expanse of gneisses extending north of the Jotunheim mountains and west of a line between Vágámo to Surnadal, to the North Sea) and Caledonized metasediments to the south and east.

iii. Regional Stratigraphic Correlations. In southern Valdres and in the Mjøsa district fossils are common, the rocks are relatively/
relatively undeformed and unmetamorphosed, and in consequence, stratigraphic correlation is relatively certain. This is not so in northern Valdres, the Jotunheim and north to the basal gneiss area where correlations must be made solely on the basis of lithologic comparisons. In this respect, a knowledge of the structure is imperative.

iv. Recognition of a root zone for the Upper Jotun Nappe. There is no obvious root exposed at the present level of erosion between the nappe and the North Sea. Possible root zone sites are west of the present Norwegian mainland, or within the large syncline in which the nappe is now situated.

In investigating these problems, the author has done detailed mapping at two places. Lom where a complete sequence of rocks from the Upper Jotun Nappe to the Basal Gneiss Area is exposed, and in Sjodalen where relationships between high grade crystalline nappes, the Valdres Sparagmite and the Cambro-Ordovician sequence are seen.

To enable the reader to appreciate the regional significance of the results of this mapping, it is necessary to give an account of relevant previous research. Following is a review of stratigraphy and structural evolution of the Caledonides of southern Norway.

**STRATIGRAPHY OF THE CALEDONIDES OF SOUTHERN NORWAY.**

1. Introduction.

In Table 1 (pocket) are presented the stratigraphic
and structural-stratigraphic sequences for 6 areas in the southern Norwegian Caledonides and one area (Oslo region) on the foreland. Distinct variations in lithology from region to region are easily seen. In general, thickness of Edicambrian and Cambro-Silurian increases in a north to north-westerly direction. Miogeosynclinal Cambro-Silurian rocks at Oslo and Mjøsa are interbedded limestones, fine grained sandstones and shales. In the west and north-west, the same miogeosynclinal sequences become more coarse grained and dominantly clastic (Holtedahl, 1960, p.172).

Miogeosynclinal and eugeosynclinal sedimentary sequences, approximately contemporaneous in age, have been set up in various parts of the area under discussion (Sel-Vågå, Leirdalen-Hestreppigga areas). Eugeosynclinal rocks are everywhere allochthonous (Strand, 1961) and in many cases brought into direct contact with miogeosynclinal rocks by thrusting. Norwegian geologists have used the following criteria to distinguish the two groups of rocks:

i. Presence or absence of Caledonian volcanics and related intrusives. These, occurring only in the eugeosynclinal rocks, include peridotites (some altered), gabbros, amphibolites, and trondhjemitic intrusions.

ii. Sedimentary rock type. Eugeosynclinal rocks are metamorphosed equivalents of graywackes and shales with no extensively developed limestones. Miogeosynclinal rocks contain no graywackes and, in the south, are limestone rich.
In the Oslo region, Mjösa district and southern Valders, miogeosynclinal rocks are in authochthonous position overlying rocks of the Baltic Shield. In more northerly areas, however, (Sjodalen, Sel-Vågå, Lom, Hestbreippigan-Leirdalen areas) because of structural complications it is not possible to state whether or not miogeosynclinal rocks are allochthonous or autochthonous, Strand 1951 has used the term paraautochthonous for these doubtful rocks.

2. Stratigraphy.

Eo-Cambrian Rocks. The oldest sediments in the Norwegian Caledonides belong to the Sparagmite group, (sparagmite is a Norwegian term for feldspathic quartzites and arkoses) best developed in the Mjösa district (Skjereth, 1963). Total thickness is more than 1450 m (base not exposed) and the sequence is autochthonous. These unfossiliferous rocks underlie fossiliferous lower Cambrian rocks, and to indicate their close time relationships to the Cambrian, they have been termed Eocambrian (Brögger, 1904). The typical sparagmite has been derived from granite and related rocks and its high feldspar content suggests rapid deposition in dry conditions. Hossack (personal communication) has found ventifacts in the basal conglomerates of the sparagmite succession at Grönsenknipa.

Cambro-Silurian Stratigraphy. Details of the Cambro-Silurian stratigraphy can be seen in Table I. Rocks in the Oslo and Mjösa districts are fossiliferous, folded but unmetamorphosed.
In Valdres, 75 km. to the north and west of Mjøsa rocks differ in three respects from those to the south-east.

i. They are coarser grained and contain relatively more terrigenous material.

ii. They are low grade metamorphics deformed to the extent that fossils are preserved only in a few localities.

iii. They are in allochthonous position. (Strand 1951; Holtedahl 1960; Skjeseth, 1963).

Correlation of Cambro-Ordovician stratigraphy can be made between Oslo, Mjøsa and Valdres. Correlation of younger rocks, though, is more doubtful. In the Valdres district there is a continuous sequence above the Phyllite formation consisting of the Mellsenn and Valdres Sparagmite formations. The Mellsenn has been dated (fossils: Mellene) as middle Ordovician and the upper part of the Valdres Sparagmite is considered to be lower Silurian. (Holtedahl, 1960). There are no upper Ordovician rocks in the Mjøsa district. Lower Silurian ortho-quartzite lies immediately above middle Ordovician Mjøsa limestone, (Skjeseth, 1963). At Oslo the sequence is also incomplete with a hiatus at the middle-upper Ordovician junction and a slight angular unconformity at the Ordovician-Silurian junction (Holtedahl, 1960).

In the Valdres district, the Valdres Sparagmite lying above the Mellsenn consists of light coloured arkoses passing upwards into conglomerates. (Holtedahl, 1960). At Røssjokollan/
Røssjokollan, 25 km. east of Mellene, a gabbro massif is thrust over the Mellsenn formation. The nappe is overlain by Valdres Sparagmite containing a conglomerate composed of boulders of the same type of gabbro forming the massif. (Strand and Holmsen 1960). This observation and others similar to it led Goldschmidt (1916a) initially and later Strand (1951, 1957, 1959, 1960, 1961, 1962) to conclude that there had been emplacement of a high grade crystalline nappe, the Lower Jotun Nappe, over the Cambro-Ordovician sequence. Erosion contemporaneous with, and after emplacement of this nappe produced the Valdres Sparagmite. In some places as at Mellene the Sparagmite was deposited directly on Cambro-Ordovician rocks, elsewhere it was deposited on the nappe. The similarity of detritial feldspar (perthites) in the Valdres Sparagmite to that in the nappe is a consequence of the nappe being the source of the sparagmite. In northern Valdres the Sparagmite is overlain by a large high grade nappe emplaced after deposition of the Valdres Sparagmite. It has been termed the Upper Jotun Nappe.

The lower Silurian ortho-quartzite (p. 5) exposed in the Mjøsa district contains feldspar similar to those in the Valdres Sparagmite and these formations have been correlated by Skjæreseth (1963). Because the ortho-quartzite is overlain by shale and sandstone devoid of perthites, it marks the end of deposition of the Valdres Sparagmite in the Mjøsa district. Deposition of Sparagmite is believed to have ended in the Valdres district/
district at the same time (Holtedahl, 1960). There is no continuity of outcrop between the two areas.

The period of non-deposition in the Mjøsa district and hiatus and angular unconformity in the Oslo region (p. 5) are attributed to up-lift during the orogenic phase when the Lower Jotun Nappe was emplaced (Skjeseth, 1963).

Recently, evidence has been found which casts doubt on this interpretation of the Valdres Sparagmite. Strand and Holmsen (1960, p.31) state: "Current-bedding shows that the Mellsenn is inverted in the type-section in the south slope of Mellene. . . . The Mellsenn formation must . . . have a tectonic contact with the underlying Phyllite formation and its relation to the Valdres Sparagmite needs to be cleared up by future research." Using this evidence, Kulling (1961) proposed that the Valdres Sparagmite was Eocambrian in age, and it and the underlying Mellsenn formation are part of a large nappe extending throughout the Valdres district. The nappe has been thrust over the Ordovician Phyllite formation. Strand (1962) discounting younging evidence, still believed that the Valdres Sparagmite was structurally and stratigraphically above the Mellsenn formation. He considered the former as "...an Ordovician or Silurian deposit in para-autochthonous position rather than an Eo-cambrian deposit belonging to a far travelled nappe."

More recent work (Loeschke, 1967; Nichelsen, 1967; Hossack, personal communication) supports Kulling's view that the Valdres Sparagmite is part of an inverted nappe and is/
is probably Eo-cambrian in age, equivalent to the Sparagemite Group.

At Grønsennknipa (Hossack, personal communication) a nappe composed of granite, and meta-sediments which young off the granite, is thrust discordantly over Cambro-Ordovician phyllite. The meta-sediments include a basal conglomerate passing upwards into sparagmite. Previously the granite was considered as part of the Lower Jotun Nappe and the meta-sediments were part of the Valdres Sparagmite (Høltedahl, 1960). Hossack prefers to consider the granite as the Pre-cambrian basement upon which Eo-cambrian rocks were deposited.

Nickelsen (1967) in a regional study of the Valdres Sparagmite in the Valdres district has observed abundant primary structures (cross-bedding oscillation ripple marks) in the Valdres Sparagmite indicating overturning to the south in the lower limb of a large recumbent anticline. With these primary structures, he was able to use bedding-slaty cleavage relationships throughout the Valdres outcrop to distinguish inverted from un-inverted rocks. He thinks the Valdres Sparagmite is Eo-cambrian as:

i. It youngs off the basement at Grønsennknipa and Røssjokollan.

ii. Identification in the type-section at Mellene of a tilloid conglomerate at the stratigraphic top of the Valdres Sparagmite which can be correlated with the Moelu tilloid conglomerate (see Table I) near the top of the Eo-cambrian succession in the/
the Mjøsa district (see also Loeschke, 1967).

iii. The inverted Meløsenn formation can be correlated with uninvited Cambrian.

iv. Great sparagmite thickness, more than 3000 m.

v. The contact between Valdres Sparagmite and Cambro-Ordovician phyllite is everywhere a thrust (the Valdres thrust. This thrust had previously not been observed at Melle, the type locality of the Meløsenn and Valdres Sparagmite formations).

An important implication of this conclusion is that the Valdres Sparagmite was not deposited in the period between emplacement of the Upper and Lower Jotun Nappes. Nichelsen (1967) prefers to explain high grade crystalline rocks formerly termed Lower Jotun Nappe as cores of recumbent south closing anticlines in which the basement was folded along with Eocambrian rocks.

Stratigraphy North of the Jotunheim Mountains: Lithologically similar rocks to the Eocambrian and Cambro-Ordovician sequences found in Valdres can be seen north of the Jotunheim mountains (see Table 1). In the Sel-Vågå area a basal group of light sparagmites is overlain by phyllites and carbonaceous schists which resemble Eocambrian and Cambrian rocks to the south.

Correlations between the Sel-Vågå area and the Leirdalen Nettosester-Hestbreiggan area 65 km. to the west have been made by Cowan (1966) and Banham and Elliott (1965) but are hampered by the lack of knowledge of intervening ground. These/
These writers have correlated eugeosynclinal rocks in the Leirdalen-Nettoseter-Hestbrepiggan area with eugeosynclinal rocks in the Sel-Vågå area. There are important differences in the stratigraphy of these rocks in the two areas, especially the stratigraphic position of volcanics.

3. Conclusions.

In general, regional variation in type and amount of Eo-cambrian and Cambro-Silurian sediments is well documented for the southern Norwegian Caledonides. There is a north-west increase in clastic content and increase in grainsize and thickness of Cambro-Silurian rocks. Unstable landmasses exposed for short periods in the centre of the mountain belt, and north-west of present exposures of Cambro-Silurian rocks, have been postulated as the source of sediment. (Holtedahl, 1960).

In detail the position is more complex and suggested stratigraphic correlations suffer from the lack of detailed work done in this part of the Norwegian Caledonides.

A sequence of basal sparagmites overlain by finer grained sediments (shales, slates, sandstones) is seen at Mjøsa, the Valdres district, and north of the Jotunheim mountains in the Sel-Vågå area. Fossils are found only in the Mjøsa district where basal sparagmites are Eo-cambrian and overlying rocks are Cambro-Silurian. Correlations made to unfossiliferous areas (most of the Valdres district and all areas north of the Jotunheim mountains) are solely on the basis of lithology. There/
There is no continuity of outcrop of Cambro-Ordovician rocks between Mjøsa and Valdres (Eo-cambrian sparagmite intervenes) nor between Valdres and areas north of the Jotunheim mountains (the Otta and Upper Jotun Nappes intervene). The possibility of regional transgressions and facies changes, especially with respect to basal sparagmitian rocks, is sufficient to cast strong doubt on age correlations made by Banham and Elliott (1965) and Cowan (1966) in the Leirdalen-Nettoseter-Hestbre-piggen area, and Strand (1951a) in the Sel-Vågå area, with areas south of the Jotunheim mountains.

The correlation by Strand (1951a) of eugeosynclinal rocks in the Sel-Vågå area with eugeosynclinal rocks of the Trondheim region is open to question. Correlation of such successions can be made only over short distances because of variability of these sequences. Also, the structure of intervening ground is not sufficiently well known (see Holtedahl, 1960 p. 202-203) who states: "A solution of the tectonic problem of the Trondheim region must...be left to future work, we do not know at present whether rocks belong to one or more tectonic units," to allow such a correlation. The correlation (p. 9) of eugeosynclinal sequences made by Banham and Elliott (1965) and Cowan (1966) is open to the same criticism.

The Valders Sparagmite must now be considered Eocambrian in the light of recent work done by Nichelsen (1967) (see p. 7-9, contrast with ideas of Goldschmidt and Strand, p. 6).
Strand (1951a, 1961, 1964a) extrapolated his concept of the Valdres Sparagmite (synorogenic flysch deposit) to the Sel-Vågå area and west along Ottadal and Boverdalen to Leirdalen. This work cannot be accepted as:

i. Rocks Strand (1951a, 1955) interpreted as "gabbro conglomerates of the Valdres Sparagmite" have been shown by Dietrichson (1957) to be augen phyllonites developed from the Upper Jotun Nappe during its emplacement.

ii. There are no rocks in Leirdalen (Cowan, 1966) resembling the type Valdres Sparagmite of the Valdres district.

Part estimates of the areal extent of the Valdres Sparagmite must be revised and the idea of it being a synorogenic flysch deposit derived from crystalline nappes emplaced in Caledonian times is not tenable as the Valdres Sparagmite is Eo-cambrian in age.

STRUCTURE OF THE CALEDONIDES OF SOUTHERN NORWAY.

1. Introduction.

Thrust tectonics are remarkably well developed in the Caledonides of southern Norway, the mountain chain border being defined by the eastern edge of a sequence of nappes overthrust onto/
onto autochthonous Cambro-Ordovician sediments. The concept of overthrust nappes was introduced into the Norwegian Caledonides by A Tornebøhm, a Swedish geologist, in 1888 and this concept formed the basis of his definitive work - "Grundstrogen af det centrala Skandinaviens berglynod" - published in 1896. Brögger (1893) was unable to accept Tornebøhm's ideas and interpreted gneisses thrust over Cambro-Ordovician rocks in the Hardangervidda district as the product of an upward increasing metamorphic grade. The metamorphic grade supposedly was due to overlying intruded igneous rocks, now removed by erosion. However, Tornebøhm's tectonics were accepted by Björlykke (1902) for the Hardangervidda district. He recognized mylonites at the thrust between gneisses and metasediments. Later, Björlykke (1905) reverted to Brögger's (1893) views and thought the mylonites signified only local movements. Goldschmidt (1916b) recognized that Bergen-Jotun kindred rocks (The Upper Jotun Nappe) had moved on thrust planes, but thought transport distances were not great, the nappe being rooted in the Faltungsgraben, a large synform that the rocks now occupy. Holtedahl (1936) suggested that the Upper Jotun Nappe was not rooted in the Faltungsgraben but had been thrust from the north-west. Thrusting of Eocambrian sparagmite over Cambrian alum shales at the south-east margin of the mountain belt was demonstrated by Schiötz (1902) and Holtedahl/
Holteådahl (1915).

The nappe concept is now generally accepted for the southern Norwegian Caledonides. The main criteria used to define a thrust are:

1. Mechanical contacts between rock units.
2. Differences in petrographic constitution and sedimentary and/or metamorphic facies across the thrust (Holteådahl, 1960; Strand, 1961).

Strand (1961) recognized two groups of Caledonian nappes in southern Norway. There is a group in a lower tectonic position composed of Eocambrian and Cambro-Ordovician rocks of miogeosynclinal facies with, in some cases, a slice of basement at their base. Nappes of this group have moved eastwards or south-eastwards on flat thrust planes. The rocks in lower nappes of this group, found closest to the eastern edge of the mountain belt, are unmetamorphosed and little deformed; rocks in higher nappes are metamorphosed and deformed.

The second group is tectonically above the first and consists of nappes carrying rocks of eugeosynclinal facies, also in some cases with a slice of basement at their base. Because of their facies, these nappes are supposed to have originated to the west of the first group.

With this general picture in mind, I will go on to describe the important aspects of the structural geology in the areas listed in Table I.

2. Regional/
FIG. 2: THE RINGSAKER INVERSION.

S. BERGSVARDEN BIRI

CAMBRO-ORDOVICIAN SUCESSION.

- SILURIAN
- ORDOVICIAN
- CAMBRIAN

EO-CAMBRIAN

- QUARTZ SANDSTONE
- EKRE SHALE
- MOELV SPARAGMITE
- BIRI ST. SH. AND LST.

PRE CAMBRIAN

- 

5KM.
2. Regional Descriptions.

Oslo Region. The Oslo region extends from Langesunds-
fjorden to the southern end of Lake Mjøsa in the north. Cambro-
Ordovician sediments have been folded by décollement along the
lower Cambrian alum shales. Basement and sediments below
the alum shales remained undeformed. The folded rocks have
been considerably shortened on axes trending east to north-
east. Intensity of folding increases northwards.

Mjøsa District. The details of the south-east margin
of the lower group of nappes (p. 14) can be seen in the Mjøsa
district. (section Bergsvarden to Biri, Fig. 2).

In this section, Eo-cambrian rocks of the Quartz-
Sandstone Nappe are thrust over autochthonous Cambrian sediments.
Movement has been concentrated in Cambrian alum shales and the
Eo-cambrian Ekre shale. Between here and Redalen, schuppen
(imbrication) structures are common causing duplication of
outcrops of Cambrian shales. At Redalen is the Ringsaker
Inversion, a syncline-anticline pair folding rocks of the
Quartz Sandstone Nappe. The "Inversion" consists of an
isoclinal syncline (containing in its core Cambro-Silurian
sediments in stratigraphic sequence above the Eo-cambrian) with
an inverted north limb bordered on the north by an isoclinal
anticline. Axial planes dip moderately north. The north
limb of the anticline is replaced by a slide along which the
Biri nappe has been emplaced. The Quartz-Sandstone Nappe
contains the upper three formations of the Sparagmite group
(see Table 1). The Biri nappe contains all recognized
divisions/
divisions of the Sparagmite group. The base of the Biri nappe is not exposed and the amount of transport is not known. (Holtedahl, 1960; Skjeseth, 1963).

The Valdres District and Bygdin: In the Valdres district the Quartz-Sandstone Nappe consists of Eo-cambrian and Cambro-Ordovician rocks in normal stratigraphic succession. This nappe is thrust over thin (10 - 50 m, Strand and Holmsen, 1960) autochthonous Cambrian alum shales. Effects of metamorphism and deformation in the rocks of this nappe increase northwards (Strand, 1955b; Holtedahl, 1960). Above the Quartz-Sandstone Nappe is an inverted nappe composed of Eo-cambrian and Cambrian rocks of different facies than rocks in the Quartz-Sandstone Nappe (Nichelsen, 1967). The inverted nappe is the lower limb of a large south closing recumbent anticline. Above this nappe lies the highest tectonic unit in the district, the Upper Jotun Nappe composed of granulite facies rocks. (Hossack 1965; McRitchie, 1965).

Cross-folding (folding on axes not parallel to the north or north-east trend of the Caledonian mountain belt) on east-west axes has been recognized as post-nappe emplacement, but detailed analysis of fold systems (Bygdin excepted, Hossack, 1965) has not been undertaken. (Holtedahl, 1960).

At Bygdin, a well exposed deformed conglomerate has attracted the attention of geologists for many years. It was first investigated by Goldschmidt (1916a) who recognized that the/
the conglomerate was deformed, due, he suggested, to emplace-
ment of the Upper Jotun Nappe. Strand (1945) studied petra-
fabrics of deformed pebbles. Flinn (1961) attempted to ex-
plain the variable shape of deformed pebbles. The pattern of
cake-like and rod-like pebbles was thought to be related to
initial irregularities in the base of the Upper Jotun Nappe.
Hossack (1965), however, has shown that these irregularities
in the thrust plane are due to "second" folds formed in a post-
pebble deformation event. Hossack attributed pebble deformation
to a flattening phase following emplacement of the Upper Jotun
Nappe. The distribution of cakes and rods is difficult to
explain on Hossack's interpretation.

Hossack (1965) also tried to trace Bygdin structures
down into the Cambro-Ordovician rocks and Pre-cambrian basement.
He concluded that the contact between Valdres Sparagmite and
underlying Mellsenn was a sedimentary one and postulated that
first structures in the Cambro-Ordovician were the same age as
first structures at Bygdin. The Valdres Sparagmite and Mellsenn
formations are inverted and thrust over Cambro-Ordovician rocks
(Nichelsen, 1967) and a thrust must exist below the Mellsenn
formation at Bygdin. The correlation of first structures made
by Hossack (1965) must be reconsidered in the light of Nichelsen's
hypothesis.

Sjodalene. Very little work has been done in Sjodalene.
Dietrichson (1957) recognized the following tectonic units.
1. Upper/
i. Upper Jotun Erruptive Nappe: syenite, monzonite, mangerite, jotun norite, peridotite, greenstones.

ii. Valdøes Sparagmite.

iii. A nappe composed of labradorite-fels and norite.


The same tectonic elements are recognized by Strand and Holmsen (1960).

The Sel-Vågå Area. The present outcrop pattern of tectonic units is determined by post-thrust folds. The Trondheim Synclinorium is a structure trending north-east to north. The thrust at the base of the eugeosynclinal rocks of the synclinorium (tentatively correlated with the Otta thrust at the base of the eugeosynclinal succession in the Sel-Vågå area, Strand, 1961) and the Otta thrust, rest with marked tectonic unconformity on the miogeosynclinal Cambro-Ordovician ? sediments, e.g. the nappe of eugeosynclinal rocks lies on different parts of the miogeosynclinal succession (Holtedahl, 1960).

Other major post-thrust folds include the Lomskollen antiform extending from Lomskollen to Vágåmo and the Faltungsgraben system to the south of this fold. (Strand, 1951a).

Strand (1951a, 1964) differentiated between (in the Finna river west of Vágåmo) first recumbent folds in the Otta nappe, with axial planes parallel to the major thrust plane at the base of this nappe, and later tight east-west trending folds of Lomskollen antiform age.

Sotaseter/
Sotasetter, Hestbreipiggan-Nettoseter-Leirdalen Areas.

Detailed work has been done in these areas by workers from Nottingham University (Shouls (1958) - Sotasetter; Banham (1962), Elliott (1965), Banham and Elliott (1965) at Hestbreipiggan-Hoydalen-Bovertun; Cowan (1966), and Elliott and Cowan, (1966) at Nettoseter-Leirdalen).

Shouls (1958) described the geology around Sotasetter. He distinguished an infrastructure of Caledonized Pre-cambrian rocks from a superstructure of thrust masses of Eo-cambrian and Cambro-Ordovician sediments, and Lower and Upper Jotun Nappes which have moved eastwards off the Basal Gneiss Area (infrastructure). There are thrusts at the base of the meta-sediments and at the base of the anorthosites of the Lower and Upper Jotun Nappes lying above the meta-sediments. A north-west, south-east stretching lineation is considered to be parallel to the direction of movement along the thrusts. Large recumbent folds, north-south plunging, and with gently east dipping axial planes, are related to thrusting.

Banham (1962) investigated basal gneisses south of Hestbreipiggan, and Elliott (1965), meta-sediments between Hestbreipiggan and Hoydalen and Bovertun. Important aspects of this work are described in Banham and Elliott (1965). The contact between Eo-cambrian meta-sediments and Pre-cambrian basal gneisses, interpreted as a thrust, strikes at $45^\circ$ to the gneissosity (respectively north-east and east) and cuts across/
across a large igneous granite body concordant with the gneissosity. Structures parallel to the thrust (and first foliation in the meta-sediments) extend only a short distance into the gneisses; in this zone retrogression and mechanical breakdown of the gneisses has occurred. The metamorphic grade in the gneisses on Hestbreppigan (Banham, 1962) is green-schist, the same as in the meta-sediments, but there is ample evidence of retrogression in the former, probably from the almandine amphibolite facies. There is no evidence for metamorphism higher in grade than greenschist in the meta-sediments and it is concluded that meta-sediments were deposited or emplaced after the almandine-amphibolite metamorphism affecting the basal gneisses and before the greenschist metamorphism affecting sediments and basal gneisses.

Cowan (1966) who investigated a section from the Upper Jotun Nappe to the basal gneisses, has confirmed Banham and Elliott's (1965) views on the thrust nature of the basal gneiss, Eo-cambrian meta-sediment contact. There are also thrusts separating miogeosynclinal from eugeosynclinal rocks and at the base of the Upper Jotun Nappe.

An interesting conclusion reached by Cowan concerns the age of the Upper Jotun Nappe. At the base of this nappe is a mylonite zone which contains very few "second" minor folds. Below the thrust, quartzites are tightly folded by these folds and the thrust bringing in the Upper Jotun Nappe is parallel to their/
their axial planes. Cowan concludes that the Upper Jotun Nappe was emplaced during the second period of deformation.

Although metamorphic history in Cowan's (1966) area is difficult to work out because of the dominant psammitic lithology, the following is suggested:

**Table 11.**

<table>
<thead>
<tr>
<th>$F_1$</th>
<th>$M_1$</th>
<th>$M_2$</th>
<th>$M_3$</th>
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<tr>
<td></td>
<td>Syn-tectonic crystallization of muscovite, chlorite in $S_1$.</td>
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<tr>
<td>$F_2$</td>
<td></td>
<td>Syn and post $S_2$, crystallization of garnet and amphibole</td>
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<tr>
<td>$F_3$</td>
<td></td>
<td></td>
<td>Crystallization of muscovite in $S_3$.</td>
</tr>
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</table>

**The Basal Gneiss Area:** The contact between Eocambrian rocks and basement to the south of the "geanticlinal region" of Skjeseth (1963) is a discordant one. The "geanticlinal region" extends from Vigelen at the Swedish border south-west through Rondane to northern Valdres and is defined by windows in the Eocambrian and Cambro-Ordovician rocks exposing Pre-cambrian basement. It is either an unconformity as at Mjøsa, a thrust and/
and unconformity as at Grønsennknipa (Hossack, personal communication) or a thrust as at Bygdin (Hossack, 1965). North of the geanticline, however, a conformable contact between Eocambrian and basement can be traced from Lomskollen in the west of the Sel-Vågå area (Strand, 1951a) to Vågåmo and from there, north along the western border of the Trondheim synclinorium.

This concordancy has implications in the origin of the rocks in the Basal Gneiss Area and the following ideas have been suggested:

i. The area is composed entirely of Pre-cambrian rocks.

a. These were undeformed and sub-horizontal prior to deposition of Eocambrian and Cambro-Silurian sediments. Concordancy between Pre-cambrian and Eocambrian is primary (Oftedahl, 1963).

b. Pre-cambrian rocks have been deformed, metamorphosed, migmatized and intruded before deposition of Eocambrian and Cambro-Silurian rocks. Concordancy between Pre-cambrian and Eocambrian is due to structural and metamorphic processes. (Geological Map of Norway, 1915; Rosenquist, 1944, Strand, 1949, Birkland, 1958; Muret, 1960; Holmsen, 1964; Banham and Elliott, 1965; and Cowan, 1966). (Migmatite is used in Norwegian literature as defined in Turner and Verhoogen, 1960, p.359)

ii. The/
The Basal Gneiss Area consists of migmatized, metamorphosed and granitized Pre-cambrian, Eo-cambrian and Cambro-Silurian rocks. Concordant contacts of these rocks with rocks of the Trondheim Synclinoorium and miogeosynclinal rocks north of the Jotunheim mountains are caused by migmatite fronts passing up into these rocks, structural and metamorphic processes. (Holtebdahl, 1936, 1938b, 1944; Kolderup and Kolderup, 1940).

Rocks of the Basal Gneiss Area consist of metamorphosed, migmatized and granitized Eo-cambrian and Cambro-Silurian sediments. Concordant contacts are migmatite fronts. (Holtebdahl, 1938a; Barth, 1938; Richter, 1945; Ramberg, 1944; Strand and Holtebdahl, 1960).

Strong arguments favour the acceptance of ib above. Strand (1949) has pointed out that: (a) gneisses have a sharp boundary with overlying sparagmites, (b) there is no discordance at the contact as the upper parts of the gneisses are mylonites of Caledonian age, (c) on the south-east side of the Jotunheim, Eo-cambrian sparagmites and overlying sediments resting discordantly on the gneisses are similar to sparagmites and overlying sediments immediately above the contact at the north side of the Jotunheim, and (d) there is a strong similarity in gneisses immediately below sparagmites north and south of the Jotunheim mountains. If gneisses to the south are/
are considered Pre-cambrian, those occupying this position to the north must also be that age.

Where the contact has been investigated in detail, (Holmsen, 1961; Banham and Elliott, 1965; Cowan, 1966; Author's work at Lom) it has been shown to be a thrust.

Thirdly, the highest grade attained at the "concordant" contact during Caledonian metamorphism was greenschist facies (Holtedahl, 1960, map, p. 203). Migmatites do not form under these low grade conditions (Turner and Verhoogen, 1960, p. 370-375; Winkler, 1965, p. 176,199,203; Lundgren, 1966) and it is impossible for Pre-cambrian? and Cambro-Silurian sediments along the northern border of the Jotunheim mountains or western border of the Trondheim synclinorium to have been converted to migmatites. The concordant contact cannot have this origin.

The Upper Jotun Nappe: The Upper Jotun Nappe is the highest tectonic unit in the southern Norwegian Caledonides. It is an elongate body 180 km. long in a north-east direction by 61 km. at its east end to a minimum of 22 km. in a north-west direction from Lake Tyin Area is approximately 8600 - 8700 km². Everywhere, this nappe is in thrust contact with underlying rocks. (Strand, 1951a,b, Holtedahl, 1960).

Goldschmidt (1912) introduced the term Faltungsgraben for the large synform in which the Upper Jotun Nappe is found. Goldschmidt (1916b) recognized the unity of the Upper Jotun Nappe/
Nappe rock suite but over-emphasized the amount of igneous differentiation which had occurred. He thought all rock types present in the nappe were developed from jotun norite (K-feldspar, quartz, two pyroxenes, biotite) and in his Bergen-Jotun Kindred included pyroxenite, anorthosite, microperthite rocks, intermediate jotun norite and mangerite. Goldschmidt suggested that the Upper Jotun Nappe originated from within the Faltungsgraben.

Hødal (1945) revised rock nomenclature as, while Goldschmidt's terms suggested igneous rocks, they actually are metamorphic. He suggested the term jotunite for rocks containing K-feldspar as an essential phase in addition to plagioclase, two pyroxenes, biotite, ± quartz, K-feldspar free rocks are termed two pyroxene granulites.

Batley (1960) who studied the Upper Jotun Nappe in the Visdalen region, found that ultrabasics in the granulite facies country-rocks were not alpine peridotites, Bushveld types, or the products of differentiation as found at the base of large sills, but are similar to peridotites and pyroxenites found in the Lewisian of Scotland. The ultramafic bodies are sheets generally concordant with foliation in the gneisses, and there are no lateral connections nor feeding dykes between sills. Mineralogy is olivine, ortho and clino pyroxenes, and bodies are poorly to well foliated. There is evidence of crushing and shearing within the bodies. Ultrabasics are retrograded at/
at their contacts with the surrounding gneisses and although they are generally concordant with country-rock foliation, small-scale discordances can be seen.

Battey (1960) postulated injection of the ultramafics along the foliation largely in a solid state with cataclasis, but when under P.T. conditions permitting mobility, mineral readjustment at boundaries and some modification of adjacent gneisses. The intrusion could represent re-injection under orogenic conditions of material formerly separated by crystallization differentiation from jotun norite magma.

Battey (1965) described work done north-west of Lake Bygdin where lineated and foliated jotunite and two pyroxene granulites contain textures indicating recrystallization under granulite facies metamorphic conditions. Locally, rocks of gabbroic or norite mineralogy with well preserved igneous textures are seen; in the field they are always closely related to the foliated rocks. Battey, agreeing with Goldschmidt that gneissic jotunites originally had an igneous history, rejects his conclusion of solid state intrusion of ultrabasics (Battey, 1960) and believes that they and the rest of the gneisses are products of igneous sedimentation. This view is held by Dietrichson (1958) who believes the Upper Jotun Nappe was originally a layered intrusion.

McRitchie (1965) described in detail the area near Eidsbugarden. He has mapped the east-north-east striking high/
high angle, north dipping, reverse Tyin-Gjende fault. The rocks on either side of the fault are different. On the north are layered pyroxene rich granulites in more coarse grained jotunite. Here, McRitchie postulates reconstitution of igneous rocks at granulite facies conditions with production of layering by granulitization and metamorphic differentiation. To the south of the fault distinct thrust sheets can be recognized in the upper-most being a gabbro in which well developed layering is present.

Formation of the initial igneous rocks both north and south of the fault and subsequent metamorphism in the granulite facies are pre-B1, of Hossack (1965) e.g., pre-thrusting. The Tyin Gjende fault is correlated with the third movement phase of Hossack. (McRitchie, 1965).

Smithson's (1964) conclusions based on a gravity survey of the north-eastern half of the Upper Jotun Nappe are dependent on assumed density contrasts between the nappe and basement. Calculations made with a value of 0.12 gm/cc (from mean values of 2.86 gm/cc for the Upper Jotun Nappe and 2.74 gm/cc for the basement) give a thickness of 1.2 km for the nappe and suggest a root zone within the Faltungraben. If a value of 0.20 gm/cc is used in calculations, a thickness of 6.2 km which suggests to Smithson, the nappe is a far-travelled thrust sheet.
DISCUSSION AND CONCLUSIONS.

Discussion of structural evolution and conclusions which can be made are not presented at this point. This aspect of the thesis will be dealt with more effectively following chapters devoted to the Lom and Sjodalen areas which I have studied in detail.
INTRODUCTION.

Lom, a village of 400 people 320 km. north of Oslo, is situated in Ottadalen at the junction of the Otta river and its tributary, the Bøvra. These rivers flow in U-shaped glaciated valleys which have been cut deeply into pre-existing more subdued topography. Maximum relief in the area mapped is 1550 m. and the river valleys are 600 - 800 m. deep.

Valley sides of Ottadalen are steep, rising on the north to rounded topography developed on gneisses. To the south, the valley wall rises steeply to 1000 - 1200 m. The ground rises more gently out of the valley culminating in the high peaks of the Jotunheim mountains. Spectacular topography has been developed during Pleistocene glaciation.

The Lom area of 64 km² lies to the south of the Otta river extending 9.6 km south from, and 14.5 km east of Lom. Outcrop is accessible in the main valley except where slopes are too steep to be negotiated. In the south there is only one access road necessitating much walking and climbing to reach/
<table>
<thead>
<tr>
<th>Stratigraphy of tectonic units.</th>
<th>Thickness.</th>
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<tbody>
<tr>
<td>Upper Jotun Nappe.</td>
<td>initially 15 km?</td>
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<tr>
<td><strong>Upper Jotun Nappe Thrust Plane.</strong></td>
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<td>Salellseter Nappes.</td>
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<tr>
<td>Hovilelangtjern group</td>
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<tr>
<td>thrust</td>
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<tr>
<td>Vindsjokamp - Kyreggi group.</td>
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<td>Skårdalen thrust.</td>
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<td>Ottadalen Nappe</td>
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<td>Ausfjell Striped group</td>
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<tr>
<td>Lom-Lomskollan Psammites = Vulu-Glomsteinhøe Psammites.</td>
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<tr>
<td>Soleggje Rivillan group</td>
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<tr>
<td>Ottadalen Thrust.</td>
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<tr>
<td>Basal Gneisses.</td>
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</table>
reach exposures.

Outcrop is adequate. It is better on the higher ground and in river sections with variable exposure on slopes. The field season on ground above 1000 m is limited by snow to mid June - mid September. In valleys the season is extended to early May - mid October. The latter half of the summer is wet especially on higher ground.

DESCRIPTION OF TECTONIC UNITS.

Introduction.

At Lom a section extending from the Basal Gneiss Area to the Upper Jotun Nappe is exposed. The mainly south dipping tectonic units that make up this section have been brought together by thrusting. The tectonic units, considered as map units (see map 1, pocket), and their thicknesses are presented in Table III.

Before describing the tectonic units a brief outline (in greater detail below) of the structural geometry and relationships of units to each other is necessary.

Miogeosynclinal rocks form a large syncline with an inverted southern limb, the Ottadalen Nappe. The uninvetered limb lies on basal gneiss. In this fold sedimentary structures/
structures indicate the stratigraphy presented in Table III. The Soleggje-Rivillan group and Vulu-Glomsteinhoe psammites are inverted.

The Vindsjokamp-Kyreggi group bounded at top and bottom by phyllonites, lies on rocks stratigraphically higher in the Soleggje-Rivillan group at Soleggje than at Rivillan. The base of the Vindsjokamp-Kyreggi group is a thrust. A thrust separates this group from the overlying Hovilelangtjern group. The Upper Jotun Nappe has been thrust above the Hovilelangtjern group.

Tectonic Units.

Basal Gneisses. Well banded gneisses in which quartzo-feldspathic bands alternate with bands rich in mafic minerals including biotite, hornblende and garnet. Biotite concentrations at the edges of mafic bands are common. Mafic and quartzo-feldspathic bands vary in thickness from 5 mm to 10 cm. Small lenses (30 cm by 1.5 m to 1 m by 8 m) and bands of amphibolite (30 cm to 3.5 m thick) are common.

Gneisses almost imperceptibly become more schistose and micaceous towards the Ottadalen thrust. Locally, gneisses up to 200 m below the contact are highly schistose with strong/
strong developments of porphyroblasts of muscovite lying in a foliation parallel to the thrust plane.

In thin section the effects of more than one metamorphism affecting the gneisses are apparent. Observations supporting this conclusion are summarised below.

1. Plagioclase feldspar is of variable composition from An35 to An3 (Michel-Levy method).
   a. All plagioclase except some albite is more or less altered containing numerous minute inclusions of epidote (used in the sense of Miyashiro and Seki (1958) to include solid solution from clino-zoisite to Fe-epidote) and sericite. The less calcic a plagioclase the more altered it is.
   b. Clear unaltered albite has been seen in most thin sections.

ii. Epidote in more mafic gneisses (biotite rich rocks, amphibolites) is often well zoned. Epidotes in quartz-feldspathic gneisses are usually unzoned. Outer zones of zoned epidotes invariably have higher birefringence than inner zones and higher Fe content (Deer et al. 1962, VI, p. 202).

iii. K-feldspar locally has been partially altered to muscovite.

iv. Biotite replaces common green hornblende at grain boundaries and along cleavages. Hornblende is also replaced by chlorite and locally is surrounded by rims of a pale green amphibole/
amphibole with lower birefringence than, and in optical continuity with hornblende. It probably is tremolite-actinolite.

v. Garnet is altered at grain boundaries and along fractures to biotite and chlorite.

vi. Locally biotite is altered to chlorite.

(See plates 1A,B-F).

Retrogression of a more calcic to less calcic plagioclase is indicated by inclusions of epidote, relatively richer in CaO, in the altered plagioclase. The presence of unaltered albite, the stable plagioclase at the P,T. conditions during which retrogression occurred, indicates that most plagioclases have not reached equilibrium during the retrograde metamorphism.

Miyashiro and Seki (1958) suggest that epidote composition is temperature dependent and reflects bulk composition. In a sequence of metamorphic rocks from green phyllites to amphibolites (sodic oligoclase, epidote and common hornblende) it was observed that higher grade epidotes are less Fe-rich than lower grade ones. During retrograde metamorphism of amphibolites, Fe-rich outer zones were formed on prograde epidotes. A similar origin for zoned epidotes in gneisses in the Lom area, often associated with hornblende and An30 plagioclase assemblages, is proposed. In quartzofeldspathic gneisses unzoned epidotes were formed during the retrograde metamorphism as, probably, these rocks were epidote free prior to/
to retrogression.

Prograde metamorphism was in the staurolite-almandine or kyanite-almandine-muscovite sub-facies of the almandine amphibolite facies (Winkler, 1965 p. 88-91, especially p. 90-91). Critical phases, staurolite and kyanite have not been observed and no choice between the two sub-facies can be made.

During retrograde metamorphism at middle greenschist conditions, Fe-richer epidote, biotite, tremolite-actinolite, chlorite albite and muscovite were stable phases. Retrograde effects extend throughout all gneisses investigated in the Lom area and do not decrease in intensity structurally downwards. Prograde metamorphism in the meta-sediments is middle greenschist in grade and it is postulated that it was synchronous with retrograde metamorphism in the gneisses. The almandine amphibolite metamorphism was pre-emplacement of the meta-sediments. (Cf. Banham, 1962; Banham and Elliott, 1965; Cowan, 1966).

Ottadalen Nappe: In description of meta-sediments in this nappe, the rest of the rocks at Lom, and those in Sjodalen, the following classification of meta-sediments has been used. The classification of Brown (1964) was modified by the author to make it more suitable to lithologies in the Lom and Sjodalen areas.
areas was used.

A. Psammitite: Poorly foliated, homogenous rock with less than 10% micas. Sedimentary structures common. Depending on mineralogy, divisible into -

i. Quartzite. 90% quartz.

ii. Feldspathic quartzite. 90 – 75% quartz.

iii. Arkose. 75% quartz.

B. Semi-pelite. Homogenous generally well foliated rock intermediate between psammitite and pelite in composition.

C. Pelite. Homogenous very well foliated rocks usually dark brown and weathering rusty brown. Mica content greater than 50% (phyllites and schists of Winkler 1965, p.209 - 213).

D. Striped and banded rocks: Heterogenous rock units where pelite, semi-pelite and psammitite are inter-banded in layers generally less than 0.5 m thick. Relative amounts of pelite, semi-pelite and psammitite are variable.

Three distinct groups of rocks make up the Ottadalen nappe. In stratigraphic order from highest to lowest they are (See Table 111):

i. Ausfjell Striped Group.

a. Semi-pelite unit: Mainly well foliated quartz-feldspar-mica (< 50%) schist with minor amounts of pelite and psammitite (feldspathic quartzite and arkose). Rocks are dark grey and grey brown weathering.

b. /
b. Striped unit: Interbanded pelite (dark grey, carbonaceous, rusty-brown weathering, segregation quartz (as described by McNamara (1965), p. 372-375) is common between mica foliae), psammite (quartzite and feldspathic quartzite) and minor orange marble near the base of the unit. Pelite and psammite bands, 30 cm. to 3.5 m thick at the top of the unit increase in thickness downwards to 10 - 20 m. at the base. Contacts between adjacent bands are sharp; bottom structures are not present. The contact with the Lom-Lomskollen psammites is gradational with an increase in the absolute amount of psammite.

ii. Lom-Lomskollen Psammites.

Feldspathic quartzite with minor quartzite (light grey to white, weathering light to medium grey and rusty-brown) in thrust contact with underlying gneisses. These rocks present the same appearance and approximate thickness over a strike distance of 11 km. from Lom to the south side of Lomskollen. Thin pelite partings become more frequent towards the Ausfjell Striped Group. At the junction with the Ausfjell Striped Group there is a rusty orange weathering band of micaeous marble and carbonate rich semi-pelite.

Current-bedding is common at the base of the Lom-Lomskollen psammites adjacent to basal gneisses, indicating the rocks are uninverted. Bedding is defined throughout by bands/
FIG. 3: Younging Evidence.

A. Lom-Lomskollan Psammites. Pebbles occasionally present.

B. Vulu-Glomsteinhöe Psammites. Cross-bedding defined by truncated beds rich in heavy minerals.

C. Vulu-Glomsteinhöe Psammites. Truncated micaceous partings. 1.5 km. north-east of Rivillan.

D. Ripple-current bedding in Soleggje-Rivillan group 1 km. north-west of Soleggje.
bands rich in detrital feldspars and in heavy minerals.

In thin section feldspar porphyroclasts (plagioclase, orthoclase, perthite, microcline, size range 2 mm. by 0.5 mm to 0.2 by 0.1 mm.) are set in a finer grained matrix of quartz and albite (0.2 by 0.8 mm. to 0.04 by 0.03 mm). There is accessory muscovite, biotite, chlorite and epidote. Metamorphic grade is middle greenschist (Turner and Verhoogen, 1960, p.537).

iii. Vulu-Glomsteinhöe Psammites:

Feldspathic quartzites and arkoses, light in colour and buff. Bedding is well defined by thin (13 to 2 mm.) layers rich in heavy minerals and rich in detrital feldspars. Current bedding, common throughout the outcrop of these rocks, everywhere indicates they are inverted (fig. 3). The contact with structurally higher, stratigraphically lower rocks, the Sollegje-Rivillan Group, is gradational, through a 30 m. thickness due to increasing pelite content. Contact with the Ausfjell Striped Group is nowhere seen. However, on the slope rising from the Vulu river to Glomsteinhöe and Rivillan at 1000 m. elevation, there are outcrops of orange weathering carbonate rich micaeous psammites similar to those seen at the top of the Lom-Lomskallen psammites. These underlie structurally, (and are stratigraphically above), the Vulu Glomsteinhöe psammites. Similar observations have been made at the east end of Glomsteinhöe.

In thin section, rocks are characterised by porphyroclasts of feldspar (plagioclase, orthoclase, perthite, microcline)/
microcline) 3.2 by 2 mm. to 0.15 by 0.1 mm. Porphyroclasts are set in a finer grained matrix of quartz and albite of variable size ranging from 0.3 by 0.1 to 0.08 by 0.03 mm., and accessory muscovite, biotite, chlorite and epidote. Grade is middle green schist.

The Vulu-Glomsteinhøe psammites are correlated with the Lom-Lomskollen psammites as:
a. Similar lithology - mainly feldspathic quartzites and arkoses.
b. Similar mineralogical make up.
c. The Lom-Lomskollen psammites are un inverted. At their top is a distinctive orange weathering micaceous marble and calcareous psammitite. The same rock is found at the stratigraphic top of the inverted Vulu-Glomsteinhøe psammites.
d. Both psammitite groups young into the Ausfjell Striped Group.

There is no continuity of outcrop between psammitite groups.

iv. Soleggje Rivillan Group.

This group stratigraphically below and structurally above the Vulu-Glomsteinhøe psammites, is bounded to the south by a thrust plane. The group is well exposed north of Soleggje in the west of the Lom area and at Rivillan 8 km. to the east, but intermediate ground is only poorly exposed. Units within this group include:
a./
### TABLE IV.

**DETAILED STRATIGRAPHY**

of

**PART OF THE SOLEGGJE-RIVILLAN GROUP.**

<table>
<thead>
<tr>
<th>Description of rock unit.</th>
<th>Thickness.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Skardalen thrust</td>
<td></td>
</tr>
<tr>
<td>light grey and buff limestone weathering orange</td>
<td></td>
</tr>
<tr>
<td>Quartzite and feldspathic quartzite with occasional thin pelite interbands.</td>
<td>160m.</td>
</tr>
<tr>
<td>Platy quartzite, rusty weathering.</td>
<td></td>
</tr>
<tr>
<td>Dark grey pelite.</td>
<td></td>
</tr>
<tr>
<td>Micaceous feldspathic quartzite.</td>
<td></td>
</tr>
<tr>
<td>Light grey rusty weathering quartzite.</td>
<td></td>
</tr>
<tr>
<td>Platy quartzite, feldspathic quartzite.</td>
<td></td>
</tr>
<tr>
<td>Medium grey feldspathic quartzite.</td>
<td></td>
</tr>
<tr>
<td>Dark grey pelite.</td>
<td></td>
</tr>
<tr>
<td>Light grey-brown platy quartzite.</td>
<td>19m.</td>
</tr>
</tbody>
</table>
a. Dark grey rusty weathering pelite containing a few psammite bands (quartzite, feldspathic-quartzite). Segregation quartz is common in the pelites. Except for bedding defined by compositional layering, sedimentary structures are absent.

b. Interbedded arkoses, feldspathic quartzites and quartzites. This unit is not well exposed and contacts with adjacent units have not been seen. Arkoses are light to medium grey, quartzites are platy and weather light rusty brown.

c. Dark grey, dark rusty brown weathering pelite. Locally segregation quartz is common. Contact with d. is gradational over 3.5 to 5 m.

d. Striped unit. Interbedded psammite (light to medium grey, weathering medium grey to rusty brown) and pelite (dark grey, rusty weathering with segregation quartz). Quartzite forms the bulk of the psammite; there is only minor feldspathic quartzite. A detailed section, measured in the Vulu river immediately north of Sæløsseter, is presented to give the reader an appreciation of the types and variation of lithologies in this unit. (See Table iv).

Thicker rock members of this unit, such as the 150 m. band of quartzite and feldspathic quartzite, can be traced along strike for as much as 6.5 km. but lack of sufficient outcrop renders it impossible to show continuity along the strike of thinner units.

Sedimentary structures, mainly bedding, are common in
the thicker quartzites and feldspathic quartzites. Locally bedding truncations are suggested. Immediately north of the summit of Soleggje, good ripple current bedding (Fig. 3) indicates ricks are inverted.

At Rivillan, structurally above the Vulu-Glomsteinhoe psammites is a striped series of pelites and lesser amounts of quartzites. Quartzites are light yellow brown to light grey and weather light rusty brown. Pelites are medium to dark grey, and dark rusty brown weathering. These rocks are correlated with d. above. The thrust, which defines the upper boundary of the Soléggje-Rivillan group, lies on higher stratigraphic units of the group at Rivillan than at Soleggje.

Salelseter Nappes.

i. Vindsjokamp-Kyreggi Group:

Phyllonites are found at the base and top of this unit. Thrust contact with the Soleggje-Rivillon group has been shown above and the upper contact with the Hovilelangtjern group is probably a thrust.

Variable appearance in the field characterises this group. Immediately above and below phyllonites, are dark quartz-feldspar-mica schists. In the middle of the group, rocks are feldspar-rich psammites in which foliation is sometimes only poorly developed. Locally, compositional banding is/
is well defined by alternating amphibole and quartz-feldspar rich layers (band 2.5 to 8 cm). Locally an augen gneiss (especially east and south-east of Rivillan) facies is developed with lensoid feldspars elongated in the regional foliation.

In thin section the rocks show variable composition. Feldspathic quartzites and arkoses are common, (with feldspar porphyroclases of microcline, orthoclase, perthite and plagioclase). Amphibole rich bands have been mentioned above. The amphibole is dark green hornblende rimmed by pale green to clear amphibole (Plate 3A). The clear amphibole is in optical continuity with the hornblende and as it is less pleochroic, and has a lower R.I., and lower birefringence than the hornblende it is tremolite-actinolite. Other minerals present in these rocks include muscovite, albite, biotite and epidote. No garnet has been seen. These rocks are probably meta-sediments locally rich in hornblende (Graywackes with 10.5% hornblende have been reported, see Pettijohn (1957), p.304, Table 50, No.E). During middle greenschist metamorphism affecting these rocks, hornblende being unstable, was partially retrograded to tremolite-actinolite, stable at these metamorphic conditions. (Turner and Verhoogen, 1960, p.537). Tremolite rimed hornblende from lower and middle greenschist facies sediments in the Grampian Highlands (in Green Beds and other rocks) have been reported by Tilley (1937). The author proposes the same origin for rimmed/
rimmed amphiboles in the Lom area as Tilley (1937).

The relatively high mica content associated with quartz and feldspar found in these rocks suggests that they have been formed from clay-rich arkoses or graywackes. The presence of detritial amphibole further suggests an immature sediment and it is reasonable to correlate this group of rocks with eugeosynclinal rocks, (Strand 1951a) has mapped 4 km. east of the Lom area.

ii. Hovilelangtjern Group.

Hovilelangtjern Group: This group, found only in the south-western part of the Lom area, lies between phyllonites at the top of Vindsjokamp Kyreggi group and phyllonites at the base of the overlying Upper Jotun Nappe. In the east of the Lom area (at Kyreggi) phyllonites at the base of the Upper Jotun Nappe rest directly on rocks of the Vindsjokamp-Kyreggi group.

The Hovilelangtjern group is divisible into two units. The lower unit is of light brown, brown weathering semi-pelite, feldspathic-quartzite and quartzite alternating in bands 30 cm. to 3-5 m. thick to form a striped series. The upper unit is composed of distinctive greenish-grey feldspathic quartzites and arkoses and its transition to the lower unit is sedimentary. Banding is well developed in the upper unit and is defined by layers (usually 1.5 cm. to 5 mm.) rich in heavy minerals, mica and detrital feldspars. Porphyroclasts of feldspar, some purplish suggesting garnet, are easily seen in the field.
FIG. 4: MODAL ANALYSES.

<table>
<thead>
<tr>
<th>Phases</th>
<th>Lom-Lomskollan Psammites</th>
<th>Vulu-Glomsteinhoe Psammites</th>
<th>Upper unit Hovilelangtjern group</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz</td>
<td>82.7</td>
<td>83.1</td>
<td>73.4</td>
</tr>
<tr>
<td>K-feldspar</td>
<td>7.6</td>
<td>6.2</td>
<td>12.4</td>
</tr>
<tr>
<td>Plagioclase</td>
<td>-</td>
<td>4.1</td>
<td>3.0</td>
</tr>
<tr>
<td>Muscovite</td>
<td>5.2</td>
<td>2.3</td>
<td>4.9</td>
</tr>
<tr>
<td>Biotite</td>
<td>1.3</td>
<td>2.1</td>
<td>6.2</td>
</tr>
<tr>
<td>Epidote</td>
<td>0.2</td>
<td>P</td>
<td>0.1</td>
</tr>
<tr>
<td>Chlorite</td>
<td>1.2</td>
<td>0.7</td>
<td>P</td>
</tr>
<tr>
<td>Sphene</td>
<td>P</td>
<td>0.9</td>
<td>P</td>
</tr>
<tr>
<td>Ore</td>
<td>1.1</td>
<td>1.3</td>
<td>P</td>
</tr>
<tr>
<td>TOTAL</td>
<td>99.3%</td>
<td>100.7%</td>
<td>100.0%</td>
</tr>
</tbody>
</table>

P: present.
2000 points in each case.
In thin section rocks of the upper unit are distinctive. Porphyroclasts of plagioclase, microcline orthoclase and perthite (3.5 by 2.5 mm. to 0.3 by 0.1 mm.) are set in finer grained well recrystallized matrix (0.2 by 0.1 to 0.08 by 0.03 mm.) of quartz, albite, light green slightly pleochroic muscovite, epidote, chlorite and minor biotite (Plates 4B,C). Epidote content of these rocks is distinctive being higher than in any other psammites in the Lom area (see model analyses, Fig. 4.).

Metamorphic grade is middle greenschist (Turner and Verhoogen, 1960, p.537)

The Upper Jotun Nappe: A study of the Upper Jotun Nappe is a study in itself and has not been attempted in this thesis. At the base of the nappe phyllonites have been produced by granulitization and metamorphic retrogression occurring during and after emplacement of the Upper Jotun Nappe. The phyllonites have been studied because they contain Caledonian structures and have suffered Caledonian metamorphism.

Cataclastic rocks at Lom have been termed phyllonites as they are recrystallized and too coarse grained to be termed mylonites in the sense of Lapworth (1885) who defined mylonite as "a microscopic pressure - breccia with fluxion-structure in which the interstitial dusty siliceous and kaolinitic paste has only recrystallized in part". In the field the rocks, recrystallized, are phyllites or fine grained schists, produced by the degradation of more coarse grained rocks, for which the name/
FIG. 5.

A. Lenses of epidosite in phyllonite. As epidote was not a stable phase in granulite facies gneisses from which phyllonites have been derived, these lenses are syn-
phyllonite layering.

B. Detail of the junction between a zone of phyllonite and un cataclased unbanded gneiss.

C. Section normal to phyllonite layering showing transition zone between phyllonites and gneisses (stippled).
FIG. 5.

A. PHYLLOMUTES AT THE BASE OF THE UPPER JOTUN NAPPE.

B.

C.
name phyllonite is apt (Christie, 1969).

Phyllonites are well layered (0.5 mm. to 1.5 cm.)
dark green and grey rocks. Individual layers can be traced for
a considerable length relative to their thickness but generally
pinch out so that all layers, in fact, are very acute lenses.
Lenses (2.5 cm. by 6 mm. to 15 cm. by 4 cm.) and bands (5 mm. to
5 cm.) of light coloured quartzofeldspathic material are common,
and occasionally lenses (15 cm. by 2 cm. to 70 cm. by 15 cm.)
of epidote are seen (Fig. 5). In all cases these lenses are
elongated in the phyllonite layering.

Phyllonites are of variable thickness ranging from
40 to 100 m. The passage from phyllonites to undeformed
granulites of the Upper Jotun Nappe is gradational over a
distance of 10.25 m. normal to phyllonite layering (Fig. 5).
Lenses of mechanically undeformed gneiss isolated in phyllonite
mark the beginning of this passage. Upwards, lenses become
larger and more common, and at the top of the transition zone
bands of phyllonite occupy only thin zones between undeformed
lenses. Above rocks are undeformed gneisses.

In thin section phyllonites are well recrystallized
equigranular (0.1 to 0.05 mm.) and inequigranular rocks (Plates 2C,
E, F). The latter contain porphyroclasts of feldspar and
porphyroblasts of epidote up to 3 by 2 mm. The rocks, granu-
litized, have also been retrograded and the present mineral
assemblages have developed at middle greenschist metamorphic
conditions. As epidote is not stable in the granulite facies
(Winkler, 1965, p. 74), epidote in phyllonites has been formed
during/
during retrograde metamorphism; locally epidote porphyroblasts developed. Conversely, most of the feldspar larger than equigranular matrix material are calcic plagioclases and K-feldspar stable in Upper Jotun Nappe granulite facies gneisses but not at middle greenschist conditions. The feldspars could not have been formed during retrograde metamorphism and must be porphyroclasts derived from the gneisses.

Minerals formed during retrogression include quartz, albite, epidote, muscovite, biotite, chlorite, minor carbonate, sphene, and microcline (middle greenschist facies, Turner and Verhoogen, 1960, p. 537). Breakdown of calcic plagioclase has produced epidote and albite with minor calcite. Ferromagnesian minerals have been retrograded to biotite, chlorite and tremolite-actinolite. Orthoclase and perthite have been partially retrograded to muscovite and quartz. Garnet, granulitized and partially retrograded to biotite and chlorite, has been seen.

Locally, well above the main phyllonites, marginal granulitization and alteration of granulite facies has occurred. The metamorphic grade indicated by alteration products is greenschist and this retrogression event could have been synchronous with retrogression in the main phyllonites.
<table>
<thead>
<tr>
<th>Movement Phase</th>
<th>Planar Structures</th>
<th>Linear Structures</th>
<th>Fold Structures</th>
<th>Occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pre-deformation</strong></td>
<td>S₀: Bedding, current bedding gneissic foliation in the Basal Gneisses and Upper Jotun Nappe.</td>
<td>L₀: intersection of bedding and current bedding</td>
<td>None preserved.</td>
<td></td>
</tr>
<tr>
<td><strong>First</strong></td>
<td>S₁: Penetrative muscovite-biotite foliation in which quartz grains have been flattened. S parallel to major thrusts.</td>
<td>B₅ⁿ</td>
<td>F₁: &quot;Similar&quot; folds, S₁ is axial planar.</td>
<td>Ottadalen Nappe, Saleilseter nappes, basal gneisses, phyllonites.</td>
</tr>
<tr>
<td><strong>Second</strong></td>
<td>S₂: Crenulation cleavage and biotite-muscovite foliation developed only in cores of F folds.</td>
<td>B₅ⁿ, B₅ⁿ</td>
<td>F₂: Flattened flexural folds.</td>
<td>Mainly the Ottadalen nappe.</td>
</tr>
<tr>
<td><strong>Third</strong></td>
<td>S₃: geometric axial planes of F folds.</td>
<td>B₅ⁿ, B₅ⁿ, L₃: Mullion structure and strong rodding parallel to B</td>
<td>F₃: Variable ranging from open concentric to tight folds thickened in the axial zone.</td>
<td>Ottadalen Nappe, Saleilseter Nappes?</td>
</tr>
<tr>
<td>Fourth.</td>
<td>$S_{4}:$ geometric axial plane of $F_{4}$ folds.</td>
<td>$L_{4}:$ Mica crinkling parallel to $B$.</td>
<td>$F_{4}:$ Chevron or kink folds.</td>
<td>Northern half of the outcrop of the Otta-dalen nappe.</td>
</tr>
<tr>
<td>---------</td>
<td>-----------------------------------------------</td>
<td>-----------------------</td>
<td>-----------------------</td>
<td>--------------------------------------------------</td>
</tr>
<tr>
<td>Fifth.</td>
<td>$S_{5A}, S_{5B}:$ Crenulation cleavage and muscovite foliation.</td>
<td>$L_{5}:$ Girdle axis defined by $S_{5A}, S_{5B}$. $L_{5A}: S_{s} / S_{1}$ and $S_{S_{5A}}$.</td>
<td>$F_{5A}:$ Minor folds of variable style ranging from &quot;similar&quot; to concentric.</td>
<td>Ottadalen Nappe, Salellseter Nappe.</td>
</tr>
</tbody>
</table>
STRUCTURAL GEOLOGY.

Introduction.

Five sets of minor structures (see Table IV) were recognised using interference relationships. Where interfering structures were not available, structures were tentatively assigned on age by comparison of their style with the style of structures of known age from other outcrops. Major structures are synchronous with the first and third sets of structures.

Foliation, axial plane cleavages, lineations and minor folds are associated with these five sets of structures. The first foliation generally strikes east-west and dips moderately to steeply south. In some sub-areas, though, S1 foliation planes lie on a girdle with an east plunging axis and reflect major folding of the third movement phase. First linear structures plunge 20° to 30° to the east in the north of the Lom area and are variable in trend in the south.

Second structures are minor folds of S1, and their axial plane cleavage, sub-parallel to S1, dips to the south except where folded by structures of the third movement phase.

Third structures include major east plunging folds with moderate to steep dipping axial planes, and minor east plunging/
folds.

There remain three sets of minor structures, one well developed in the north of the Lom area where it constitutes the fourth movement phase, while the other two are well developed in the south. The two sets in the south, not seen to interfere with each other, have been shown to post-date structures of the fourth movement phase mentioned above. It is suggested that the two sets of structures in the south are synchronous and constitute the fifth movement phase.

Vertically dipping east-north-east striking faults post-date the fifth movement phase.

Phyllonite layering at the base of the Upper Jotun Nappe breaks down gneissic banding in rocks of the nappe. The banding is pre-thrusting and considered Pre-cambrian in age.

Structural Movement Phases.

First Movement Phase. During the first movement phase, tectonic units (p. 31 - 45) were brought together by thrusting and the large field constituting the Ottadalen Nappe was formed. Minor structures of the first movement phase were generated after thrusting.

a. Description of thrusts.

"Younging evidence supports the existence of at least one major fold of this phase but is not abundant enough to rule out the possibility of more than one fold."
i. Ottadalen Thrusts. The author has postulated a major fold forming the Ottadalen Nappe (see page 49) and lateral equivalents of the Soleggje-Rivillan group should be found below the Lom-Lomskollen psammites. On the south side of Lomskollen, between basal gneisses and light grey feldspathic quartzites of the Lom-Lomskollen psammites, are 80-90 m. of dark grey pelite, and semi-pelite. The sedimentary passage to the Lom-Lomskollen psammites is similar to the passage between the Vulu-Glomsteinhoe psammites and the Soleggje-Rivillan group exposed immediately north of Rivillan, and in the Vulu river. The 80-90 m. of pelites and semi-pelites are tentatively correlated with the stratigraphically highest rocks of the Soleggje-Rivillan group. Elsewhere, along the north slope of Ausfjell, Lom-Lomskollen psammites lie directly on basal gneisses. This contact must be a thrust which has cut out the whole of the Soleggje-Rivillan group with the exception of 80 m. of that group exposed below Lom-Lomskollen psammites on Lomskollen.

Basal gneisses become intensely granulitized towards the contact with the Lom-Lomskollen psammites and within 20 m. of the contact have been retrograded to dark grey phyllonites. Phyllonites were formed during thrust emplacement of the Ottadalen Nappe.

ii. Skardalen Thrust. There are 10-20 m. of dark grey phyllonites at the base of the Vindsjokamp-Kyreggi group.
In Skardalen the Vindsjokamp-Kyreggi group rests on an orange limestone (unit d., p.39), the stratigraphically lowest unit of the Soleggje-Rivillan group. At Rivillan, 8 km. to the east, the same rocks rest on stratigraphically higher rocks of the Soleggje-Rivillan group (unit a., p.39).

There is an abrupt change in sedimentary facies at the contact between rocks of the Vindsjokamp-Kyreggi group and the Soleggje-Rivillan group, respectively, between eugeosynclinal and miogeosynclinal rocks (Strand, 1961, 1964a; Holtedahl, 1960). No transitional rocks between the Vindsjokamp-Kyreggi and the Soleggje-Rivillan groups have been seen and the contact between the groups is a thrust.

### iii. Hovilelangtjern Thrust

Phyllonites are found in the top of the Vindsjokamp-Kyreggi group at its contact with the overlying Hovilelangtjern group. At Kyreggi, the Vindsjokamp-Kyreggi group is in thrust contact with phyllonites at the base of the Upper Jotun Nappe, and the outcrop of the Hovilelangtjern group is wedge-shaped, pinching out to the east. This, and the presence of phyllonites at the top of the Vindsjokamp-Kyreggi group, indicate that contact between the Hovilelangtjern and Vindsjokamp-Kyreggi groups is a thrust.

### iv. The Upper Jotun Nappe Thrusts

A major thrust plane which separates middle greenschist meta-sediments from granulite facies metamorphic rocks. Phyllonites developed at the thrust have been described above (p.43).
FIG. 7: Location of Younging Evidence.

U. UNINVERTED ROCKS.
I. INVERTED ROCKS.

OBSERVATIONS FROM 26 LOCALITIES.
b. The Ottadalen Fold. There is evidence suggesting the existence of a major first fold, the Ottadalen Fold (which makes up the Ottadalen Nappe). (See cross section Fig. 6 pocket). Some of this evidence has been described on page 38 and is summarized here:

i. Lithological similarities between Lom-Lomskollen psammites and Vulu-Glomsteinhoe psammites suggesting correlations of these groups.

ii. All sedimentary structures (see fig. 3, 19 localities in the Vulu-Glomsteinhoe psammites, 5 localities in the Lom-Lomskollen psammites, 2 localities in the Solaggje-Rivillan group) seen in the Lom-Lomskollen psammites indicate it is un inverted; all seen in the Vulu-Glomsteinhoe psammites and Solaggje-Rivillan groups indicate inversion.

iii. The Lom-Lomskollen psammites and the Vulu-Glomsteinhoe psammites young into the Ausfjell striped group.

The? Ottadalen fold has been assigned to the first movement phase because:

i. By analogy with other areas in the southern Norwegian Caledonides, major folds producing regional inversion of stratigraphy are primary structures (Nichelsen, 1967).

ii. It is easiest to visualize contemporaneous formation of the? Ottadalen fold, and the Ottadalen thrust. The latter was formed during the first movement phase as was the major fold. It is realized that these arguments are suggestive at/
at best.

Evidence against a? major fold is:

1. Lack of continuity of outcrop. The Lom-Lomskollen psammites are not seen passing into the Vulu-Glonsteinhoe psammites.

ii. Regionally, F minor folds do not show change in shape from Z to S - profile (looking down east plunging fold axes) which would be expected from one limb of the Ottadalen fold to the other. Fold profile data is summarized in Table V.

<table>
<thead>
<tr>
<th>Rock Unit</th>
<th>Percentage Z-profile east plunging F₁ folds</th>
<th>Percentage S profile east plunging F₁ folds</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lom-Lomskollen psammites</td>
<td>65% (13)</td>
<td>35% (7)</td>
</tr>
<tr>
<td>(uninverted)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Vulu-Glonsteinhoe psammites</td>
<td>48% (37)</td>
<td>52% (33)</td>
</tr>
<tr>
<td>(inverted)</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Data from the Vulu-Glonsteinhoe psammites is inconclusive. First minor folds are rare in the Lom-Lomskollen psammites and the sample shown in Table V might not be representative. No conclusion can safely be drawn from these data.
A. Tight $F_1$ fold of phyllonite layering $S_1$ axial planar. 0.7 km, east-north-east of Kyreggi.

B. $F_1$ minor fold in a thin quartzite band, from south slope of Lomskollan. Note boudinage of fold limbs in a direction parallel to the fold hinge and $L_1$ stretching lineation.

C. $L_1$ stretching lineation exactly parallel to $F_1$ fold hinge over observed distance parallel to the hinge of 6.5m, South slope of Lomskollan.

D. Highly attenuated limb of an $F_1$ minor fold, Glomsteinhoe.

E. Clastic feldspar grains lying in the axial plane of an $F_1$ minor fold, Hovilelangtjern.
FIG. 8: F₁ MINOR STRUCTURES.

A. L, STRETCHING LINEATION

B.

C. L, STRETCHING LINEATION

D.

E.
S\textsubscript{i} is sub-parallel to bedding (S\textsubscript{0}) and the sense of S\textsubscript{i}/S\textsubscript{0} intersection is, over the outcrop of the Ottadalen Nappe, inconclusive.

In the next section it is argued that S\textsubscript{i} and F\textsubscript{1} are post-thrusting structures. The lack of correlation of S\textsubscript{i}/S\textsubscript{0} intersections and F\textsubscript{1} fold profiles (Table V) from one limb to the other of the? Ottadalen fold is readily explained if F\textsubscript{1} and S\textsubscript{i} are not synchronous with that fold.

iii. No major fold closure has been seen in the field.

\textbf{c. Post-thrusting structures of the First Movement Phase.}

\textbf{i. S\textsubscript{i} foliation:} S\textsubscript{i} is found throughout the Lom area and is a biotite-muscovite foliation in which lie partially and completely re-crystallized flattened quartz, albite and microcline grains (see plates 3B-F, 5C).

S\textsubscript{i} is usually parallel to cataclastic layering in phyllonites at the base of the Upper Jotun Nappe, at the top and bottom of the Vindsjokamp-Kyreggi group and in phyllonites derived from basal gneisses below the Ottadalen thrust. Locally, however, phyllonite layering is folded by F\textsubscript{1} folds. In F\textsubscript{1} fold cores, S\textsubscript{i} is at a high angle to phyllonite layering and is axial planar to these folds. The phyllonite layering has been folded about S\textsubscript{i} and pre-dates S\textsubscript{i} and F\textsubscript{1} formation (see Fig. 8) Cf. Hossack, 1965, p. 20).

S\textsubscript{i} is parallel to the major thrusts throughout the Lom/
Lom area. At the Ottadalen thrust (exposed only on the north-east slope of Ausfjell), S\textsubscript{1} bedding-foliation in the Lom-Lomskollen psammites is parallel to the phyllonite layering and muscovite-biotite foliation in the basal gneisses.

The Skardalen thrust can be seen at the east end of Skardalen and immediately east of the Vulu river at Sålelssetter. S\textsubscript{1} in orange limestone at the top of the Solåggje-Rivillan group (structural top) is parallel to phyllonite layering in overlying rocks.

The Hovilelangtjern thrust is nowhere exposed but, can be placed in a 10 m. covered interval immediately west of Hovilelangtjern. Phyllonite layering at the top of the Vindsjokamp-Kyreggi group on the north side of the covered interval is parallel to S\textsubscript{1} bedding foliation in rocks of the Hovilelangtjern group.

In the Kvitinge river section south-east of Salellseter, observations similar to those at Hovilelangtjern show that S\textsubscript{1} bedding foliation in the Hovilelangtjern group is parallel to the thrust and phyllonite layering at the base of the Upper Jotun Nappe.

ii. \textit{L\textsubscript{1} Lineation}. The first lineation structure is a fine, faint to strong rodding best developed in psammites. The rodding is defined by ellipsoidal deformed quartz and clastic feldspar grains flattened in S\textsubscript{1} and elongated in the L\textsubscript{1} lineation.

L\textsubscript{1}
L₁ rodding is well developed on the surface of quartz rods (intrafolial segregation quartz, and quartz and pink feldspar (microcline) masses) and rods are invariably elongated in the L₁ direction. F₁ fold hinges are always exactly parallel to L₁. On Lomskollen in the Lom-Lomskollen psammites, L₁ paralleling F₁ minor fold hinges has been seen over exposed distances of as much as 5 m. in the hinge direction. Some of these F₁ folds have wavelengths as small as 7 cm.; wavelength and style in the profile plane are constant over the exposed rectilinear hinge distance.

The trend of first linear structures is variable (see Fig. 9 pocket). From the basal gneisses through most of the Ottadalen Nappe to the lower half of the Soleggje-Rivillan group (sub-areas 1 to 5), L₁ plunges 20° - 30° to 085° - 095°. Structurally above, in sub-area 6, L₁ is distributed in a poor girdle. In sub-area 7, the girdle distribution of L₁ is better developed. The L₁ girdle coincides with the trace of S₁, and, within the girdle, the maximum L₁ concentration is 20° - 30° to 150°.

Locally, L₃ is similar in style to L₁. It has the same trend as L₁ (see Fig. 9 pocket) and in some localities it was not possible to tell whether a particular rodding lineation was L₁ or L₃. However, third minor folds fold L₁ (see Fig. 14), and while L₃ is developed extensively only in areas of strong F₃ folding, L₁ is ubiquitous. Most L₁ and L₃ rodding lineations could/
could be distinguished in the field but it must be accepted that doubtful indentifications could not be avoided.

iii. Minor Folds of the First Movement Phase. Minor folds of the first movement phase which occur in the Lom-Lomskollen and Vulu-Glomsteinhöe psammites, the Hovilelangtjern group and Upper Jotun Nappe phyllonites are nowhere common. Everywhere, first folds plunge exactly parallel to \( L_1 \) (see p. 54, and Fig. 9 pocket). Detailed investigation of these folds and related \( S_1 \) and \( F_1 \) structures has been carried out. The two-fold aim of this work has been to establish the inter-relationships between \( S_1 \), \( L_1 \), and \( F_1 \) and the stress and strain ellipsoids, and establishment of a detailed first movement phase chronology.

**F. minor fold styles:** Ramsay (1962) has developed methods allowing fold style to be readily described. A plot of thickness of a folded layer measured in the direction parallel to the fold axial plane versus distance normal to the axial plane, and a plot of thickness of a layer normal to the layer versus distance along the layer shows whether, in the profile plane (terminology of Fleuty, 1964, is used in this thesis), a fold is concentric, similar, or of intermediate geometry. The thickness of a band measured parallel to the axial plane in a similar fold is constant, and in a plot of this thickness versus distance normal to the axial plane, a straight line of zero slope is obtained. For a similar fold, thickness normal to a band varies along that band, and, in
a plot of thickness versus distance along the band, a curve with maximum values at axial planes is obtained. The latter plot for a concentric fold yields a straight line. Intermediate folds yield curves; both plots reflecting a combination of similar and concentric geometry (fig. 10A).

Nineteen \( F_1 \) folds have been analysed in the Lom area using these methods and results are shown in Fig. 10B-D. Most folds \((13)\) have similar geometry. The remaining 6 folds are intermediate between similar and concentric in their geometry.

Areal distribution of S and Z-profile \( F_1 \) folds has been presented in Table V p. 50, and the relationship of \( L_1 \) to \( F_1 \) minor folds has been described on p. 54.

Strain analysis of first folds. In folds intermediate in geometry between concentric and similar, Ramsay (1962, p. 313) has presented a method, permitting the amount of flattening which a fold has undergone, to be determined. Unfortunately, similar folds retain similar geometry during flattening, and in the majority of first folds at Lom, flattening cannot be determined by Ramsay's methods. Ramsay's analysis is further limited, allowing only two dimensional flattening in the profile plane to be determined. During this plane strain, it is assumed that no deformation in the fold axis direction has occurred. At Lom strong rodding lineation parallel to first fold axes is defined by quartz and feldspar grains elongated in/
FIG. 10A.
Plots obtained from geometrically similar and concentric folds are presented and should be contrasted with plots from natural folds presented in Figs. 10B, 10C, 10D, 13.
FIG. 10A.

PLOT I.

GEOMETRICALLY SIMILAR FOLD

THICKNESS PARALLEL TO AXIAL PLANE

DISTANCE NORMAL TO AXIAL PLANE

SIMILAR FOLD
CONCENTRIC FOLD

PLOT II.

THICKNESS NORMAL TO BANDING

DISTANCE ALONG BANDING

SIMILAR
CONCENTRIC

CONCENTRIC FOLD

DETERMINATION OF FOLD STYLE
A. \( F_1 \) fold of light and dark layers in phyllonite, 2.5 km. south-east of Salellsseter. This is a "similar" fold. In plot I gentle positive and negative slopes of bands A, B and C reflect the acute wedge shape of phyllonite bands.

B. "Similar" \( F_1 \) fold in phyllonite, 2.7 km. south-east of Salellsseter.

C. "Similar \( F_1 \) fold of quartzite band, lower unit of the Hovilelangtjern group, 2.3 km. south of Salellsseter.

D. Modified flexural? \( F_1 \) fold, upper unit of the Hovilelangtjern group, 1.2 km., east-south-east of Hovilelangtjern. Feldspar porphyroclasts lie in the \( S_1 \) axial plane schistosity.
E. "Similar" fold from upper unit of the Hovilelangtjern group, 2.5 km, south-west of Hovilelangtjern.

F. Strongly modified flexural F₁ fold of phyllonite layering 2.1 km, south of Hovilelangtjern.

G. Complex F₁ fold in feldspathic quartzite south slope of Lomskollan. Band A has "similar" geometry while band B₂ is concentric.

H. Modified (thickened in the hinge zone) isoclinal fold. Interbanded micaceous feldspathic quartzite, and quartzite, 5.5 km east of Lom.

I. "Similar" F₁ fold, south slope of Lomskollan.
FIG. 10D.
J. Complex tight \( F_1 \) fold from south slope of Lomskollan 3.3 km. west-north-west of Garmo. Irregularity in plot 1 is possibly due to pre-\( F \) folding variation in band thickness (cf. A of Fig. 10B).

K. "Similar" \( F_1 \) fold from Ausfjell.
in the fold axis direction and flattened in S, axial plane cleavage (see below in more detail). The symmetry of deformed grains is orthorhombic and can be correlated with strain symmetry also orthorhombic (Turner and Weiss, 1963, p. 454-455). During three dimensional deformation, there has been strain in the fold axis direction and Ramsay's (1962) analysis cannot be applied.

Strain analysis using deformed feldspar porphyroclasts has been attempted. In the Lom-Lomskollen psammites, Vulu-Glomsteinhöe psammites and upper unit of the Hovilelangtjern group, feldspar porphyroclasts are numerous enough to be measured in thin section (see plates 3C-F). Porphyroclast phases present include plagioclase, micro-perthite, orthoclase, and microcline. Plagioclase, micro-perthite and orthoclase are not stable phases at middle greenschist metamorphic conditions (Turner and Verhoogen, 1960, p. 108-109, p. 537; Winkler, 1965, p. 76-97) and these minerals must be detrital in origin. Microcline is stable at middle greenschist metamorphic conditions and in the fine grained equigranular groundmass in which porphyroclasts sit, it could have been formed during progressive metamorphism. Large grains of microcline could be porphyroclasts or syn-S porphyroblasts. They are the same size as plagioclase, orthoclase and micro-perthite porphyroclasts. Johnson (1967a, p. 247) indicated that shape and orientation of feldspar porphyroclasts could be used to define the deformation ellipsoid in mylonites. The author has attempted this analysis in the Lom area/
FIG. 11. Feldspar Porphyroclast Data.

A. $S_1$ - $L_1$ mesogranitic fabric in feldspathic quartzite, Gloeinsteinhöe.

B. The planes in which porphyroclasts have been measured.
   i. Plane normal to $S_1$ and $L_1$. The angle between $S_1$ and $Y$ is also determined.
   ii. Plane normal to $S_1$ and parallel to $L_1$. The angle between $Z$ and $S_1$ is also determined.
FIG. II: Feldspar Porphyroclast Data

A.

B.

$S_1$, foliation

$L_1$, 'stretching' lineation

1 cm.
FIG. II.A. LOCATION OF MEASURED SPECIMENS.

DATA FROM 23 LOCALITIES.
### TABLE VI.

**SUMMARY OF FELDSPAR PORPHYROCLAST DATA.**

<table>
<thead>
<tr>
<th>Spec. No. and rock group</th>
<th>Number of grains measured in section (Ls., Ll., Ls., Ll.)</th>
<th>X:</th>
<th>Y:</th>
<th>Z:</th>
<th>k:</th>
</tr>
</thead>
<tbody>
<tr>
<td>65-9B</td>
<td>4.0 0.40</td>
<td>1: 1.58</td>
<td>2.24</td>
<td>1.14</td>
<td>72%</td>
</tr>
<tr>
<td>65-14A</td>
<td>4.0 0.45</td>
<td>1: 2.70</td>
<td>2.93</td>
<td>1.40</td>
<td>47%</td>
</tr>
<tr>
<td>65-14D</td>
<td>5.0 0.40</td>
<td>1: 1.51</td>
<td>2.59</td>
<td>0.60</td>
<td>40%</td>
</tr>
<tr>
<td>65-15C</td>
<td>4.6 0.45</td>
<td>1: 1.76</td>
<td>2.51</td>
<td>0.60</td>
<td>40%</td>
</tr>
<tr>
<td>66-14F</td>
<td>4.0 0.40</td>
<td>1: 1.82</td>
<td>2.77</td>
<td>0.64</td>
<td>40%</td>
</tr>
<tr>
<td>66-11A</td>
<td>4.0 0.40</td>
<td>1: 1.63</td>
<td>2.15</td>
<td>0.51</td>
<td>40%</td>
</tr>
<tr>
<td>66-12D</td>
<td>4.0 0.40</td>
<td>1: 1.61</td>
<td>2.32</td>
<td>0.72</td>
<td>40%</td>
</tr>
<tr>
<td>65-26A</td>
<td>4.0 0.40</td>
<td>1: 1.62</td>
<td>2.62</td>
<td>0.98</td>
<td>40%</td>
</tr>
<tr>
<td>65-42C</td>
<td>4.0 0.40</td>
<td>1: 1.69</td>
<td>2.03</td>
<td>0.28</td>
<td>40%</td>
</tr>
<tr>
<td>66-13C</td>
<td>5.0 0.40</td>
<td>1: 1.91</td>
<td>2.51</td>
<td>0.35</td>
<td>40%</td>
</tr>
<tr>
<td>66-19B</td>
<td>4.0 0.40</td>
<td>1: 1.75</td>
<td>2.29</td>
<td>0.41</td>
<td>40%</td>
</tr>
<tr>
<td>66-20A</td>
<td>7.0 0.40</td>
<td>1: 1.56</td>
<td>2.13</td>
<td>0.65</td>
<td>40%</td>
</tr>
<tr>
<td>66-20B</td>
<td>4.0 0.40</td>
<td>1: 2.02</td>
<td>3.07</td>
<td>1.03</td>
<td>40%</td>
</tr>
<tr>
<td>66-23C</td>
<td>4.0 0.40</td>
<td>1: 1.91</td>
<td>2.61</td>
<td>0.40</td>
<td>40%</td>
</tr>
<tr>
<td>66-24B</td>
<td>5.0 0.40</td>
<td>1: 1.79</td>
<td>2.56</td>
<td>0.54</td>
<td>40%</td>
</tr>
<tr>
<td>66-24C</td>
<td>4.0 0.40</td>
<td>1: 1.59</td>
<td>2.62</td>
<td>1.09</td>
<td>40%</td>
</tr>
<tr>
<td>66-25A</td>
<td>4.0 0.40</td>
<td>1: 1.76</td>
<td>2.29</td>
<td>0.40</td>
<td>40%</td>
</tr>
<tr>
<td>66-25B</td>
<td>4.0 0.40</td>
<td>1: 1.74</td>
<td>2.21</td>
<td>0.37</td>
<td>40%</td>
</tr>
<tr>
<td>65-44A</td>
<td>5.0 0.40</td>
<td>1: 1.92</td>
<td>2.21</td>
<td>0.16</td>
<td>40%</td>
</tr>
<tr>
<td>65-45B</td>
<td>4.0 0.40</td>
<td>1: 2.03</td>
<td>2.62</td>
<td>0.28</td>
<td>40%</td>
</tr>
<tr>
<td>66-41A</td>
<td>6.0 0.40</td>
<td>1: 2.39</td>
<td>2.44</td>
<td>0.002</td>
<td>40%</td>
</tr>
<tr>
<td>66-41C</td>
<td>5.0 0.40</td>
<td>1: 2.03</td>
<td>4.20</td>
<td>1.04</td>
<td>40%</td>
</tr>
<tr>
<td>66-71H</td>
<td>4.0 0.40</td>
<td>1: 1.51</td>
<td>2.02</td>
<td>0.58</td>
<td>40%</td>
</tr>
</tbody>
</table>
Explanation of Table VII: Histograms show the range of variation in the trend of principal axes of feldspar porphyroclasts. The heavy vertical line above the histograms shows the orientation of $S$ in the thin section to which the histogram applies. Where $S_1$ is not apparent in thin section a vertical line is not present. Histograms have been subdivided into four classes:

a. Histogram for measurements made in a section normal to mesoscopic $S_1$ and $L_1$. The variation of the $Y$ principal axis with respect to $S_1$ is depicted.

b. Measurements in a section normal to $S_1$ and parallel to $L_1$. The Histogram depicts the variation in trend of the $Z$ principal axis with respect to $L_1$.

c. Measurements in a section parallel to $S_1$ and $L_1$. The variation in trend of $Z$ with respect to $L_1$ is depicted.

d. Sections normal to $S_1$ and in which there is no visible mesoscopic $L_1$ lineation.
### TABLE VII.

<table>
<thead>
<tr>
<th></th>
<th>a</th>
<th>b</th>
<th>a</th>
<th>d</th>
</tr>
</thead>
<tbody>
<tr>
<td>66-25A</td>
<td><img src="image" alt="Histogram" /></td>
<td><img src="image" alt="Histogram" /></td>
<td>66-41A</td>
<td><img src="image" alt="Histogram" /></td>
</tr>
<tr>
<td>66-25C</td>
<td><img src="image" alt="Histogram" /></td>
<td><img src="image" alt="Histogram" /></td>
<td>66-41G</td>
<td><img src="image" alt="Histogram" /></td>
</tr>
<tr>
<td>65-44A</td>
<td><img src="image" alt="Histogram" /></td>
<td><img src="image" alt="Histogram" /></td>
<td>66-41H</td>
<td><img src="image" alt="Histogram" /></td>
</tr>
<tr>
<td>65-45B</td>
<td><img src="image" alt="Histogram" /></td>
<td><img src="image" alt="Histogram" /></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
area.

In each specimen investigated (23) the arithmetic mean of the ratio of major to minor axes of 40 clastic feldspar grains was measured in each of two sections, one normal to $S_1$ and $L_1$, and the other normal to $S_1$ and parallel to $L$ (see Fig. 11). Variation in trend of the major axes of feldspar porphyroclasts in thin section was noted. An estimate of $S_1$ and $L_1$ orientation was made in each section and compared with the trend of feldspar porphyroclast major axes.

Major intermediate, and minor axes of porphyroclasts have been designated $Z$, $Y$ and $X$ respectively. In thin sections normal to $S_1$ and $L_1$, $Y/X$ was determined, and in thin sections normal to $S_1$ and parallel to $L$, $Z/X$ was determined. As $X$ is common to both ratios, $X: Y: Z$, the axial ratios of feldspar were determined.

Trends of porphyroclast major axes are presented graphically in the percent of grains in a 10° division versus 10° divisions. Estimated $S_1$ and $L_1$ orientations measured in thin section are means of ten readings. These data are presented in Tables VI and VII.

There are definite limitations to this method which must be pointed out. These limitations affect the validity of the assumption that $X$, $Y$ and $Z$ axes of the deformed feldspar porphyroclasts can be correlated with $X$, $Y$ and $Z$ of the strain ellipsoid (terminology of Flinn, 1962, is used in this thesis). Limitations/
Limitations are:

i. Limited occurrence of feldspar porphyroclasts (see page 57). This does not influence reliability but makes it difficult to establish a regional picture of strain variation.

ii. Marginal recrystallization of some porphyroclasts has altered their shape. Recrystallization has preferentially affected major axes of porphyroclasts resulting in inaccuracies in determination of these axes (cf. more highly deformed oolites of Cloos, 1947, p. 879). Recrystallization of the quartz-albite matrix is most extensive in the Nobilelangtjern group followed by Lom-Lomskollen psammites, then the Vulu-Glomsteinhöe psammites. In the first two rock groups, there has been recrystallization of feldspar porphyroclasts; no feldspar porphyroclast recrystallization has occurred in the Vulu-Glomsteinhöe psammites.

The variable extent to which clastic quartz grains have recrystallized renders them unsuitable as strain indicators except locally in the Vulu-Glomsteinhöe psammites.

iii. Other problems affecting reliability of feldspar porphyroclasts as indicators of strain have been discussed by various authors (Higgins, 1964; Hossack, 1965; Hobbs and Talbot, 1966) and are discussed below.

Variation in feldspar porphyroclast or composition affects their viscosity during deformation, probably their rate of strain and, consequently, their final shape. As the majority/
majority of porphyroclasts are K-feldspar (microcline, orthoclase, micro-perthite, see modal analyses, Fig. 4), it is expected that viscosity variation, if any, will have little or no effect. More importantly, it is probable that feldspar porphyroclasts behaved as brittle substances during deformation. Internal deformation of porphyroclasts is, in the vast majority of cases, negligible. Twinning in plagioclase and microcline, and exsolution bands of albite in micro-perthite are undeformed and only in a very few grains is undulatory extinction evident. Many feldspar porphyroclasts have been fractured. This is best seen in thin sections normal to S₁ and parallel to L₁ where porphyroclasts have been elongated in the L₁ direction by fracturing sub-normal to L₁ (see plate 3C, cf. Bossack (1965), fig. 3c; Johnson (1967c) p. 268, and text-fig. 1). Deformation of clastic feldspars has largely been restricted to granulitization at grain boundaries and by brittle fracturing of the grains.

Quartz, making up most of the matrix in which porphyroclastic, has suffered ductile deformation, as shown by strong undulatory extinction and continuity of deformed ribbon-like grains. Different mechanisms of strain have operated in feldspar and quartz and absolute strains undergone by quartz and feldspar grains probably are different (Table VII, specimens 66 - 186, 20A, 20B).

Significance/
Significance of Results of Feldspar Porphyroclast Measurements. The relationships between the fabric of a metamorphic tectonite and the states of stress and strain which give rise to that fabric are discussed before considering the results of feldspar porphyroclast measurements.

Paatterson and Weiss (1961) have rigorously justified the use of symmetry arguments in analysis of metamorphic tectonites. The argument stated by them (p. 359) is: "...the symmetry of any physical system must include those symmetry elements that are common to all the independent factors (physical fields and physical properties of the medium) that contribute to the system...". Applied to rocks, the symmetry of the tectonite fabric is that combination of symmetry elements which are common to both the undeformed body and system of forces giving rise to the deformed body.

Abundant experimental work has borne out the above. (See Turner and Weiss, 1963, p. 327 - 335, 385 - 391). An example is found in experiments conducted by Paatterson and Weiss (1962, 1966). A stress field of axial symmetry was applied to a specimen of phyllite. Symmetry of the specimen was also axial with an axis of infinite rotation normal to the phyllite cleavage. The infinite rotation axis of the stress field was normal to the infinite rotation axis of the specimen during the experiment. The combined symmetry of this system is/
is orthorhombic, and was reflected in the orthorhombic symmetry of conjugate folds produced in the phyllite after deformation.

In three dimensional strain of the type described by Flinn (1962, 1965) only two symmetries of strain are possible - orthorhombic (the majority of cases) and axial. If a sedimentary fabric is not present in a small volume of rock, the symmetry of the array of grains in this volume will be spherical (Patterson and Weiss, p. 871). After three dimensional strain, symmetry of the deformed fabric will be orthorhombic (most cases), or axial.

In structural analysis it is not the final state of strain in a tectonite which is predicted, but just the opposite. If the symmetry of a tectonic fabric is known and if its pre-deformation fabric symmetry is known, the symmetry of deformation can be ascertained. As examples, triaxial oolites and clastic grains flattened in a foliation and elongated in a lineation in that foliation probably have undergone orthorhombic strain. This conclusion is limited by the extent of knowledge of the pre-deformation oolite and clastic grain shapes and orientations. By symmetry argument, principal planes of the strain ellipsoid are equated with the planes of symmetry of the orthorhombic fabric of the tectonite. Thus, in the above examples, the largest principal axis of the oolite or clastic grain is parallel to Z, the major axis/

Applying the symmetry argument further, an orthorhombic or axial stress field must have been responsible for an orthorhombic strain. Principal stress directions coincide with principal strain directions. The X-axis of the strain ellipsoid is parallel to the maximum stress.

**Results.** Clastic feldspar grains have not deformed passively in an amount equal to the deformation in the predominantly quartz matrix. Instead, long axes of detrital feldspars have been rotated towards the direction of extension in the rock (Z of the strain ellipsoid) and the amount of deformation these grains have undergone is not known (cf. Johnson, 1967c). Feldspar porphyroclast data from 23 rocks is presented in Tables VII and VIII.

In the Vulu-Glomsteinhöe psammites and Lom-Lomskollen psammites a consistent pattern has emerged from feldspar porphyroclast measurements. In sections normal to S\textsubscript{1} and parallel to L\textsubscript{1}, 90% of major axes lie within a 45° angle, and the L\textsubscript{1} trend established in thin section bisects this angle. In sections normal to S\textsubscript{1} and L\textsubscript{1}, 90% of the intermediate feldspar porphyroclasts lie in a 100° angle in the plane of this section and S\textsubscript{1} bisects this angle (see Table VIII). In some thin sections normal to mesoscopic S\textsubscript{1} and L\textsubscript{1}, a foliation was not apparent.

In the above rocks there is a strong preferred orientation/
orientation of the major axes of feldspar porphyroclasts in
the direction of the \( L_1 \) lineation, and a poor alignment of
intermediate axes in \( S_1 \).

Flinn's (1962) \( k \)-value has been calculated from feldspar
porphyroclast ellipsoids and ranges from 1.40 to 0.37
(see Table VIII). Because initial porphyroclast shapes, and
the amounts of deformation they have undergone are unknown,
no significance can be attached to \( k \)-values from the point of
view of placing deformation in the field of flattening \((1 > k > 0)\)
or the field of constriction \((< k < 1)\). In three cases where
quartz grains could be measured (see Table VIII), \( k \) of the
quartz matrix was lower than the \( k \)-value derived from the
associated feldspars and the amount of deformation in the quartz
was greater than in the feldspar (as determined from comparison
of \( \Xi \) natural octahedral unit shear, see Hossack, 1965, p. 48).

The strong preferred orientation of feldspar porphyro-
clast major axes indicates considerable extension in the preferred
orientation direction, \( L_1 \), regardless of whether matrix
deformation was in the field of flattening or constriction.
Where \( k \) in deformed matrix quartz can be determined it lies in
the field of flattening and it is tentatively suggested that
flattening with \( 1 > k > 0 \) occurred throughout the Lom-Lomskollen
psammites and the Vulu-Glomsteinhoe psammites.

\( L_1 \); on elongation or stretching lineation defined
by/
by preferred orientation of elongate mineral grains, is
tentatively correlated with Z of the strain ellipsoid. The
$S_1$ foliation is parallel to the $Y-Z$ plane of the strain
ellipsoid. Symmetry of strain is orthorhombic and $\sigma_3$ normal
to $S_1$, $\sigma_2$ lies in $S_1$ and is normal to $L_1$ and $\sigma_3$ is parallel to
$S_1$ and $L_1$.

There are important differences, both in mesoscopic
appearance of $L_1$ and $S_1$, and in feldspar porphyroclast data,
between rocks of the Lom-Lomskollen and Vulu-Glomsteinhöe
psammites contrasted with the Hovilelangtjern group.

In the Hovilelangtjern group, $L_1$ has variable
orientation (see p. 54). It is similar in style but nowhere
as strong as the $L_1$ rodding lineation in the Vulu-Glomsteinhöe
and Lom-Lomskollen psammites. The first foliation, however,
is better defined, especially in sections normal to $L_1$ (see
plates 4B, contrast with plates 3D, F).

In rocks of the Hovilelangtjern group, the quartz-
feldspar matrix has completely recrystallized and cannot be used
to locate the strain ellipsoid. Invariably the shortest axis
of feldspar porphyroclasts is normal to $S_1$. There is a large
variation in intermediate and major axes orientations in $S_1$.
$K$-values range from 0.002 to 1.04, but, again, probably do not
systematically reflect variation (if any) in symmetry of
deformation which may have affected the matrix (see p. 69).
Major axes of feldspar porphyroclasts are statistically parallel to \( L_1 \) (see Tables VII, VIII).

The \( Y - Z \) plane of the strain ellipsoid is tentatively correlated with the \( S_1 \) foliation in the Hovilelangtjern group as intermediate and major axes of feldspar porphyroclasts lie in this plane. \( \sigma_1 \) is normal to \( S_1 \); \( \sigma_2 \) and \( \sigma_3 \) lie in \( S_1 \) and have variable trends over the outcrop of the Hovilelangtjern group. The maximum concentration of \( \sigma_3 \) in the \( Y-Z \) plane, parallel to \( L_1 \), is \( 20^\circ - 30^\circ \) to \( 150^\circ \).

Flattening has been the important deformation in rocks of the Hovilelangtjern group. Possibly, extension in the plane of flattening has not been as important as in rocks at structurally lower levels.

Hossack (1965) has noted that the absolute amount of deformation in pebbles at Bygdin and Olefjell first decreases downwards from the thrust with the overlying Upper Jotun Nappe, and then increases. In the Lom area, \( S_1 \) is synchronous in age throughout the meta-sediments. It post-dates (see p. 52) but is parallel to thrusts in the Lom area. Because of the regional extent of thrusts, especially the Upper Jotun Nappe thrust, they must have been sub-horizontal when formed. \( S_1 \) also was horizontal as its present dip is due to post-\( S_1 \) folding. Load due to overlying rocks will be greatest in the Lom-Lomskollen psammites, and, if the deformation giving rise to \( S_1 \) and \( L_1 \) is due to load of/
of overlying rocks (mainly due to the Upper Jotun Nappe, cf. Hossack, 1965), it is possible that the amount of deformation in the Lom area has increased downwards below the Upper Jotun Nappe thrust plane. Even if k remained constant at all structural levels, the greater the absolute amount of deformation, the more pronounced becomes elongation in the plane of flattening (see fig. 2). This could explain the increase in intensity of L structurally downwards. This problem is further discussed in Chapter V.

The majority of F folds have similar geometry. If they are synchronous with S and L, they cannot have been formed by simple shear parallel to the axial plane as there has been flattening in the axial plane and extension parallel to the fold axis. Probably first folds pre-date S and L, and have been modified by these structures.

Bedding is sub-parallel to S, which is a plane of flattening. Consequently during the deformation producing S and L, if bedding is deformed inhomogenously, it would boudinage, not fold (Flinn, 1962 p. 403). In a few localities boudins trending parallel to L have been seen, and on the south slope of Lomskollen limbs of a tight F fold have boudinaged in a direction parallel to the fold axis (see fig. 8).

First folds post-date thrusting (p. 52), and were not generated during formation of S and L. Nor do they fold S and L, and their formation must have occurred in a phase between/
FIG. 12.
The Y-Z and X-Y principal planes of the strain ellipsoid at $k = 0.5$ and $b = 1, 3, 5$ and 7. Lines 1, 2, 3 and 4 depict the rotation and extension of a line (1) initially at 45 to Z. It is suggested that as at higher values of $b$, directions lie closer to Z in the Y-Z plane than at lower $b$ values (less absolute deformation), a stretching lineation defined by sub-parallelism of long dimensions of deformed and rotated clastic grains will be more intense the greater the value of $b$ (greater absolute deformation).

$$b = a/k.$$
FIG. 12: VARIATION IN INTENSITY OF MESOSCOPIC $L_1$, STRETCHING LINEATION AT CONSTANT $k$-VALUE.
between thrusting and flattening.

The exact parallelism between $F_1$ fold hinges and $L_1$ is not due to rotation of hinges towards $L_1$ during flattening as the condition of exact parallelism is fulfilled only at infinite deformation. The author cannot explain why $F_1$ fold hinges and $L_1$ are parallel.

**Summary.**

i. Flattening ($l > k > 0$) is postulated to be the dominant type of deformation occurring during $S_1$ and $L_1$ formation. $S_1$ is parallel to the $X$-$Z$ plane of the strain ellipsoid and $L_1$ is parallel to $Z$.

ii. Where the quartz matrix has not recrystallized it has been deformed to a greater extent than feldspar porphyroclasts in the same rock and deformation is in the field of flattening.

iii. $S_1$ is well developed everywhere in accordance with deformation being in the field of flattening ($S_1$ and $S > L$ tectonites of Flinn, 1965).

iv. $L_1$ is not equally developed everywhere. It is best developed in the Lom-Lomskollen Vulu-Glomsteinhøe psammites. $L_1$ has the same style in the Hovilelangtjern group but it is not as strongly developed and is of more variable trend. Locally it is not seen in $S_1$.

v. It is tentatively suggested that $L_1$ increases in intensity downwards as the absolute amount of deformation also increases downwards.
d. Strain Chronology of the First Movement Phase.

i. Emplacement of tectonic units and synchronous formation of the Ottadalen fold. Formation of phyllonite layering.

ii. Formation of \( F_1 \) minor folds.

iii. Flattening phase during which phyllonites and \( F_1 \) folds were modified and \( S_1 \) and \( L_1 \) were generated.

The Second Movement Phase.

Introduction. During the second movement phase, minor folds, two types of lineation, crenulation cleavage and schistosity were developed. Second structures were found throughout the Ottadalen nappe but were not found in the Sælleseter Nappes (see Map II, pocket).

Description of Structures of the Second Movement Phase. Orientation of \( F_2 \) fold hinges is variable. \( S_2 \) axial plane cleavage and schistosity is sub-parallel to \( S_1 \), and lies in a girdle whose axis coincides with the \( S_1 \) girdle axis. Girdle distributions of \( S_2 \) and \( S_1 \) have been produced by \( F_3 \) folding (fig. 9). Dihedral angles of \( F_2 \) folds range from close to isoclinal. The geometry in the profile plane of seven folds was analysed by Ramsay's (1962) methods described on p. 55, and the results are/
are presented in Table IX (see also fig. 13A)

TABLE IX.

<table>
<thead>
<tr>
<th>Fold Style</th>
<th>Number</th>
</tr>
</thead>
<tbody>
<tr>
<td>Similar</td>
<td>0</td>
</tr>
<tr>
<td>Flattened-flexural</td>
<td>4</td>
</tr>
<tr>
<td>Flattened-isoclinal</td>
<td>3</td>
</tr>
</tbody>
</table>

Cleavage crenulating \( S_1 \), and \( S_2 \) schistosity parallel to axial planes of second folds, are commonly developed in \( F_2 \) fold cores. (Crenulation cleavage is the term used by Knill (1960) for strain slip cleavage (as defined by Turner and Weiss, 1963, p. 98) and is preferred to that term as it is non-genetic). Tabular micas and chlorite lying in \( S_1 \) are intensely crenulated in "microlithons" between \( S_2 \). Syn-tectonic recrystallization of muscovite and chlorite parallel to \( S_2 \), and post-tectonic crystallization of biotite porphyroblasts parallel to \( S_2 \) has formed a new schistosity in the cores of second folds. \( S_1 \) on the limbs of second folds is undeformed by \( S_2 \) (see plates 5B-D).

With a few exceptions \( L_2 \) is found only in the hinges of \( F_2 \) folds where it is an \( S_2/S_1 \) intersection lineation parallel to fold hinges. Although, regionally, \( F_2 \) fold hinges and \( L_2 \) are variable in orientation, only where \( F_2 \) was overprinted on \( F_1 \),
FIG. 13.
FIG. 13: Style of $F_2$ minor folds.

A. Modified concentric fold in a band of pegmatite in biotite gneiss. North slope of Lomskollan 3.9 km, northwest of Garmo.

B. Modified concentric $F_2$ fold ("flattened-flexural") defined by thin bands of quartzite. Ausfjell.

C. Modified concentric $F_2$ fold in micaceous feldspathic quartzite. West of Glomsteinhoe.
FIG. 13: $F_2$ FOLD STYLES.
FIG. 13A.
FIG. 13A: Representative Second Structures.

A. Detached $F_2$ fold cores of psammite in pelite matrix. Ausfjell.

B. $L_2$ stretching lineation oblique to curvilinear $F_2$ fold hinge. West end of Ausfjell.

C. $F_1$ minor fold folded across $F_2$ fold hinge, 0.4 km. west of Rivillan.

D. $F_2$ isoclinal fold folded by an upright $F_2$ structure.
F₁, and where a stretching lineation oblique to F₂ fold hinges is found, have curvilinear fold hinges been seen. Elsewhere, F₂ fold hinges are rectilinear on the scale of an outcrop (see fig. 13A).

Detached second folds (rootless intrafolial folds of Turner and Weiss, 1963, p.117, fig. 4-246; cf. Knill and Knill, 1958) are fairly common. Detached fold hinges are composed of thin bands of psammite and segregation quartz within pelitic schist. Detached fold hinges show the same variation in orientation as undetached fold hinges (see fig. 9).

Discussion. The following processes can give rise to fold hinges of variable orientations:


v. Rotation of fold hinges during later homogenous or inhomogenous deformation (Flinn, 1962).

The applicability of these processes to the variable trend of F₂ fold hinges is now considered.

Refolding by later folds can account for some of the variation of F₂ fold hinge orientations. Second fold hinges do not trend parallel to F₃ and F₄ fold axes and have been folded/
folded across axial planes of these structures (see fig. 13A). However, on planar limbs of both $F_3$ and $F_4$ folds variable trends of $F_2$ fold hinges are common and refolding by later folds can explain only part of the variation of $F_2$ hinge orientation.

Overprinting of second folds on first folds has been seen at two localities, at the west end of Ausfjell and 1.5 km. north-west of the summit of Soleggje (see fig. 13A). This mechanism of producing variable $F_2$ fold hinge orientation is of only limited importance.

Variable stretching is produced by inhomogenous deformation and is caused by change in the amount of extension in a particular direction from point to point in a tectonite. Only if the stretching direction is at an angle to the fold hinges, will the geometry of the fold in the profile plane change along the fold hinge (cf. Voll, 1960, p. 557, fig. 19a) and the fold hinge be curvilinear. The direction of elongation in a rock is generally marked by a mesoscopic stretching lineation defined by elongate mineral grains (see p. 57). $L_2$ stretching lineation has been seen at Lom at only three localities where it is at an angle to second fold hinges and the $S_2/S_1$ intersection lineation. At one of the localities at the west end of Ausfjell curvilinear fold hinges were associated with the $L_2$ elongation lineation (see fig. 13A). Differential stretching could account for some variation in the trend of $F_2$ fold hinges, but/
but it is difficult to estimate the importance of this mechanism as the $L_2$ stretching lineation is so seldom developed.

Knill and Knill (1958) have described detached fold hinges (rootless intrafolial folds of Turner and Weiss, 1963, p.117) trending at discordant angles (measured in the common axial plane schistosity) to the trend of concordant fold hinges. They suggest that detached fold hinges have been rotated into different orientations than concordant hinges which have not become detached and have a constant trend. In the Lom area, "attached" and detached $F_2$ folds show the same variation in trend and rotation of detached hinges relative to concordant ones has not occurred.

The effect of rotation of hinges during inhomogenous deformation is equivalent to variable stretching and has been discussed. The distinction between $\text{iii.}$ and $v$, is between stretching synchronous with fold formation and stretching which is post-fold formation. The measurements of seven second folds (p.70) suggests that there has been flattening in the axial plane. Supporting evidence comes from syn-$S_2$ quartz grains which have been flattened in the $S_2$ schistosity. Apparently second folds have undergone three dimensional deformation.

Where the $L_2$ stretching lineation is developed it lies in $S_2$ and the symmetry of this fabric is orthorhombic. $S_2$ is tentatively correlated with the $Y-Z$ plane of the strain ellipsoid/
ellipsoid, and the \( L_2 \) stretching lineation is parallel to \( Z \). Possibly, in all second folds, \( S_2 \) can be correlated with the \( Y-Z \) plane of the strain ellipsoid. The lack of a stretching lineation (except in three isolated cases) could mean that almost pure flattening (\( k \) close to zero) has operated. Variability in trend of \( F_2 \) fold hinges would not then be related to flattening as, during axial compression there is no rotation of linear elements in the plane of flattening.

If \( S_2 \) is the \( Y-Z \) plane of the strain ellipsoid, \( \sigma_1 \) was sub-normal to \( S_2 \) during flattening and folds of \( S_1 \) (which is sub-parallel to \( S_2 \)) could not have been generated as \( S_1 \) was in the field of extension (see Flinn, 1962, p. 404). The flattening phase during which \( S_2 \) and locally the \( L_2 \) stretching lineation were developed probably post-dates formation of second folds e.g. variable stretching, where it occurs, is post fold formation.

**Summary and Conclusions.** Second folds are developed on a minor scale only. They are flattened-flexural and flattened isoclinal folds with axial plane crenulation cleavage and schistosity. The lineation parallel to fold hinges is an \( S_2/S_1 \) intersection lineation. Locally, an \( L_2 \) stretching lineation is developed at an angle to the \( L_2 \) fold hinges, and rectilinear fold hinges are curvilinear.

**Factors/**
FIG. 14: Representative Third Minor Structures.

A. Open $F_3$ fold folding $F_2$ detached isocline and $F_2$ quartz-rod, 1.5 km. north-west of summit of Ausfjell.

B. Disharmonic "second" fold in phyllonite 2.1 km. south-east of Hovilelangtjern.

C. "Similar" minor $F_3$ fold from Vulu river.

D. $F_3$ syn-form core, Vulu river.

E. Downward facing folds, Vulu river.

F. $L_1$ stretching lineation folded over $F_3$ minor fold hinge. Generally the angle between $F_3$ and $L_1$ is less than 5°.
FIG. 14: THIRD STRUCTURES.
Factors contributing to variable trend $F_2$ fold hinges in order of importance are:

i. Refolding by $F_3$ and $F_4$ folds.

ii. Variable stretching in a flattening phase post-dating fold formation.

iii. Superposition of $F_2$ folds on $F_1$ structures.

$S_2$ is correlated the Y-Z plane of the strain ellipsoid, and, where $L_2$ stretching lineation is developed, it is parallel to $Z$. $S_2$ is sub-parallel to $S_1$ and the Y-Z plane of the strain ellipsoid of the second movement phase is parallel to the Y-Z plane of the first deformation.

The Third Movement Phase.

Introduction. Major and minor structures have been generated during the third movement phase. Minor structures include folds, lineation and mullion structure. Major structures producing the present dip of tectonic units are synchronous with the minor structures.

Description of Structures.

a. Minor Structures. Minor structures are best developed in sub-areas three, four and five (see fig. 9). Minor fold style/
style is variable. In sub-area three, gently east plunging concentric open folds with steeply north dipping axial planes are found. These folds are plane cylindrical (Turner and Weiss, 1963, p. 108) with a wavelength of 2 to 10 m. in which the middle limb dips down the fold plunge. The S-profile (looking down fold axes) of these folds indicates a major antiform to the north of the Lom area.

Upright east plunging minor third folds are found in the cores of major third folds in sub-areas four and five. Dihedral angles range from open to close and there is a strong rodding lineation parallel to their fold axes.

$L_3$ lineation is well developed in sub-areas four and five where it is a strong rodding and mullion structure. First and third lineations have statistically the same trend (see fig. 9) and locally the same style and in some places in sub-areas three, four and five, they were inseparable in the field (see p.54).

Irregular mullions (Wilson, 1953, p.127) are developed parallel to $L_3$. Curving surfaces defining the mullions are not parallel to bedding, false bedding, or $S_1$. Mullions are best developed in psammite lithology in the cores of major third folds and are restricted to the Vulu-Glomsteinhöe psammites although major $F_3$ folds occur elsewhere.

In the Sålellsseter Nappes no structures of similar style and orientation to second folds found in the Ottadalen nappe have been seen. Second folds in the Sålellsseter Nappes, folding/
### TABLE VII D

DETAILS OF MAJOR F FOLDS, LOM AREA.

<table>
<thead>
<tr>
<th>Fold</th>
<th>Axial Plane</th>
<th>Fold</th>
<th>Dip of</th>
<th>Dip of</th>
<th>Remarks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>strike, dip</td>
<td>trend</td>
<td>north limb</td>
<td>south limb</td>
<td></td>
</tr>
<tr>
<td>Lomskollan anti-form</td>
<td>E-W, steeply N, 20 - 30</td>
<td>690 - 20</td>
<td>30 - 60 to N</td>
<td>65 - 80 to S</td>
<td>Equivalent to the Lomskollan anticline of Strand (1951a). Hinge not exposed in the Lom area (see p. 77).</td>
</tr>
<tr>
<td>Heimfjell fold</td>
<td>E-W, moderately to steeply N, 10 - 30</td>
<td>090 - 10</td>
<td>60 - 85 to S</td>
<td>30 - 70 to S</td>
<td>Complex fold whose geometry varies throughout the Lom area (see p. 77 and fig. 15).</td>
</tr>
<tr>
<td>Vulu synform</td>
<td>E-W, steep to N and S, 20 - 30</td>
<td>085-095</td>
<td>50 - 75 to S</td>
<td>30 to 80 to N and NE</td>
<td>Developed on the middle limb of the Heimfjell fold in the east of the Lom area. The hinge is complex being made up of numerous minor F folds. Strong rodding and mullion structure parallel to fold hinge and best developed in fold core. Downward facing. Conical fold opening eastwards.</td>
</tr>
<tr>
<td>Rivillan antiform</td>
<td>E-W, sub-vertical</td>
<td>090, 25</td>
<td>45 - 85 to N and NE</td>
<td>85 - 40 to S</td>
<td>Downward facing. Conical fold increasing in amplitude and opening out towards the east.</td>
</tr>
</tbody>
</table>

Description of the middle limb of the Heimfjell fold.

- Developed on the middle limb of the Heimfjell fold in the east of the Lom area. The hinge is complex being made up of numerous minor F folds. Strong rodding and mullion structure parallel to fold hinge and best developed in fold core. Downward facing. Conical fold opening eastwards.
FIG. 15.
FIG. 15: THE HEIMFJELL FOLD.
folding $S_1$ and $F_1$ but trending parallel to $L_1$ (see fig. 9) are locally well developed, especially in phyllonites. They are variable in style between similar and strongly disharmonic and usually have moderate to steeply dipping axial planes. Wavelength is variable from 5 to 15 m.

The age of these folds is not known. They pre-date structures of the fifth movement phase but cannot be more closely dated. Correlation with third minor structures is not justified.

b. Major Structures. Girdle distribution of $S_1$ about east plunging axes in sub-areas three, four and five reflects major $F_3$ folding (see fig. 9). The major features of third folding in the Lom area are shown in a block diagram (fig. 15), and north-south cross-sections (fig. 6, pocket).

Third major folding consists of a large antiform the Heimfjell fold (see Table V where structural data pertinent to this fold and other subsidiary major $F_3$ folds are tabulated) in which development of the middle gently east plunging limb increases in importance eastwards.

In the west of the Lom area, the middle limb of the Heimfjell fold is restricted in development so that $S_1$ bedding foliation in a section due south from Lom to Movilelantjern retains the same steep to moderate dip.

In a north-south section at the Vulu river (fig. 15) the middle limb of the Heimfjell fold outcrops over a north-south/
south distance of 1.8 km. It is rucked by 21 upright downward facing third folds with wavelengths ranging from 20 to 200 m. All these folds have constant plunge of 20° - 30° to the east and axial planes dip moderately to steeply north and south.

Immediately west of the Vulu river, the middle limb of the Heimfjell fold is about 1 km wide, east dipping and planar so that folds plicating it are conical, increasing in amplitude and wavelength to the east.

In the east of the Lom area (Glomsteinhöe, Rivillan, (see section fig. 6, pocket) development of folds on the middle limb of the Heimfjell fold is strong with the formation of the Vulu synform, and Rivillan antiform to the south of that fold. See Table VII B for details of these folds.

Vergence of third minor folds on the south dipping north limb of the Heimfjell fold indicates the presence of a large scale antiform to the north of the Lom area. Strand (1951a) has shown that basal gneisses and overlying meta-sediments on Lomskollen lie on the south limb of a major antiform, the "Lomskollen anticline", which folds the contact between basal gneisses and overlying psammites. This fold closes east of Vågåmo. Rund (personal communication) has mapped this closure in the basal gneisses westwards to 7 km. north-east of Lom. It is reasonable to correlate this fold with the inferred antiform north of the Lom area.

Strand (1951a) has grouped together the Lomskollen antiform/
antiform and Faltungsgaben as post-thrusting structures but does not indicate whether these major structures are contemporaneous. This is a difficult problem involving the origin of the Upper Jotun Nappe and is discussed in Chapter V.

Geometry of the Ottadalen Fold. A major F₁ fold closure of the Ottadalen fold has been postulated. Arguments for the existence of this structure have been presented (p. 38, 50-52) and will not be discussed here.

The axial trace of the Ottadalen fold is defined by the line separating uninvolved from involved rocks (Stauffer and Rickard, 1966). Everywhere in the Lom area the uninvolved Lom-Lomskollen and involved Vulu-Gломстенхёе psammites young into the Ausfjell Striped group, and the axial trace of the Ottadalen fold lies within that group. As sedimentary structures have not been found in the Ausfjell Striped group, the axial trace cannot be more clearly defined.

F₃ major folding renders it difficult to work out the style of the Ottadalen fold. In the west of the Lom area where the middle limbs of the Heimfjell fold is not developed, dip of S₁ bedding foliation is moderate to steep to the south from Lom to the southern boundary of the Lom area (see section; fig. 6, pocket). The closure of the Ottadalen fold lies in the Ausfjell Striped group and as S₁ bedding foliation above and below these rocks/
rocks is sub-parallel, limbs of the Ottadalen fold are parallel.

If the axis of the Ottadalen fold were inclined to \(L_3\), a deflection of the \(F_1\) axial trace would be apparent in the field. The axial trace of the Ottadalen fold lies in the Ausfjell Striped group and is roughly parallel to \(F_3\) axial traces, suggesting that first and third major fold axes are roughly co-axial (see section, fig. 16, pocket for possible closures of the Ottadalen fold).

Summary and Conclusions. Minor structures of the third movement phase include folds and lineation. Minor folds are best developed in the core of the Vulu synform and in the flat limbs of the Heimfjell fold. Lineation including mullion structure, is best developed in the cores of major third folds.

The present dip of tectonic units in the Lom area is due to major folding of the third movement phase. The middle limb of the most important major structure, the Heimfjell fold, is complex. It is absent in the west but is important in the east of the Lom area where two major folds, the Vulu synform and Rivillian antiform, make up this limb.

The Heimfjell fold is bounded in the north by the Lomskollen antiform and on the south by the Faltungssgraben. It is not known whether the Lomskollen antiform and Faltungssgraben are/
FIG. 17: Fourth Minor Structures.

A. Minor $F_4$ folds over-printed on the limbs of an $F_3$ fold. 1.1 km. south of Ausfjell.

B. Small $F_4$ folds over-printed on a detached $F_2$ fold core.

C. Deformation of $L_1$ stretching lineation by an open concentric $F_4$ fold. 2.3 km. south-east of Lom.

D. Detail of the conical shape of small $F_4$ chevron folds.
FIG 17 FOURTH STRUCTURES

A.

B.

C.

D.

N.  S.  N.  S.

3M.  10CM.

20CM.  2CM.

L. STRTCHING LINEATION
are synchronous structures.

It is tentatively suggested that the Ottadalen fold is roughly co-axial with $F_3$ major folds, and that this fold is sub-isoclinal in style.

Symmetry of structure of the third movement phase in monoclinic with a single plane of symmetry normal to easterly plunging structures. This plane of symmetry coincides with a principal plane of symmetry of the strain ellipsoid, and the axis normal to the plane of symmetry with a principal direction in the strain ellipsoid.

Fourth Movement Phase.

Description of Structures. Minor upright folds found throughout the Ottadalen Nappe, but best developed in the northern part of the Lom area in sub-areas one to four, are the most important structures formed during this movement phase (see fig. 9, pocket, also fig. 17, and plates 5A, 6A-C).

Fourth folds plunge 30° to 95°; axial planes are sub-vertical, and, looking down fold axes, these folds have Z-profiles. $F_4$ folds have various wavelengths from 1 mm. to 35 m. The larger/
larger folds are best developed in sub-area three where congruous smaller $F_4$ folds are common on limbs of the larger structures. The variability in style of fourth folds is related to lithology. In pelitic schist chevron or kink folds (Turner and Weiss, 1963, p.114) occur. Axial plane cleavage and lineation parallel to fold hinges are absent. In semi-pelite and psammite, fold hinges show varying degrees of curvature and the style of folding is concentric. Dihedral angles of $F_4$ folds are usually less than 90°.

Where $S_1$ has a southerly dip of 65° - 75°, north and south limbs of $F_4$ folds have approximately the same orientation - the attitude of $S_1$ bedding schistosity prior to $F_4$ folding. Thus during $F_4$, rotation of early $S$-planes has been largely confined to the middle north dipping limbs of $F_4$ folds. Where $S_1$ is sub-horizontal (on middle limbs of minor $F_3$ folds) both $F_4$ fold limbs have been rotated, one to a northerly and the other to southerly dip.

Discussion of $F_4$ fold formation. Fourth folds at Lom generally have dihedral angles less than 90° and have planar north dipping limbs (see fig. 17).

Roberts (1966) and Patterson and Weiss, 1962, 1966 have discussed the origin of folds similar in style to fourth folds in the Lom area. Patterson and Weiss (1962, 1966) deformed specimens.
specimens of phyllite orientated at various angles to an axial compressive stress. When compression was parallel to foliation in the phyllite, conjugate kink bands developed. Where these intersected, similar chevron folds with dihedral angles less than 90° and axial planes normal to the direction of compression were generated.

At Lom F, folds are not everywhere developed and, if these folds have originated by the intersection of conjugate kinks locally individual kinks should be seen. Isolated kinks have not been seen in the Lom area and the mechanism of fold formation suggested by Paterson and Weiss (1962, 1966) cannot explain the origin of fourth folds in the Lom area.

Roberts (1966) suggested that when a pre-existing foliation is stressed, conjugate cleavages develop in that foliation. As deformation proceeds one of the conjugate cleavages is suppressed and a set of folds with the other cleavage parallel to its axial plane is developed. In these folds, deformation is initially confined to one limb of the fold which rotates towards the plane normal to the maximum compressive stress. During rotation this limb first undergoes shortening then extension (measured in the plane of the folded surface) so that rotation is caused by translation parallel to the axial plane of the fold, i.e., by similar folding (Roberts, 1966, fig. ).

The deformed limb can rotate no further than the plane of/
of flattening normal to $\sigma_1$. Thus, if $\sigma_1$ was initially sub-parallel to the foliation, a condition assumed by Roberts (1966, p.), dihedral angles less than $90^\circ$ can only be produced by rotation of long limbs. In the latter stages of evolution, both limbs probably undergo three dimensional deformation.

Curvi-planar long limbs of $F_4$ folds have been observed in the Lom area (see fig. 17) and the folds have probably originated by mechanisms proposed by Roberts (1966). Short north dipping limbs have been rotated into a position sub-normal to and the long limbs have rotated varying amounts towards the short limbs producing dihedral angles less than $90^\circ$. Both limbs probably have undergone three dimensional strain.

The fourth fold system has monoclinic symmetry in which east plunging fold axes are normal to the plane of symmetry. This symmetry plane coincides with a principal plane of the strain ellipsoid and the fold axes with a principal direction. The short limb lies in the Y-Z plane of the strain ellipsoid (see above) and during three dimensional deformation either Y or Z coincided with the fold axis.

Fifth Movement Phase.

Introduction. Two sets of minor structures with conjugate geometry/
FIG. 18.
FIG. 18: Representative Fifth Minor Structures.

A. S$^5_A$ crenulation cleavage and quartz-infilled fracture cleavage parallel to the axial plane of an $F^5_A$ fold. Såellsseter.

B. Minor $F^{5_A}$ folds. 3.2 km. east-south east of Salellsseter.

C. Minor $F^{5_A}$ folds in psammite, 4.1 km. south-west of Soleggje.

D. $F^{5_A}$ and $S^{5_A}$ overprinting an $F^1$ isocline; Soleggje.

E. $S^{5_A}$ crenulation cleavage deforming an $F^4$ fold, 1.7 km. north-east of Soleggje.

F. $F^{5_B}$ minor fold and associated $S^{5_B}$ axial plane crenulation cleavage. At Hovilelangtjern.

G. Large $F^{5_B}$ folds immediately west of Hovilelangtjer.

H. Conjugate $F^{5_A}$ and $F^{5_B}$ folds, Rivillan.
FIG. 18: FIFTH STRUCTURES.

A. 10CM.

B. 5CM.

C. 20CM.

D. 10CM.

E. 5CM.

F. 10CM.

G. 10CM.

H. 5CM.
geometry constitute the fifth movement phase. There is ample evidence to show that these sets of structures, designated A and B sets for convenience of description, post-date the fourth movement phase.

Description of Structures. Style, size range, mutual relationships between A and B structures and relationships of these structures to F folds are shown in fig. 18. At Rivillan 1.5 Km north of Kyreggi, south of Soleggje and near Hovilelangtjern, A and B structures developed in the same outcrop have never been seen to interfere. At these localities, conjugate folds have been seen and axial trends in these conjugate folds are parallel to the trends of \( F_{SA} \) and \( F_{SB} \) folds in the same outcrop. Because of this, and because of the regional orthorhombic symmetry of A and B structures, their synchronous formation is postulated.

a. A. Set of Structures. A. structures include \( F_{SA} \) minor folds, \( S_{SA} \) crenulation cleavages and \( S_{1}/S_{SA} \) intersection lineation parallel to \( F_{SA} \) fold hinges.

\( S_{SA} \) is a strong crenulation cleavage best developed in pelites and semi-pelites, and is separated by "microlithons" 1 mm - 2.5 cm. wide. Locally, \( S_{SA} \) is a quartz filled fracture cleavage; quartz veins range in thickness from 2 to 15 mm. Locally syn-\( S_{SA} \), muscovite has recrystallized parallel to \( S_{SA} \) producing a schistosity. Post-tectonic stubby biotite and very small grains of epidote, sphene and magnetite have also recrystallized in the \( S_{SA} \) schistosity.
FIGS. 19, 20: Determination of Principal Stresses During the Fifth Movement Phase, Lom Area.

A. (Fig. 19) Composite plot of 237 poles to $S_{5A}$ and $S_{5B}$ crenulation cleavages in the Lom area. Contours at 0.4%, 1.3%, 2.5%, 4.6%, 8.7% per 1% area.

B. (Fig. 20). Orientation of principal stresses for the fifth movement phase.
\( S_{5A} \) is axial planar to \( F_{5A} \) folds of \( S_1 \). Most folds plunge gently south-west and, looking down plunge, have Z-profiles. \( S_{5A} \) is present in both fold limbs and folds vary from open to close. Variable styles of \( F_{5A} \) folds range from concentric to similar. Disharmonic folds are common and in all cases fold symmetry is monoclinic. Wavelengths of \( F_{5A} \) folds are usually 0.5 cm. to 20 cm, but locally structures 10 m. across have been seen. \( S_1 \) is often intensely crumpled in "microlithons" between \( S_{5A} \) surfaces.

b. B. Set of Structures. B structures include \( F_{5B} \) minor folds, \( S_{5B} \) crenulation cleavage and \( L_{5B} \) intersection lineation parallel to \( F_{5B} \) fold hinges.

\( F_{5B} \) folds are usually "similar" in style, are open to close, have monoclinic symmetry, and are generally disharmonic. \( S_{5B} \) axial plane crenulation cleavage is present in both limbs. Looking west and south-west along sub-horizontal \( F_{5B} \) fold hinges, all folds have S-profiles. Most folds are small, wavelengths of 2.5 to 10 cm. being common, but locally (near Hovilelangtjern), \( F_{5B} \) folds have wavelengths up to 20 m.

Relationships between A and B Structures. \( \sigma_2 \) is parallel to the intersection of axial planes of synchronous \( F_{5A} \) and \( F_{5B} \) folds (Johnson, 1956; Ramsay, 1962b; Roberts, 1966). Thus \( \sigma_2 \) coincides with the axis of the girdle defined by poles to \( S_{5A} \) and \( S_{5B} \) axial plane crenulation cleavages. Fig. 19 is a combined plot of \( S_{5A} \) and \( S_{5B} \) cleavages in the Lom area from which it is seen that \( \sigma_2 \) is/
is sub-horizontal and trends 662°.

In the girdle of $S_{5a}$ and $S_{5b}$ axial plane crenulation cleavages, there is a pronounced gap of poles corresponding to planes striking and dipping in the range 062°, 15° N to 062°, vertical. $\sigma_1$ bisects the obtuse angle of conjugate axial planes (Patterson and Weiss, 1962, 1966; Roberts, 1966) and during the fifth movement phase $\sigma_1$ bisected this gap of planes and plunged 53° to 332°. $\sigma_3$, normal to $\sigma_1$, and lying in the $S_{5a}$, $S_{5b}$ girdle plunged 37° to 152°. The postulated orientation of $\sigma_1$, $\sigma_2$ and $\sigma_3$ are shown in fig. 20.

$F_{5a}$ fold hinges and $S_{5a}/S_1$ intersections lie in a girdle which coincides with $S_{5a}$ (see fig. 9 combined plot of A linear structures in the Lom area). The variability in A linear structures is due to pre-$F_{5a}$ folds in $S_1$. Similarly, $F_{5b}$ fold hinges lie in a sub-horizontal partial girdle coinciding with $F$ (see fig. 20).

The variable trends of A and B linear structures have arisen as the line of intersection of conjugate axial planes is inclined to $S_1$ (Ramsay, 1962b).

**Late Stage Faulting.** Post-$F_5$ faulting has been found in the east of
of the Lom area. North of Rivillan and in Glomsteinhoe a zone of faulting can be traced east-north-east for 3.2 Km (see Map I, pocket).

The fault is defined by a fault breccia 10 - 20 m. wide. The breccia consists of randomly orientated blocks of feldspathic quartzite in a very fine grained structureless matrix. Blocks, invariably angular, range in size from 2 mm. by 10 mm to 5 m. by 9 m. (see fig. 21). L₃ millions are the latest structures seen in these blocks, but as deformation producing the breccia has been brittle, faulting probably post-dates F₅ plastic structures. The breccia zone, east-north-east striking, dips vertically and S bedding foliation on either side of the fault zone has been rotated into parallelism with the fault (see fig. 21).

The deformation of S₁ in the vicinity of the fault suggests that there has been left-hand movement on the fault. Possibly such movement has brought the south dipping limb of the F Rivillan antiform against the north dipping limbs of the F Vulu synform (see fig. 15).

Summary of The Structural Evolution of the Lom Area.

Structures belonging to five movement phases and a period of late faulting have been recognised in the Lom area.
FIG. 21. Structures associated with late-stage faulting.

A. Fault-breccia. Structures found in blocks in breccia include $S_1$, $L_1$, $F_1$, and $L_3$ mullions. Matrix (stippled) is very fine grained quartz and epidote.

B. Rotation of $S_1$ bedding foliation into sub-parallelism with the vertically dipping fault-zone, 0.8 km. north of Rivillan.
i. The First Movement Phase. During this phase, tectonic units have been brought together and penetrative foliation and lineation have been developed. Sub-division of this phase is into:
a. Emplacement of tectonic units and synchronous formation of the Ottadalen fold. Formation of cataclastic rocks.
b. Formation of \( F_1 \) minor folds.
c. Generation of penetrative \( S_1 \) and \( L_1 \) during orthorhombic flattening. As the present dip of \( S_1 \) is probably due to \( F_3 \) major folding, when formed, \( S_1 \) was sub-horizontal. Consequently \( \sigma_1 \) was sub-vertical, \( \sigma_3 \) was parallel to \( L_1 \) as the latter is parallel to \( Z \) of the strain ellipsoid, and trended east-west, and was sub-horizontal. \( \sigma_2 \) lay in \( S_1 \) and was normal to \( \sigma_3 \).

ii. The Second Movement Phase. During this phase only minor folds restricted to the Ottadalen nappe were developed. Fold style is flattened flexural, axial plane crenulation cleavage \( (S_2) \) is sub-parallel to \( S_1 \) and trend of \( F_2 \) fold hinges is variable. Variable trend of fold hinges is due to refolding by \( F_3 \) and \( F_4 \) folds, variable stretching during flattening and superposition of \( F_2 \) on \( F_1 \) folds. \( S_2 \) is a plane of flattening and \( \sigma_1 \) of the second deformation coincided with \( \sigma_1 \) of the first deformation.

iii. Third Movement Phase. Generation of major structures which have produced the present dip of tectonic units occurred during this phase. The most important major structure is the Heinfjell fold which is flanked on the north by the Lomskollen antiform and on the south by the Faltungsgraben. The Heinfjell fold is complex in the east of the Lom area where the Vulu synform and Rivillan
Orientation of principal stresses for five movement phases, Lom area. Sub-scripts 1, 2 and 3 refer to maximum, intermediate and minor principal stresses, respectively. Super-scripts 1-5 refer to the five movement phases.
FIG. 22: ORIENTATION OF PRINCIPAL STRESSES LOM AREA.
Rivillan antiform, major $F_5$ folds are developed in its middle limb. The shape of folded $S$, suggests that $\sigma_1$ lay in the symmetry plane normal to the hinge direction of $F_3$ folds, and was probably sub-horizontal.

iv. Fourth Movement Phase. Minor folds of this phase are restricted to the Øttadalen Nappe. They have been generated in a stress field in which the principal stress, $\sigma_1$ was normal to the north dipping fold limbs. $\sigma_3$ probably coincided with the fold axis direction and $\sigma_2$, 90° from $\sigma_1$, lay in the symmetry plane normal to fold axes.

v. Fifth Movement Phase. Two sets of synchronous structures with conjugate geometry make up this movement phase. The ortho-rhombic symmetry of these structures allows the attitude of the stress field during their formation to be determined.

vi. Late Faulting. A vertically dipping east-north-east striking fault is found in the east of the Lom area. Sinistral movement has occurred on this fault.

Postulated orientations of principal stress axes for the first five deformations are shown in fig.22.
111. GEOLOGY OF SOUTHERN SJODALEN.

INTRODUCTION.

Southern Sjodalen, about 320 Km. north of Oslo, 55 Km. from Lom and 32 Km. from Bygdin, is drained to the north by the Sjoa river. Sjodalen is a wide glaciated valley with a flat floor between three and four Km. wide. An area of 30 Km.² extending from Gjendesheim in the south to Nedre Sjodalsvatn was mapped in the summer of 1966.

Valley sides of Sjodalen are irregular rising 600 - 1000 m. on east and west sides. Topography on valley sides is controlled by varying lithology. Sparagmite rocks stand up as well defined ridges producing a noticeable break in slope above gentle slopes underlain by Cambro-Ordovician rocks. Topography developed on rocks of the Upper Jotun Nappe is a combination of rugged well exposed ground on the steep slopes and poorly exposed moorland topography where slopes are gentle.

In general, exposure is adequate for mapping purposes in country underlain by sparagmite and rocks of the Upper Jotun Nappe especially where slopes are steep; it is poor in ground underlain by Cambro-Ordovician rocks. Access to exposure is good on the west side of Sjodalen and around Gjendesheim. Without a boat to cross Övre Sjodalsvatn, the east side of the valley is inaccessible/
# TABLE IX.

**STRATIGRAPHY AND TECTONIC UNITS.**

**SJODALEN AREA.**

<table>
<thead>
<tr>
<th>Stratigraphy of tectonic units.</th>
<th>Thickness</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper Jotun Nappe.</td>
<td>Initially 14 km?</td>
</tr>
<tr>
<td><strong>Upper Jotun Nappe thrust plane.</strong></td>
<td></td>
</tr>
<tr>
<td>Sjodalen Nappe. Mola - Besstrondfjell</td>
<td></td>
</tr>
<tr>
<td>Semi-pelite unit (top (structural base) not exposed).</td>
<td>&gt;200 m</td>
</tr>
<tr>
<td>Quartzite-pelite unit.</td>
<td>100 m</td>
</tr>
<tr>
<td>Pelite unit.</td>
<td>380-400 m</td>
</tr>
<tr>
<td>Sjodalen Sparagmites.</td>
<td>250-0</td>
</tr>
<tr>
<td></td>
<td>1000 m</td>
</tr>
</tbody>
</table>


inaccessible and was not mapped.

**DESCRIPTION OF TECTONIC UNITS.**

Thrust contacts between tectonic units in Sjodalen have been folded by a major open gently south plunging antiform.

The highest unit, the Upper Jotun Nappe, strikes approximately north-north-east and dips gently east on the east side, and strikes north-south and dips to the west on the west side of Sjodalen.

Immediately below the Upper Jotun Nappe on both sides of the valley are green and grey psammites and micaceous psammites. Below these rocks are interbedded dark grey pelites, thin quartzites and grey semi-pelites. Green and grey psammites are possibly the equivalent of the Valders Sparagmite south of the Jotunheim (the opinion of Dietrichson, 1957, Holtedahl, 1960) and underlying rocks might be Cambrian meta-sediments (see Table XI, map III, pocket).

Cambrian? Mola-Besstrondfjell group.

a. Introduction. Although dark pelites, semi-pelites and interbedded thin quartzites of the Mola-Besstrondfjell group are similar in facies to Cambrian? rocks in the Valdres district, it is accepted that these rocks in the Sjodalen area are not necessarily the same age.

b. Nature of the Contact with the Sjodalen Sparagmites. Dark grey phyllites pass up structurally through a transition zone 50 - 100 m. thick where sparagmite and pelite are interbedded, into greyish-green feldspathic quartzites of the Sjodalen Sparagmites (see fig. 23). Occasional thin beds of sparagmite are found in pelites up to 200 m. below the transition zone, and beds of pelite are found in sparagmite for as much as 100 m. above this zone. The contact between the Sjodalen Sparagmites and Mola-Besstrondfjell group is sedimentary.

c. Stratigraphy of the Mola-Besstrondfjell Group.

i. Pelite Unit. Phyllites and subsidiary quartzites and semi-pelites are found below the transition zone with Sjodalen sparagmites. Thickness of this unit is 350-400 m.

Phyllites are dark grey, rusty weathering, semi-pelites are medium grey and quartzites are light grey and buff, often weathering light rusty brown.

ii. Quartzite-Pelite Unit. The quartzite-pelite unit, approximately 100 m thick, underlies the Pelite unit. Quartzite, feldspathic quartzite/
quartzite, pelite and carbonate are interbedded in layers from 10 cm. to 8 m. thick. Quartzite and pelite are found in equal amounts; feldspathic quartzite and carbonate are of only minor importance.

iii. Semi-Pelite Unit. The semi-pelite unit is composed of homogenous dark grey semi-pelite with a few thin quartzite bands 50 cm. to 1 m. thick. About 200 m. of this unit are exposed; the base was not seen.

Rocks of the Moela-Bestrongfjell group are deformed, have recrystallized, and no sedimentary structures have been preserved.

In thin section quartz, albite, muscovite, chlorite, calcite, zoned and unzoned epidotes and sphene have been identified, and metamorphic grade is lower greenschist (Turner and Verhoogen, 1960, p. 534). Post-third deformation recrystallization has been extensive and rocks are generally strain free. Quartz is usually well annealed. (see plates 8 D–F.)

Eo-Cambrian Sjodalen Sparagmites. Sjodalen Sparagmites are similar in appearance throughout the Sjodalen area. They are feldspathic quartzites and arkoses, light grey and green in colour and invariably well layered. The upper limit of the Sjodalen Sparagmites is the Upper Jotun Nappe thrust plane.

Layering (bedding?) is defined by thin (2 mm. to 1 cm.) bands rich in light green muscovite, detrital feldspars and heavy minerals. Mylonite banding parallel to sedimentary? layering/
FIG. 23: SEDIMENTARY PASSAGE BETWEEN THE MOLA-BESSTRONDJELL GROUP AND SJODALEN SPARAGMITES, BESSA RIVER.
layering is conspicuous near the Upper Jotun Nappe thrust plane.

In thin section plagioclase, orthoclase, perthite and microcline occur as porphyroclasts (see p. 57). These detrital feldspars have been marginally granulitized but have not suffered internal deformation. Also present are quartz, epidote, albite, muscovite, chlorite and sphene and the rocks are lower greenschist in grade. (Turner and Verhoogen, 1960, p. 534).

Evidence of younging has not been seen in the Sjodalen area. The Sjodalen Sparagmites are similar to the Valdres Sparagmites at Bygdin and in the Valdres district, and correlation of these rock units is postulated. Valdres Sparagmite is in sedimentary contact with the Cambrian? Mellsenn formation which is similar to the Mola-Besstrondfjell group below the Sjodalen Sparagmite. The Valdres Sparagmite and underlying Mellsenn formation are inverted (Michelsen, 1967) and probably the Sjodalen Sparagmite and Mola-Besstrondfjell groups are also inverted.

The Grasviki Nappe. The Grasviki nappe composed of granulite facies gneisses is situated at the north end of Ovre Sjodalsvatn (see Map III, pocket). At its north-west boundary the nappe overlies unit II, of the Mola-Besstrondfjell group (see fig. 24) and at its south-east margin it passes under Sparagmite rocks.

Undeformed/
Undeformed gneisses in the centre of the Grasviki nappe are exposed 0.5 Km. east of Besstrondi and at Trollbotthgnappen. These rocks are light green gneisses possessing a foliation defined by lensoid feldspars and slight variation in mafic mineral content. At both upper and lower contacts with meta-sediments, gneisses have been broken down to fine grained well foliated dark green phyllonites.

During phyllonitization chlorite, actinolite, biotite, muscovite, epidote, quartz and albite have been formed from granulite facies minerals. Actinolite, chlorite and epidote give phyllonites their characteristic greenish colour and layering in these rocks is evident in thin section by alteration of layers (2 to 10 mm. thick) rich and poor in these green minerals.

The Upper Jotun Nappe. The Sjodalen sparagmite is everywhere overlain by dark green and grey phyllonites. The phyllonites (see p. 143) pass upwards into less cataclased and uncataclased granulite facies rocks of the Upper Jotun Nappe.

Phyllonites are generally well layered (see plate 2C) and are variable in thickness. East of Grasviki, phyllonites are 150-200 m. thick while north of Gjendesheim only 10 m. are developed. Layering in phyllonites is defined by light and dark coloured bands 1 mm. to 10 mm. thick, which are laterally persistent up to 1000 times their thickness. Layers taper out, however, and in detail are wedge shaped phyllonite bodies.
<table>
<thead>
<tr>
<th>Movement Phase</th>
<th>Planar Structures</th>
<th>Linear Structures</th>
<th>Fold Structures</th>
<th>Occurrence</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pre-deformation</td>
<td>S. Bedding, current bedding? Gneissic foliation in the Grasviki and Upper Jotun Nappes</td>
<td>Lineation defined by elongate mafic mineral clots in gneisses.</td>
<td>None seen.</td>
<td>Sjodalen Sparagmites, Mola-Bessstrondfjell group, phyllonites derived from crystalline nappes</td>
</tr>
<tr>
<td>First</td>
<td>Phyllonite layering $S_1$: Penetrative muscovite-chlorite schistosity in which clastic grains have been flattened. Generally parallel to phyllonite layering and major thrust planes.</td>
<td>$B^*_2$: fine rodding lineation defined by elongate quartz and feldspar grains. Parallel to $B$.</td>
<td>$F_1$: Similar folds with $S$, axial planar and hinges exactly parallel to $L$.</td>
<td>Mola-Bessstrondfjell group.</td>
</tr>
<tr>
<td>Second</td>
<td>$S_2$: Crenulation cleavage and locally a muscovite schistosity developed only in cores of $F_2$ folds.</td>
<td>$B^*_3$:</td>
<td>$F_2$: Flattened flexural folds.</td>
<td>Mola-Bessstrondfjell group.</td>
</tr>
<tr>
<td>Third.</td>
<td>S₃: Geometric axial plane of F₃ folds, locally a crenulation cleavage.</td>
<td>B₃⁺, B₃⁻ Local crenulation</td>
<td>F₃: Variable style concentric to similar, open to tight disharmonic.</td>
<td>All tectonic units.</td>
</tr>
<tr>
<td>--------</td>
<td>-------------------------------------------------</td>
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</tr>
</tbody>
</table>
FIG. 25.
FIG. 25: Transition Between Phyllonites and Granulite Facies Gneisses.

A. West of Gjendesheim.

B. East of Övre Sjodalsvatn. Three distinct zones are recognized -

i. Lower zone of well banded phyllonites.

ii. Intermediate augen phyllonite zone.

iii. Upper zone of well banded phyllonites gradational into gneisses.
FIG. 25: PHYLLONITE TRANSITION ZONES.
Between Gjendesheim and the Bessa river phyllonite development is restricted to a thin zone 10 m. thick where layering can be traced upwards into less cataclased and then into uncataclased rocks (see fig. 25).

Development of cataclastic rocks at the base of the Upper Jotun Nappe in the north east of the Sjodalen area from Griningsdalsknippa to Blossomfjell is extensive and complex (see vertical section fig. 26, pocket).

Here contact with Sparagmites is gradational through a 20 m. zone and is not a clear cut thrust. In this zone bands of sparagmite, thin thrust slices, are interlayered with phyllonite. Above the gradational zone are 30-40 m. of well layered dark phyllonites which pass upward into a large scale "augen phyllonite" zone. In this area large augen ranging from 2 m. by 5 m. to 10 m. by 40 m. of gneiss and pegmatite are set in dark phyllonite. Phyllonite layers are curved and pass around the augen (see fig. 25). Long dimensions of augen lie in the phyllonite layering and the augen phyllonite zone is 40 - 70 m. thick.

Above the augen phyllonite zone a further 50 m. of dark regularly banded phyllonites similar to those below the augen phyllonite zone are developed. Over 10 m. they pass upwards into less, and then uncataclased granulite facies gneisses (fig. 25).

Phyllonites are composed of actinolite, chlorite, biotite, epidote/
epidote, muscovite, sphene, albite and quartz. Occasionally granulite facies minerals are incompletely retrograded and clinopyroxene, olivine, k-feldspar and labradorite have been seen. In general, retrogression of granulite facies assemblages has produced albite, epidote, sphene, actinolite, chlorite, biotite and in particular:

i. Hornblende is retrograded to actinolite at grain boundaries.

ii. Labradorite and k-feldspar contain numerous minute inclusions of epidote and sericite.

iii. Rims of actinolite have formed around clinopyroxene.

Phyllonites are very well recrystallized equigranular rocks with a grainsize about 0.05 mm. Quartz and albite often show well developed annealing textures. Because of the presence of biotite, phyllonites were formed during middle-greenschist sub-facies metamorphic conditions. (Turner and Verhoogen, 1960, p. 537) (see plates 2A, B, D).

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**STRUCTURAL GEOLOGY OF THE SJODALEN AREA.**

**Introduction.**

Four sets of minor structures have been recognised in the Sjodalen area using interference relationships seen in the field/
field and in thin section. Foliation, lineations and folds are associated with each set of structures, and major structures are synchronous with the first and third minor structures. (see Table XII, also map IV, vertical sections, fig. 26, and Structural Analysis, fig. 29, in pocket).

The first foliation strikes north-south and dips east on the east side of Sjodalen and dips to the west on the west side of the valley due to open folding by Øvre Sjodalsvatn antiform. Girdle distribution of S in the northern part of the Sjodalen area reflects the effects of third folding.

Second structures are minor folds of S₁ and have axial plane crenulation cleavage. They are restricted to the Mola-Besstrondfjell group.

Third structures are found on all scales from folds visible only in thin section to structures with half-wavelengths of 2 Km. Third folds plunge gently east and have moderate to steep dipping axial planes. Locally axial plane cleavage is developed.

Fourth structures form a conjugate system with crenulation cleavage of this phase defining a gently south plunging girdle. Superposition of S₄ on the folded S₁ surface has resulted in wide variations in orientation of S₁/S₄ intersections and F₄ fold hinges.

Structural Movement Phases.

First Movement Phase. Structures generated during the first movement phase include thrusts separating tectonic units and which/
which are synchronous with regional inversion of stratigraphy (the Sjodalen nappe, see p. 93), cataclastic rocks produced during thrusting, and later development of minor folds, a penetrative schistosity and lineation.

a. Description of Thrusts.

i. Upper Jotun Nappe Thrust Plane. Phyllonite banding and mylonite layering in Sjodalen Sparagmites are parallel to the Upper Jotun Nappe thrust plane. $S_1$ schistosity is also usually parallel to the thrust, phyllonite and mylonite banding. Locally, however, cataclastic banding is folded by $F_1$ minor folds to which $S_1$ is axial planar (fig. 27). Cataclastic layering pre-dates formation of $S_1$, $F_1$, and $L_1$ (cf. p. 52; Hossack, 1965, p. 20).

ii. Grasviki Thrusts. The Grasviki nappe is in thrust contact at its south east margin with overlying Sjodalen Sparagmites and to the north west with underlying rocks of the Mola-Besstrondfjell group.

In detail, there is no clear cut thrust separating the Grasviki nappe from overlying Sjodalen Sparagmite. In a zone about 30 m. thick below the Sparagmite (see cross sections fig. 26), phyllonite, sparagmite, quartzite and dark grey phyllite are interbanded in layers 2 m. to 20 m. thick. These bands are wedge shaped thrust sheets, not fold cores as no fold closures have been seen in the centres of meta-sediment and phyllonite bands.

Throughout this zone, cataclastic layering and $S$ are parallel to thrusts separating wedge shaped sheets except in isolated cases where layering has been folded by $F_1$ folds.

Emplacement/
FIG. 24: THE GRASVIKI NAPPE.

- Strike and dip of S₁ bedding-foliation.
- Strike and dip of phyllonite layering sub-parallel to S₁.
- Boundary between gneisses and phyllonites
- Thrusts

MOLA BESSTRONFJELL GROUP
- Semi-pelite and pelite
- Quartzite

GRASVIKI NAPPE
- Greensparagmite and pelite
- Light psammite
- Phyllonite
- Gneisses
Emplacement of the Upper Jotun Nappe, Sjodalen Sparagmites and Gravviki nappe was synchronous and pre-$S_1$.

At Besstrondi, below the Gravviki nappe there is a thin thrust sheet 10-30 m. thick composed of green epidote rich sparagmite, buff quartzite and dark grey phyllite. Mylonite layering in the sparagmite and quartzite is parallel to phyllonite layering at the base of the Gravviki nappe. The nappe and thrust sheet were emplaced at the same time in a pre-$S_1$ phase (see p.100). The thrust at the base of the meta-sediment thrust sheet, however, truncates $S_1$ in semi-pelites of the Mol-Besstrondfjell group and is post-$S_1$.

iii. Gjendesheim Thrust. Post-$S_1$ thrusting has also been observed 1.2 Km. east of Gjendesheim and on Mol-flya south east of Ovre Sjodalsvattn.

East of Gjendesheim dark green micaceous psammites dipping gently to the west pass under dark grey phyllites (unit i. of the Mol-Besstrondfjell group, see p.93), which pass upwards concordantly into the Sjodalen Sparagmites. $S_1$ bedding foliation is truncated against $S_1$ in the dark green sparagmite.

The thrust cuts structurally up through the pelite into Sjodalen Sparagmite so that at the Sjoa river 1.6 Km. from Lake Gjende, Sjodalen Sparagmites are in contact with dark green micaceous sparagmite. The thrust can be followed northwards until it passes under Ovre Sjodalsvattn.

Similar observations have been made on the east side of Sjodalen on Mol-flya where a thrust discordant to $S_1$ is exposed/
FIG. 27: First Minor Structures, Sjodalen.

A. $F_1$, fold of well banded phyllonite, 1.2 km. south-east of Gravik. $S_1$ axial planar.

B. Large $F_2$ fold defined by a thin band rich in heavy minerals. Mola river.

C. Feldspar porphyroclasts lying in $S_1$ axial plane cleavage. 2.1 km. west of Bessheim.
FIG. 27: FIRST STRUCTURES.

A.

B.

C.
exposed over a distance of 1.5 km. From north to south, the thrust cuts structurally up through rocks of the Mola-Besstrond-fjell group. A cleavage crenulating $S_1$ in pelites decreases in intensity above the thrust, dying out about 20 m. from the thrust plane.

This thrust is tentatively correlated with the discordant thrust east of Gjendesheim and together they have been termed the Gjendesheim thrust. The Gjendesheim thrust has been folded by the Övre Sjodalsvatn antiform (see map 111, section, fig. 26, pocket), pre-dates that structure, but cannot be more exactly dated. The significance of late thrusting and the amount of movement which has occurred on post-$S_1$ thrusts is not known.

b. Minor Structures.

i. Minor folds. $F_1$ minor folds, developed in all rock types, are nowhere common. First folds are post-phyllonite (see p. 100) and, consequently, are post nappe emplacement. They are "similar" in style (see fig. 27), $S_1$ is axial planar, and $L_1$ is invariably parallel to $F_1$ fold hinges. Wavelengths of first folds range from 0.5 cm. to at least 15 m., the largest seen.

ii. $S_1$, Schistosity and $L_1$, Stretching Lineation. $S_1$ and $L_1$ are penetrative structures developed throughout the Sjodalen area. $S_1$ is a muscovite chlorite foliation defined by the strong preferred orientation of muscovite and chlorite grain boundaries and lenticular strain free quartz and albite grains aligned parallel to the micas and chlorite. In the Sjodalen Sparagmites/
Sparagmites the long and intermediate axes of feldspar porphyroclasts lie close to, or in $S_1$. Locally uncrystallized and only partially recrystallized quartz grains are flattened in $S_1$.

$L_1$ is a fine rodding lineation of variable intensity, defined by preferred orientation of elongate mineral grains. It is relatively constant in orientation in the Sjodalen area, plunging gently to the north-west and south-east.

The mesoscopic style of $L_1$ can be correlated with the micro-fabric, especially in Sparagmite rocks which have not undergone the extensive recrystallization shown by the Mola-Besstrondfjell group and which contain strain indications in the form of feldspar porphyroclasts.

Where mesoscopic, $L_1$ is weak or not visible, tabular micas show a high preferred orientation in $S_1$, and approximately equal long and intermediate axes of flattened quartz grains are parallel to the tabular micas. Long and intermediate axes of feldspar porphyroclasts lie in, or near $S_1$, but, in that plane, have no preferred orientation.

Where $L_1$ is mesoscopically conspicuous, deformed quartz grains are elongate parallel to $L_1$ and there is a strong preferred orientation of major axes of feldspar porphyroclasts in the $L_1$ direction. Porphyroclasts have undergone brittle deformation. Fractures have formed in some feldspars sub-normal to $L_1$, and fragments/
fragments so produced have been separated in this direction. Quartz, more coarse grained than in the matrix, has been deposited in the spaces between fragments. This behaviour reflects considerable extension in the $L_1$ direction (Hossack, 1965; Johnson, 1967a,c; author's work at Lom, see page 57-66) and $L_1$ is a "stretching" lineation (see plates 7C-F).

Fabrics in which $S_1$ is conspicuous, and $L_1$ is not present or weak, probably correspond to $S$ and $S> L$ tectonites of Flinn (1965) and have been produced during flattening ($\lambda > k > 0$). The author's studies at Lom have shown that though there might be a strong preferred linear fabric of feldspar porphyroclasts and deformed quartz grains in a rock, deformation affecting the matrix the rock sits in, generally greater than 75% of the volume of the tectonite, is probably a flattening deformation. Care must be taken when assigning first movement phase linear fabrics developed in tectonites in Sjodalen to Flinn's (1965) tectonite classification.

On the basis of qualitative observations in Sjodalen, the author considers that:

i. $S$ tectonites of Flinn (1965) do exist and are shown by complete lack of $L_1$. They are not common.

ii. Most tectonites are probably $S > L$ tectonites formed during orthorhombic flattening. Intensity of $L_1$ is dependent on the value of $k$ which can range between zero and one, and the absolute amount of deformation affecting the tectonite (see p. 66-67, and fig./
iii. No criteria are available to allow recognition of $L > S$ and $L$ tectonites and their extent is unknown. The general good development of $S_1$ precludes extensive development of these tectonites.

The symmetry of the $S_1 - L_1$ fabric is orthorhombic. $S_1$ is a plane of flattening and can probably be equated with the $X-Z$ plane of the strain ellipsoid. $L_1$ is parallel to $Z$. Consequently, $F_1$ folds could not have been formed during the flattening phase producing $S_1$ and $L_1$ as bedding and/or cataclastic layering ($S_0$) is generally sub-parallel to $S_1$. During flattening, if pre-existing $S$ deformed inhomogenously, it would be by boudinage, not folding (Flinn, 1962, p. 403). $F_1$ folds do not fold $S_1$ and $L_1$; they pre-date these structures and have been modified by them.

c. Strain Chronology of the First Movement Phase.

i. Simultaneous emplacement of the Upper Jotun Nappe, Sjodalen Nappe (inverted?) and Grasviki Nappe. Formation of cataclastic rocks.

ii. Formation of $F_1$ folds.

iii. Formation of $S_1$ and $L_1$ during an orthorhombic flattening phase. Modification of cataclastic rocks and $F_1$ folds.

The/
FIG. 28.
FIG. 28: Second Minor Structures.

A. Detached $F_2$ fold cores in psammite bands and segregation quartz. 1.3 km. east of Gjendesheim.

B. Large minor $F_2$ fold, 1.1 km. west-north-west of Besstrondi.

C. $L_1$ stretching lineation folded across $F_2$ minor fold hinge. Generally the angle between $F_2$ and $L_1$ is less than $5^\circ$. 
FIG. 28: STRUCTURES OF THE SECOND MOVEMENT PHASE.

A. [Diagram]

B. [Diagram]

C. [Diagram]
The Second Movement Phase. During the second movement phase only minor structures restricted to the Mola-Besstrondfjell group have been developed. $F_2$ fold hinges plunge gently north-west and south-east and axial planes are sub-parallel to $S_1$ (see fig. 29 pocket).

Second folds are tight or isoclinal in style. Folded bands up to three times thicker in fold cores than on limbs testify to the considerable flattening undergone by original concentric folds (Ramsay, 1962)(see fig. 28).

A crenulation cleavage restricted to $F_2$ fold cores is parallel to second fold axial planes. This cleavage has intensely deformed $S_1$ (see plate) which is crumpled in "microlithons" between $S_1$ and which locally has been transposed into $S_2$. Limited recrystallization of syn-$S_2$ muscovite parallel to $S_2$ has occurred. Intersection of $S_2$ with folded $S_1$ has produced a lineation parallel to $F_2$ fold hinges.

Most $F_2$ folds have monoclinic symmetry with a plane of symmetry normal to fold axes. A few detached $F_2$ folds with limbs of equal length have orthorhombic symmetry. In monoclinic folds the symmetry plane coincides with a symmetry plane of the strain ellipsoid; the folded form of $S_2$ and the flattened-flexural shape of these folds suggests $\sigma_1$ lies in the plane of symmetry. (cf. Turner and Weiss, 1963, p.456). In detached flattened-flexural folds with orthorhombic symmetry, symmetry planes normal to fold axes and parallel to axial planes coincide with principal planes/
FIG. 30: Representative Third Structures.

A. Large $F_3$ fold in interbanded quartzite and pelite, 3.3 km, west-south-west of Besstrondi.

B. Large $F_3$ fold in phyllonite layering folding an $F_2$ isocline.

C. $L_1$ folded across the hinge of an open concentric $F_3$ minor fold, 0.2 km, west of Grasviki.

D. $F_3$ minor fold folding an $F_1$ isoclinal 0.7 km, south-west of Grasviki.

E. Disharmonic $F_3$ fold in interbanded sparagmite and phyllonite, 0.8 km, south-west of Grasviki.
planes in the strain ellipsoid. The axial plane cleavage \((S_2)\), a plane of flattening, can probably be correlated with the Y-Z plane of the strain ellipsoid. Possibly in all second folds, as they are flattened-flexual, the Y-Z plane can be correlated with \(S_2\). As \(S_1\) is sub-parallel to \(S_2\), the \(\sigma_1\) direction during formation of \(S_1\) coincides with \(\sigma_1\) of the second movement phase.

The Third Movement Phase. Major and minor structures have been developed during the third movement phase. Third folds plunge gently south-east and have produced girdle distributions of \(S_1\) in the north of the Sjodalen area (see fig. 29 pocket).

a. \(F_3\) Minor Folds. Third minor folds are developed on all scales from micro-folds to structures with wavelengths of 100 m. Dihedral angles vary from open to tight. Open folds are generally concentric while tight folds are flattened-flexual in style. In tight folds, especially in highly micaceous rocks, axial plane crenulation cleavage and rodding lineation parallel to fold hinges are common.

\(F_3\) folds are often strongly disharmonic with the dihedral angle varying from band to band within an individual fold/
FIG. 31. Minor fourth structures, Sjodalen area.

A. F4 folds in pelite and thin banded quartzite Besstrond fjell.

B. F4 folds and associated S4 axial plane cleavage.

C. Synchronous S4 crenulation cleavages. 1 km west of Besstrondi.

D. Intersecting S2 and S4 cleavages. Besstrond fjell.

E. L1 rodding lineation folded across an F4 fold hinge. Besstrond fjell.

F. Large F4 folds, Besstrond fjell.
FIG. 31: FOURTH STRUCTURES.
## TABLE XI.
### MAJOR F<sub>3</sub> FOLDS, SJODALEN AREA.

<table>
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<tbody>
<tr>
<td>Gamloset antiform</td>
<td>130 -140, moderately to steeply N.E.</td>
<td>130 -150, gently S.E.</td>
<td>70 to S.W. to 20 to E.</td>
<td>90 to 45 N.</td>
<td>Folds thrust plane separating the Grasviki nappe from Sjodalen Sparagmites and Sparagmites from the Upper Jotun Nappe.</td>
</tr>
<tr>
<td>Besstrondfjell synform</td>
<td>130 -150, sub-vertical.</td>
<td>130 -150, gently S.E.</td>
<td>50 -80 to N.E.</td>
<td>90-70 to S.W.</td>
<td>Exposed only in the Mola-Besstrondfjell group. Dis-harmonic decreasing in amplitude upwards.</td>
</tr>
<tr>
<td>Bessheim antiform</td>
<td>140, moderately S.W.</td>
<td>135 -147, 10 S.E.</td>
<td>150, 20 S.W. to 070, 10 S.</td>
<td>90 -60 to N.E.</td>
<td>Outcrop restricted to the Mola-Besstrondfjell group.</td>
</tr>
</tbody>
</table>
fold (cf. Hossack, 1965; the Bygden Invagination; $F_3$ folds at Lom, p.75-78). Disharmonic $F_3$ folds are well exposed 0.7 km.
south-west of Grasviki seter (fig.30) where a sub-isoclinal $F_1$
fold has a wavelength of 65 m. and amplitude of 70 m. Inter-
banded sparagmites and phyllonites structurally 100 m. above
this fold are unfolded.

b. $F_3$ Major Folds. Major folding of the third movement
phase is responsible for the present dip of tectonic units in the
northern part of the Sjodalen area. There are three important
major folds, the Gamloset antiform, Besstrondfjell synform, and
Bessheim antiform, details of which are presented in Table XLIII
(see also map 111, fig.29, pocket).

Because $L_1$ and $F_1$ fold hinges have the same orientation
as major $F_3$ fold hinges, the style of major folds could not be
determined from analysis of deformed linear structures (Ramsay,
1960).

c. Discussion. Symmetry of $F_3$ folds is monoclinic
with a plane of symmetry normal to fold axes which coincides
with a principal plane of the strain ellipsoid. Because of the
folded form of $S_1$, $\sigma$, probably lies in the girdle normal to
fold axes. Axial planes of third folds dip moderately to steeply
north-east and south-west and $\sigma$ was probably normal to the axial
planes and sub-horizontal in orientation. (Turner and Weiss, 1963,
p.524).
FIG. 32: Determination of Principal Stresses During the Fourth Movement Phase, Sjodalen Area.

A. Composite plot of 115 poles to $S_i$. Contours at 0.9, 2.6, 4.3, 9.6% per 1% area.

B. Orientation of principal stresses.
The Fourth Movement Phase. Crenulation cleavage and minor folds were generated during the fourth movement phase. Orientation of \( S_4 \) and \( F_4 \) are shown in fig. 29 (pocket) and style of structures is depicted in fig. 31.

Poles to \( S_4 \) crenulation cleavage, the most important structure of the fourth movement phase, lie in a girdle whose axis is horizontal and trends north-south. Synchronous development of conjugate crenulation cleavages is suggested at four localities west of Övre Sjodalsvøtt, where cleavages inclined to each other also appear to intersect each other (cf. Rickard, 1961, p. 330). Elsewhere one cleavage has been seen to cut another but regionally there is no consistent pattern to these intersections (see fig. 31). It is suggested that all post \( F_3 \) crenulation cleavages, although not strictly synchronous, belong to one movement phase and form a conjugate system (Turher and Weiss, 1963, p. 466-467). No conjugate folds have been seen. \( S_4 \) is best developed in pelitic and semi-pelitic rocks and is largely restricted to the Mola-Besstrondfjell group.

Although fourth folds vary in style from concentric to similar, most are asymmetric similar folds with monoclinic symmetry. (see fig. 31). In the field, \( F_4 \) folds and \( S_4 / S_4 \) intersection lineation parallel to \( F_4 \) fold hinges show extreme variation in orientation. The girdle axis of \( S_4 \) axial plane crenulation cleavage plunges through folded \( S_1 \) foliation (see fig. 32). This has produced diverging trends of \( S_4 / S_4 \) intersections and \( F_4 \) fold hinges (Ramsay, 1962b).
The orientation of the stress field present during the fourth movement phase can be estimated from the symmetry and orientation of structures of this phase. $\sigma_1$ is assumed to be parallel to the line of intersection of conjugate axial plane cleavages. (Johnson, 1956; Ramsay, 1962b; Roberts, 1966), and consequently has a north-south near-horizontal trend parallel to the girdle axis of $S_\gamma$ crenulation cleavages. $\sigma_1$ and $\sigma_3$ must lie in this girdle. The girdle defined by $S_\gamma$ crenulation cleavages is a partial girdle in which there is an elongate maximum of poles to $S_\gamma$. $\sigma_1$ probably bisects this maximum in the girdle (cf. Hossack, 1965, p. 31) and $\sigma_3$ lying in the girdle is normal to $\sigma_1$. These relationships are shown in fig. 32.

The Övre Sjodalsvatn Antiform. The Övre Sjodalsvatn antiform is a broad open flexure plunging gently to the south-south-west (see Geological Map of Norway, Holtedahl and Dons, 1960). It has produced westerly dips of $S_1$ on the west side, and easterly dips of $S_1$ on the east side of Sjodalen. The age of this fold is not known. It post-dates the Gjendesheim thrust (see p. 101) and/
and could have been formed during post-orogenic regional uplift.

Summary of Structural Evolution in the Sjodalen Area.

In the Sjodalen area four periods of deformation have been recognised. They are:

i. The First Movement Phase. Regionally important, this phase has been sub-divided into:

   a. Emplacement of tectonic units and formation of cataclastic rocks.
   
   b. Formation of F₁ minor folds.
   
   c. Flattening phase during which the penetrative regional schistosity, S₁, and L₁, "stretching" lineation were formed.

As the present dip of S₁ is due to later folding, σ₁ was sub-vertical in orientation, σ₁ was sub-horizontal and trended north-west, south-east and σ₁ was sub-horizontal and normal to

ii. The Second Movement Phase. Formation of minor folds which fold S₁ in the Mola-Besstronfdjell group. These folds are flattened-flexural and it is suggested that σ₁ was sub-normal to the S₂ axial plane cleavage during its formation. The orientation of σ₁ and σ₃ within S₂ is not known. As S₂ is sub-parallel to/
FIG. 33: Postulated Orientation of Principal Stresses During Four Movement Phases, Sjodalen Area. Sub-scripts, 1, 2 and 3 refer to principal stresses, super-scripts, 1-4 refer to the four periods of deformation.
FIG. 33: ORIENTATION OF PRINCIPAL STRESSES
SJODALEN AREA.
to $S_1$, the direction of maximum compressive stress remained parallel during the two deformations.

iii. The Third Movement Phase. Major and minor folds were generated during the third movement phase. $\sigma_7$ probably lies in the plane of symmetry normal to $F_3$ fold hinges and is probably sub-horizontal. $\sigma_2$ and $\sigma_3$ cannot be precisely located. One of these principal stresses is parallel to sub-horizontal south-east trending fold hinges and the other is sub-vertical and lies in the plane of symmetry in which $\sigma_7$ lies.

iv. Fourth Movement Phase. Generation of a conjugate set of minor structures. During this phase orientation of the principal stress axes was $\sigma_1 : 270, 45^\circ$ W; $\sigma_2 : 000^\circ$, horizontal; $\sigma_3 : 090^\circ, 45^\circ$ E.

Postulated orientation of principal stresses for the four movement phases are presented in fig. 33.

Late thrusting which is post- $S_1$ and pre-Ovre-Sjodalsvatn antiform in age cannot be more accurately dated. The Ovre-Sjodalsvatn antiform is probably the latest structure in the Sjodalen area.
TABLE XH
## TABLE XII.

### METAMORPHIC HISTORY, LOM AND SJODALEN AREAS.

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<tr>
<th>Structural Event</th>
<th>chlorite</th>
<th>biotite</th>
<th>muscovite</th>
<th>quartz</th>
<th>actinolite</th>
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<td>F₁ syn-tectonic</td>
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<td>post-tectonic</td>
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<td>F₂ syn-tectonic</td>
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<table>
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<tr>
<th></th>
<th>Lom area</th>
<th>Sjodalen area</th>
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IV, METAMORPHISM.

INTRODUCTION.

In the Lom area the highest metamorphic grade attained in the meta-sedimentary rocks (Ottadalen Nappe, Salålseter Nappes) was middle greenschist. These conditions prevailed during four of five movement phases. In Sjodalen, middle greenschist metamorphic conditions existed during and after generation four sets of structures.

Effects of retrogression in cataclastic rocks derived from the Upper Jotun Nappe in the Lom and Sjodalen area has been described elsewhere (see p. 43). Retrogression is generally complete and mineral assemblages in them indicate the same grade of metamorphism as in underlying meta-sediments. Retrogression of basal gneisses in the Lom area has been described on p. 31.

METAMORPHIC HISTORY IN THE LOM AREA.

The main period of metamorphic crystallization in the Lom area was probably synchronous with $S_1$ and continued into a static phase between $S_1$ and $S_2$. The intensity of this crystallization has varied in the Lom area. In addition, there has been limited recrystallization following each movement phase (see Table XIV).
Syn and Post-\(S_1\) Crystallization.

The syn-\(S_1\) fabric, retained only where later secondary recrystallization has been minimal, is characterized by alignment of very small muscovites (0.05 by 0.01 mm. to 0.01 mm. by 0.002 mm.), chlorites and flattened highly deformed quartz grains in \(S_1\) (see plate 3C). No syn-\(S_1\) biotite has been observed and probably the metamorphic grade at the time of \(S_1\) formation was the lower greenschist (Turner and Verhoogen, 1960, p.534).

Post-\(S_1\) static recrystallization has coarsened the syn-\(S_1\) fabric. Muscovite and biotite (0.6 mm. by 0.1 mm. to 0.1 mm. by 0.02 mm.) have crystallized mainly on \(S_1\) but locally cross-cut it. Deformed quartz grains have polygonized and migration of sub-boundaries, so formed, have produced new less strained grains. Possibly, also quartz grains have been nucleated at grain boundaries of highly deformed quartzes, and have grown in size during the post \(S_1\) recrystallization. The very small grains (diameter less than 0.01 mm. see plate 3C,E) often observed at junctions between deformed quartz grains in syn-\(S\) fabrics which have not recrystallized could be such nuclei (Rast in Pitcher and Flinn, 1965, p. 87-88; Voll, 1960, p.513-520).

The variability of preservation of the syn-\(S_1\) fabric must be considered. Post-\(S_1\) recrystallization in the Vulu-Glomsteinhoe psammites east of the Vulu river is almost non-existent.
existent (see plate 3D). In the Soleggje-Rivallan group, Ausfjell striped group and Lom-Lomskollen psammites, effects of annealing are readily apparent. In addition to post-\(S_1\), muscovite and biotite, quartz, although inequigranular is well recrystallized and is only slightly strained or strain free (see plate 4A). In the Salelsetter nappes, annealing of the syn-\(S_1\) fabric is complete. Micas are strain free and locally have grown randomly across \(S_1\). Annealing texture in quartz is well developed—quartz, quartz triple junctions are common and grain boundaries are straight (see plate 4C).

Possible explanations for the variable preservation of the syn-\(S_1\) fabric are:

i. The amount of deformation which produced the \(S_1\) fabric. The greater the deformation, the greater the amount of recrystallization a material will undergo during later annealing (Burke and Turnbull, 1952).

ii. Annealing temperature during post-\(S_1\), static metamorphism has varied in the Lom area.

Considering i., the least recrystallized rocks, the Vulu-Clomsteinhoe psammites east of the Vulu river, should also be the least deformed. Conversely, rocks of the Salelsetter Nappes were the most intensely deformed. Unfortunately this conclusion contradicts evidence presented in Chapter II (the postulated increase downwards from below the Upper Jotun Nappe thrust plane of the absolute amount of deformation, see p. 68) and/
and cannot be completely accepted.

With respect to ii., mineral assemblages in the Vulu-Glomsteinhöe psammites are not distinctive enough (neither chlorite nor biotite occur) to allow the metamorphic grade of these rocks to be placed in the lower or middle greenschist sub-facies. As it is difficult to visualise a process which would result in the Vulu-Glomsteinhöe psammites being kept at a lower temperature than the rocks structurally below and above them, they were probably at the same P.T. conditions as the rest of the rocks in the Lom area during post-S₁ annealing recrystallization. The second possibility must be rejected.

The abundance of post-S₁ biotite suggests that metamorphic grade was higher in the post-S₁ static recrystallization than the syn-S₁ event. The grade of the static event was middle greenschist, the highest grade attained in the Lom area, and is characterised by the assemblage quartz-albite (An₄₋₅)–muscovite biotite ± chlorite ± epidote developed in pelites (Turner and Verhoogen, 1960, p. 537).

**Later Metamorphic Events.**

Syn and post-S₁ muscovite, biotite and chlorite are intensely crenulated by S₂. Locally, however, there has been syn-S₂ crystallization of muscovite, chlorite and biotite on S₂, so that middle greenschist metamorphic conditions prevailed during and after generation of S₂ (see plate 5C).

Because/
Because cleavage was not developed during the third movement phase and because minor folds of that phase are too large to be studied in thin-section, the metamorphic history during and after generation of third structures is not known.

F₄ folds have strained the syn and post-S₁ fabric. Micas have been strained by gliding on the basal cleavage with production of concentric and kink folds in \{001\}.

Post-F₄ recrystallization of strained micas is common producing interlocking laths of muscovite and biotite parallel to fold limbs. Metamorphic grade remained at middle greenschist during and after F₄ fold formation. Recrystallization, though common, is irregular in its distribution and probably has occurred in those F₄ folds where deformation has been most intense (see plate 6C).

Syn-S₄ and post-S₅₈ recrystallization is common. Small muscovites lying in S₅₈ could be syn-tectonic and there is abundant post-S₅₈ stubby muscovite which cross-cuts these cleavages. Magnetite? and sphene have also recrystallized in S₅₈. No syn or post-S₅₈ biotite has been seen and metamorphic grade at this time was probably lower greenschist.

METAMORPHIC HISTORY IN THE SJODALEN AREA.

Syn and Post-S₁ Crystallization.

Syn/
Syn-S$_1$ muscovite and chlorite have recrystallized on S$_1$. There is a complete gradation between highly deformed quartz grains flattened in S$_1$, which have undergone little or no post-S$_1$ recrystallization to rocks in which the quartz has recrystallized completely giving good annealing texture. The variable aspect of the S$_1$ fabric has arisen from differing amounts of secondary recrystallization in a post-S$_1$ static metamorphism.

During this static metamorphism larger muscovites than syn-S$_1$ micas have crystallized and complete annealing of quartz has resulted in triple junctions and straight quartz-quartz boundaries in an equigranular mosaic of grains. Less perfectly recrystallized quartz is inequigranular, shows undulatory extinction and sutured grain boundaries. Sutured grain boundaries between quartzes in specimens which show the least post-S$_1$ recrystallization are common and indicate that slight annealing has occurred. Very small unstrained grains at boundaries between deformed grains probably represent quartz nuclei developed during deformation and, during annealing, new grains have probably originated by growth from these syn-S$_1$ nuclei and by migration of sub-boundaries in polygonized grains (Rast in Pitcher and Flinn, 1965, p.88; Voll, 1960 p.515; Burke and Turnbull, 1952) (see plates 7E).

Post-S$_1$ static annealing has caused complete recrystallization of quartz in rocks of the Mola-Besstrondfjell group but only/
only partial recrystallization in the Sjodalen Sparagmites. This might be due to a structurally downward increase in the absolute amount of deformation in $S_1$ formation (Burke and Turnbull, 1952; see also p. 68; Hossack, 1965, p. 60).

**Later Metamorphic Events.**

Limited crystallization of muscovite and chlorite in the $S_2$ crenulation cleavage has occurred. Quartz is usually not present in the long limbs of micro-folds of the $S_1$ fabric which define the $S_2$ crenulation cleavage, but is abundant between muscovite layers in the short limbs where it is invariably well annealed. This pattern is common in crenulation cleavages (see Roberts, 1966, p. 348-351; Rickard, 1961) and has been attributed to pressure solution, affecting quartzes in the long limbs which are dissolved to be precipitated in the short limbs. The concentration of carbonaceous material in the long limbs of many $F_2$ micro-folds in Sjodalen suggests that this metamorphic process has operated and was synchronous with $S_2$ formation (see plate 88).

Locally unstrained chlorite and muscovite have crystallized in $S_3$ crenulation cleavage. This material is probably post-$S_2$ (see Rast in Pitcher and Flinn, 1965, p. 89-90) and crystallized in a static phase.

Stubby randomly orientated chlorites cut $S_4$ crenulation cleavage and crystallized in a static phase following formation of that structure. During post-$S_4$ static metamorphism, quartz
in the vicinity of $S_4$ cleavages has been annealed and is strain free (see plate 8D).

In the Sjodalen area metamorphic grade has remained constant throughout four periods of deformation. Following each deformation, static crystallization has lasted long enough to completely remove strain from most rocks.

**CONDITIONS OF METAMORPHISM IN THE JOTENHEIM AREA.**

The metamorphic grade present during and immediately after the generation of the first schistosity in meta-sediments in the Jotenheim area (Hoydalen-Nettoseter-Leirdalen, Lom, Sel-Vågå, Sjodalen and Bygdin) has varied little over a remarkably large area and contrasts strongly with the Scottish Caledonides where, in comparable areas, metamorphic zones are well developed (cf. Kennedy, 1948, 1949).

Barrovian metamorphism has affected meta-sediments in the Jotenheim area (Winkler, 1965, p. 70; Banham, 1962; Holtedahl, 1960, p. 203-204). Evidence that this is so comes from the Hestbreppigan-Hoydalen area (Banham, 1962, 1966) and the southern part of the Trondheim Synclinorium (Holtedahl, 1960, p. 203-204) where three sub-facies within the greenschist facies can be recognised/
recognised. In lower pressure types of metamorphism (Abukuma type, Buchan type), biotite is stable in the lower greenschist facies and andalusite is a stable phase in the upper greenschist facies (Hietanen, 1967, p.193; Winkler, 1963, p.99). In Barrovian metamorphism, biotite is stable only above middle greenschist conditions and in the upper greenschist conditions and in the upper greenschist facies, pyrophillite (okyanite) instead of andalusite is the stable aluminium silicate. In the Jotunheim area, biotite free lower greenschist and andalusite free upper greenschist facies have been recognized (see above) and metamorphism was of the Barrovian type.

Maximum metamorphic grade in the Hoydalens-Nettoseter-Leirdalen and Sel-Vågå areas is in the upper greenschist sub-facies; at Lom, in Sjodalen and at Bygdin, maximum grade is in the middle greenschist sub-facies.

Various authors have attempted to establish P.T. conditions prevailing during Barrovian greenschist metamorphism. McNamara (1965, p.385, fig. 10), assumed that a geothermal gradient of 25 -30 C/KM. was present during greenschist Barrovian metamorphism of Dalradian rocks of south-west Scotland and estimated that at the biotite isograd P = 4kb and T = 400 ° C. He also estimated the P.T. range for the whole greenschist facies to be P: 3kb - 6 kb; T 250 ° C - 500 ° C. Winkler (1965, p.154-159; p.168, fig./
fig. 28) on the basis of experimental data on the break-down of zeolite facies minerals at the beginning of greenschist metamorphism and the break-down of pyrophyllite at the greenschist-almandine amphibolite boundary, suggested $P$: 3kb - 8kb; $T$: 400°C - 550°C as the range of the greenschist facies. Hietanen (1967, fig. 1) suggested $P$: 2.5kb - 6.3kb; $T$: 250°C - 400°C and Turner and Verhoogen (1960, p.534) suggested $P$: 3kb - 8kb; $T$: 300°C - 500°C. In all cases, pressure is equivalent to load pressure and is probably equal to fluid pressure (McNamara, 1965, p.371). Pressures of about 4 - 5 kb and temperature in the order of 450°C probably existed during regional greenschist metamorphism in the Jotunheim area.
V. CONCLUSIONS.

REVISION OF STRATIGRAPHY IN THE JOTUNHEIM AREA, SOUTHERN NORWEGIAN CALEDONIDES.

Lom Area.

The stratigraphic sequence proposed for the Lom area, based on recognition of regional inversion of stratigraphy and the Ottadalen fold, differs considerably from that accepted in the past (e.g. Strand, 1951a, 1961a,b; Holtedahl, 1960). The author's mapping has shown that:

i. The Lom-Lomskollen psammites immediately above basal gneisses are not at the base of the stratigraphic sequence in the Lom area. In the past these rocks were considered Bo-Cambrian, but their age must now be considered unknown.

ii. The Sjodalen-Rivillan group is stratigraphically below the Lom-Lomskollen psammites. The base of this group is not exposed in the Lom area as the upper (structurally) boundary of the Soleggje-Rivillan group is the Skardalen thrust. Thus a complete sedimentary sequence is not present in the Ottadalen fold. Rocks of the Ottadalen Nappe are allochthonous, are derived from an unknown source and are of unknown age.

Rocks above the Skardalen thrust, the Sálellseter Nappes, include/
include eugeosynclinal rocks, quartzites and semi-pelites similar to rocks in the Ottadalen Nappe (lower unit of the Hovilelangtjern group, p. 42) and sparagmites similar in field appearance and mineralogy to the Valdres Sparagmite south of the Jotunheim mountains. The Vindsjokamp-Kyreggi group of eugeosynclinal rocks immediately above the Skardalen thrust is tentatively correlated with the lower part of the eugeosynclinal meta-sediment sequence of the Sel-Vaga area (see Table I; Map I, pocket). The Salellseter Nappes are made up of allochthonous rocks of unknown age and source.

**Sjodalen Area.**

The Sjodalen sparagmites are similar in field appearance and mineralogy to the Valdres Sparagmite, and are in sedimentary contact with underlying pelites, semi-pelite and quartzites of the Mola-Besstrondsfjell group. At Bygdin (Hossack, 1965); and in Valdres (Strand, 1951, 1957; Strand and Holmsen, 1960; Holtedahl, 1960) the Valdres Sparagmite is underlain by the Mellsenn formation which is similar to the Mola-Besstrondsfjell group (see Table I). The Sjodalen Sparagmites are tentatively correlated with the Valdres Sparagmite, and the Mola-Besstrondsfjell group with the Mellsenn formation. In Valdres, the Valdres Sparagmite and Mellsenn formation are inverted (Nichelsen, 1967) and probably the Sjodalen Sparagmites and Mola-Besstrondsfjell group in the/
the Sjodalen area are also inverted.

The Sjodalen Sparagmites and Mola-Besstrondfjell group make up the Sjodalen Nappe, and are allochthonous. The source area, distance and direction of transport of rocks of the Sjodalen Nappe are not known.

Revised Stratigraphy.

The author's work at Lom and in Sjodalen casts strong doubts on stratigraphic correlations made by Strand (1951a) between the Sel-Vågå area and the Valders district.

The east border of the Lom area lies in the western part of the Sel-Vågå area, thus, the Ottadalen fold and regional stratigraphic inversion extends at least into the west part of the Sel-Vågå area. There is continuity in the field between the Lom-Lomskollen psammites and basal light sparagmites in the Sel-Vågå area (Strand, 1951, 1964a; author's observation). As it has been shown that the Lom-Lomskollen psammites are not at the base of the stratigraphic sequence in the Lom area and are of unknown age, the assigned age and position at the base of the sedimentary sequence in the Sel-Vågå area are probably in error.

There is continuity of outcrop of the Mola-Besstrondfjell group (Cambrian?) in the Sjodalen area and Cambro-Ordovician miogeosynclinal rocks of Strand (1951) in the Sel-Vågå area. The possibility that "Cambro-Ordovician" rocks in the Sel-Vågå area are inverted and of different age than Cambro-Ordovician must be considered.
### Table XIII.
### Revised Stratigraphy of the North-East Jotunheim.

<table>
<thead>
<tr>
<th>Lithology</th>
<th>Local Group Names</th>
</tr>
</thead>
</table>
| **Semi-pelite, pelite thin quartzite, minor carbonate.** | **Ausfjell Striped**  
| | **group.** |
| **Light grey and buff feldspathic quartzites arkoses and quartzites** | **Lom-Lomskollen**  
| | **psammites = Vulu-Glomsteinhöe**  
| | **psammites.** |
| **Pelite, quartzite, minor semi pelite and carbonate.** | **Soleggje-Rivillan**  
| | **group.** |
| **Greenish sparagmite.** | **Upper part of the**  
| | **Hovilelangtjern**  
| | **group. In tectonic**  
| | **contact with**  
| | **adjacent rocks.** |
| **Total Thickness** | **3500m.**  
| | **2000m?**  
| | **2000 - 2500m?** |
FIG. 34: STRATIGRAPHY AND STRUCTURE OF NORTH EAST JOTUNHEIM.

MODIFIED FROM STRAND (1961).
The revision of stratigraphy in the Sel-Vågå area and the tentative new stratigraphy proposed for the north east Jotunheim area is best discussed with reference to fig. 34, a block diagram representing the regional geology of this area (see also Map I, pocket) north of Lake Tessa eugeosynclinal rocks (the Otta Nappe are synformally folded into Cambro-Ordovician and Eo-Cambrian sparagmites (cf. Strand, 1951a). The eugeosynclinal rocks are the equivalent of the Vindsjokamp-Kyreggi group in the Lom area, Cambro-Ordovician rocks are equivalent to the Soleggje-Rivillian group and Eo-Cambrian sparagmites of the Sel-Vågå area are equivalent to the Vulu-Glomsteinhöe and Lom-Lomskollen psammites. The Ausfjell striped group does not outcrop in the Sel-Vågå area.

"Cambro-Ordovician" rocks in the Sel-Vågå area are equivalents of the inverted Mola-Besstrondfjell and Soleggje-Rivillian groups and are themselves inverted. A major fold closure equivalent to the Ottadalen fold probably exists in the "Eo-Cambrian" light sparagmites in the Sel-Vågå area. The "Eo-Cambrian" sparagmites are actually younger, not older, than the "Cambro-Ordovician" in the Sel-Vågå area.

This highly tentative revised stratigraphy of the north-eastern Jotunheim is presented in Table XIII.
CORRELATION OF STRUCTURES IN THE NORTH-EASTERN JOTUNHEIM

Considering the intervening distances, the strong similarities in structural sequences worked out at Lom, in Sjodalen and at Bygdin (Hossack, 1965) suggest correlation of movement phases over this area (See Table XIV).

Correlation of Movement Phases.

First Movement Phase. Two lines of argument indicate that thrust emplacement of tectonic units pre-dates formation of the regional $S_1$ schistosity. At Lom, in Sjodalen and at Bygdin first folds to which $S_1$ is axial planar fold cataclastic layering and $S_1$ and $F_1$ must post-date thrusting which is synchronous with cataclastic rock formation.

The regional $S_1$ schistosity, sub-parallel to major thrusts in all areas, is in a plane in which clastic grains and pebbles have been flattened. If flattening and $S_1$ formation were related to thrusting, as the thrusting deformation probably has monoclinic symmetry (simple shear), intermediate pebble axes would not be deformed and the plane of pebble flattening ($= S_1$) would be truncated by the thrust. In fact, pebbles (Hossack, 1965) and clastic grains have been flattened in a plane parallel to the thrust, intermediate axes of pebble have been strained, and deformed pebbles and grains show orthorhombic or axial symmetry/
symmetry (Hossack, 1965, p. 65, also p. 61).

In all areas, pebble and clastic grain deformation, and $S_1$ generation post-date thrusting and are the result of a flattening phase following nappe emplacement. Flattening was largely orthorhombic with $l > k > 0$ (Hossack, 1965, identified areas of constructional deformation surrounded by areas of flattening) and a "stretching" lineation has developed parallel to the direction of greatest elongation in the plane of flattening.

The origin of $F_1$ minor folds is not clear. In all areas these folds must post-date most of the thrusting as they fold cataclastic layering. Hossack (1965) suggested that first folds at Bygdin must have originated at the end of thrusting and before flattening as:

i. At Olefjell, 3 Km. south-east of Bygdin, all first folds have a sense of movement consistent with thrust movement from the north-west.

ii. $F_1$ folds were not formed during the flattening phase post-dating thrusting, as phyllonite layering and bedding, the folded surfaces, were sub-parallel to the plane of flattening and would deform by boudinage, not flattening.

In the Sjodalen and Lom areas, sense of shear suggested by first folds is not constant (see p. 51) and $F_1$ fold hinges are exactly parallel to the $L_1$ stretching direction. Because the sense of shear is not constant, $F_1$ folds probably post-date all thrusting. They could not have been formed during flattening (cf/
(cf. Hossack, 1965, above) and as they do not fold $S_1$ were generated between thrusting and flattening.

Rotation of $F_1$ fold axes into the $L_1$-stretching lineation direction is improbable as this lineation is exactly parallel to $F_1$ fold hinges, a condition which would only be attained at infinite deformation if initially $L_1$ were inclined to $F_1$ fold hinges. There is some relation between $F_1$ fold hinges and $L_1$ or between $F_1$ and $L_1$ and an independent variable (pre-$F_1$ anisotropy in fabric which has controlled $F_1$ and $L_1$ orientation?) which has resulted in $L_1$ being generated exactly parallel to $F_1$.

At Lom there is a swing in the trend of $L_1$ from $20^\circ$ to $30^\circ$ to $85^\circ$ to $95^\circ$ in the Lom-Lomskollen psammites, Ausfjell Striped group, Vulu-Elomsteinhöe psammites and the lower half of the Soleggje-Rivillan group to a girdle distribution containing a well defined maximum at $20^\circ$ to $30^\circ$ to $150^\circ$ to $170^\circ$ in the overlying rocks. No change in $L_1$ trend has been seen in Sjodalen, but at Bygdin (Hossack, 1965) $L_1$ changes in trend upwards through the Mellsenn formation from east-west below to north-west above the Mellsenn. $L_1$ is everywhere a stretching lineation of variable trend post-dating thrusting and it cannot be used as evidence of the direction of nappe emplacement. No satisfactory explanation can be offered for the variation in trend of $L_1$.

In rocks immediately below the Upper Jotun Nappe thrust plane/
plane flattening with extension mainly in a north-west, south-east direction has occurred on a regional scale from Lom through Sjodalen to Bygdin.

**Correlation of Later Movement Phases.** On the basis of similarity in style and relationships of earlier to later structures, the following correlations are suggested (see Table XIV).

Second folds at Lom and in Sjodalen are correlated as they are similar in style, pre-date $F_3$ of each area and post-date $S_1$. Folds of $F_2$ (Lom and Sjodalen areas) style are not developed at Bygdin.

$F_3$ major and minor folds in the Lom area are correlated with $F_3$ folds in the Sjodalen area and $B_2$ major and minor structures at Bygdin (Hossack, 1965). All these folds fold $S_1$, show the same variability of style and between Bygdin and Sjodalen have the same fold hinge and axial plane orientations.

There has been broadly synchronous formation of fourth and fifth movement phase conjugate structures at Lom, fourth phase conjugate structures in Sjodalen and $B_3$ conjugate structures at Bygdin. Differences in orientation of fold elements and of principal stresses responsible for these structures are found between the three areas and while correlation of individual fold systems cannot be made, formation of conjugate structures occurred at approximately the same time in each area.

It is premature at this stage to correlate post-$S_1$, pre-Övre Sjodalæmatn antiform thrusting in the Sjodalen area with/
<table>
<thead>
<tr>
<th>Nettoseter-Leirdalen area.</th>
<th>Lom Area</th>
<th>Sjødalen Area</th>
<th>Vågåmo</th>
<th>Bygdin Area.</th>
</tr>
</thead>
<tbody>
<tr>
<td>B1</td>
<td>-</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>B2</td>
<td>F₁</td>
<td>F₁</td>
<td>F₁</td>
<td>B₁</td>
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<td>F₂</td>
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<td>F₃</td>
<td>F₃</td>
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<td>B₃</td>
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<td>B₃</td>
</tr>
<tr>
<td></td>
<td>F₅</td>
<td></td>
<td></td>
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</tbody>
</table>
with post-B4 thrusting at Bygdin (Hossack, 1965).

Correlation with Areas west of Lom. Correlation of structural sequences between Lom and the Nettoseter-Hoydalen area is hampered by 20 Km. of unmapped intervening ground. Cowan (1966) has postulated that emplacement of the Upper Jotun Nappe was a B2 event (see Table I). Emplacement of this nappe in the Lom area occurred at the beginning of the first movement phase so that S_2 of Cowan is synchronous with S_1 of the author. B1 structures of Cowan pre-date first movement phase structures in the Lom area. Possibly major B2 folding of Cowan (Holleindalen antiform) is synchronous with the Ottadalen fold in the Lom area. Both have caused regional inversion of stratigraphy (see Cowan, 1966, Geological and structural maps; sections 1,2, figs. 6a, 6b).

See Table XIV for other correlations made between the Nettoseter-Hoydalen and Lom areas.

Inter-relationships of Movement Phases.

The orientation of principal stresses responsible for structures of five movement phases in the Lom area and four phases in the Sjodalen area have been discussed above (see p.88-90, fig.22; p.111-112, fig.33). Summarising the stress orientation data:

i. \( \sigma_1 \) during generation of \( S_1 \) and \( L_1 \) coincided with \( \sigma_1 \) present when \( S_2 \) was formed in the Lom and Sjodalen areas.

ii./
ii. Probably \( \sigma_1, \sigma_2 \) and \( \sigma_3 \) of the first movement phase coincided respectively with \( \sigma_3, \sigma_1 \) and \( \sigma_2 \) of the third deformation in the Lom and Sjodalen areas.

iii. \( \sigma_3 \) of the first movement phase is parallel to \( \sigma_1 \) of the fourth phase in the Lom area.

iv. \( \sigma_1, \sigma_2 \) and \( \sigma_3 \) of the fifth movement phase in the Lom area, and of the fourth movement phase in the Sjodalen area did not coincide with any of the pre-existing principal stress orientations in these areas.

In the field these stress patterns are reflected by a parallelism between \( S_1, S_2, \) and \( L_1, L_3 \) in both the Lom and Sjodalen areas, parallelism between \( F_2 \) fold hinges, \( L_1, L_3 \) in the Sjodalen area, and parallelism between \( L_1, L_3, F_4 \) fold hinges in the Lom area.

In four deformations in the Lom area and three in the Sjodalen area, at least one principal axis of the stress ellipsoid was parallel to a principal axis of the stress ellipsoid of the previous deformation. For the first three movement phases in each area, directions of the three principal stresses probably coincided.

The pattern of parallelism of one, two or three principal directions of the stress ellipsoid producing parallelism of mesoscopic structures of differing ages has been described from elsewhere in the Southern Norwegian Caledonides. (Hossack (1965) has/
has suggested that the B\textsubscript{1} stress field was parallel to the B\textsubscript{2} stress field; as a result $L_1$ is parallel to $L_2$ in the Bygdin area and there was probably no time gap between B\textsubscript{1} and B\textsubscript{2}.

A similar origin is suggested for the first three periods of deformation in the Lom and Sjodalen areas. These deformations were due to varying relative magnitudes of the principal stresses in a stress field of constant orientation as follows:

<table>
<thead>
<tr>
<th>Principal stress direction</th>
<th>First Movement Phase</th>
<th>Second Movement Phase</th>
<th>Third Movement Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sub Vertical</td>
<td>$\sigma_1$</td>
<td>$\sigma_2$</td>
<td>$\sigma_3$</td>
</tr>
<tr>
<td>East-west, sub horizontal</td>
<td>$\sigma_3$</td>
<td>$\sigma_2$</td>
<td>$\sigma_2$</td>
</tr>
<tr>
<td>North-south, sub horizontal</td>
<td>$\sigma_2$</td>
<td>$\sigma_3$</td>
<td>$\sigma_1$</td>
</tr>
</tbody>
</table>

Static mineral growth following $S_1$ and $S_2$ formation (see p.114-119) indicate a time break of unknown duration between successive movement phases.

Regionally, from Lom to Sjodalen, rotation of the two sub-horizontal principal stresses from north-south and east-west, to north-east and south-east has occurred. The other principal stress, $\sigma_3$ during $S_1$ regional schistosity and $S_1$ formation, has retained its same sub-vertical orientation from Lom to Sjodalen.
The Upper Jotun Nappe.

The Upper Jotun Nappe presents one of the most perplexing problems in Norwegian geology. An igneous body modified by granulite facies metamorphism, its present thickness (Smithson, 1964) exceeds even the Bushveld Igneous Complex (Turner and Verhoogen, 1960, p. 297), one of the world's largest basic intrusions.

Metamorphic conditions in meta-sediments immediately below the Upper Jotun Nappe allow an estimate of the thickness of the nappe immediately after emplacement. At the Upper Jotun Nappe thrust plane, pressure probably was 4 kb - 5 kb and temperature about 450°C (see p. 121). Assuming a density of 2.86 gm./cc. for the nappe (Smithson, 1964) and assuming load pressure during metamorphism was due to the overlying Upper Jotun Nappe (as metamorphism is no earlier than syn-S1, which was formed immediately after Upper Jotun Nappe emplacement), a nappe thickness of 14.5 Km. to 17 Km. is obtained.

As metamorphic conditions at the Upper Jotun Nappe thrust plane are approximately the same in the Jotunheim area (they range between the middle and upper greenschist sub-facies) depth to the thrust plane must have been everywhere the same - at the time of metamorphism the thrust plane must have been sub-horizontal.

The thickness of 14 Km. - 17 Km. for the Upper Jotun Nappe estimated from metamorphic grade could represent a thinner Upper Jotun Nappe and one or more higher nappes. The Upper Jotun Nappe is the highest tectonic unit identified in the southern Norwegian/
Norwegian Caledonides (Strand, 1961) and the probability of higher nappes cannot at present be evaluated.

Fluid pressures approximating lithostatic pressure were necessary for the emplacement of the Upper Jotun Nappe (Hubbert and Rubey, 1959; Rubey and Hubbert, 1959). The lithostatic fluid pressure was generated in the meta-sediments immediately below the Upper Jotun Nappe by dehydration reactions occurring during regional metamorphism. Because the Upper Jotun Nappe is impermeable, pore fluids could escape only laterally, and lithostatic pressures were established. (Hubbert and Rubey, 1959; Rubey and Hubbert, 1959; Platt, 1962; Henshaw and Zen, 1965; Heard and Rubey, 1965; also McNamara, 1965, p.371: During green-schist facies recrystallization it is believed that an "intergranular fluid was maintained at a pressure approximately equal to the mean lithostatic pressure....Water and CO from deeper down, expelled as long as transitions to higher T assemblages continued streamed...upwards maintaining an abundant fluid supply throughout lower grade metamorphic facies...."

Any hypothesis of the origin of the Upper Jotun Nappe must explain:

i. the thickness of the nappe.

ii. the contact with underlying meta-sediments, everywhere a thrust.

iii. The granulite facies metamorphism the nappe has undergone.

If the Upper Jotun Nappe has originated within the Faltungsgraben/
Faltungsgraben numerous high angle reverse faults along which up-thrusting (a Caledonian event as the Upper Jotun Nappe has been thrust over Caledonian sediments) occurred should have been noted within the outcrop of the Upper Jotun Nappe (see fig. 1).

With the exception of the Tyin-Gjende fault (McRitchie, 1965) no such system of faults has been described within the Upper Jotun Nappe.

If the Upper Jotun Nappe originated from without the Faltungsgraben, then the source area is unknown. A large mafic intrusion in the Pre-cambrian basement underlyin the Valdres district has been identified by Smithson (1964). Possibly the Upper Jotun Nappe is a similar but much larger basic mass intruded into basal gneisses north and north-west of the nappe's present position, and was metamorphosed to the granulite facies in the Pre-cambrian. During evolution of large recumbent nappes postulated by Muret (1960) which now underlie the Basal Gneiss area, the Upper Jotun Nappe could have been thrust south to its present position.

The regional flattening deformation following emplacement of the Upper Jotun Nappe and other nappes occurred when pore pressure decreased from lithostatic to hydrostatic.

The Basal Gneiss Area. Basal gneisses in the Lom area show evidence of metamorphism in the almandine amphibolite facies followed by greenschist facies retrogression. Effects of the greenschist/
greenschist metamorphism extend throughout the basal gneisses. As rocks in the Ottadalen and Sålellseter nappes do not show effects of a metamorphism higher in grade than greenschist, and do not show the effects of an extensive retrograde metamorphism, it is concluded that the greenschist metamorphism is synchronous in meta-sediments and basal gneisses and that the almandine amphibolite metamorphism predates emplacement of the meta-sediments.

The author's work supports completely conclusions reached by Banham (1962), Banham and Elliott (1965), and Cowan (1966). The nature of the contact between meta-sediments and gneisses of the Basal-Gneiss Area is probably everywhere tectonic.

General Considerations of the Evolution of the Southern Norwegian Caledonides.

Muret (1960) and Ramberg (1966) have suggested the types of large-scale tectonics occurring in the southern Norwegian Caledonides. Muret (1960) has proposed that the Basal Gneiss Area is underlain by large Pennine-type recumbent nappes. Eocambrian and Cambro-Silurian rocks were considered cover rocks of nappe cores of basal gneiss. Consequently, cover rocks constituting the Ottadalen and Sålellseter nappes have become detached from core rocks (Ottadalen thrust) and folded (Ottadalen fold) during emplacement of the large nappes underlying the Basal Gneiss Area.

Ramberg (1966) has postulated that the Basal Gneiss Area/
Area represents rocks, because they are less dense (see Ramberg, 1966, p. 8-10), which have risen through overlying more dense rocks (see fig. 59, p.55; fig. 62, p.58). Gneisses upon rising, have spread laterally overriding the more dense rocks, and in shape are like recumbent nappes. The product of this lateral spreading, a nappe sequence similar to that envisioned by Muret (1960), needs only gravity as a driving force and crustal shortening need not be invoked.

In the Lom area, at the edge of the Gneiss Area, nappe evolution as seen by Ramberg is identical to Muret (1960). The Upper Jotun Nappe, because of its high density, cannot be visualized as a body which has risen up through less dense rocks and its emplacement cannot be adequately explained by Ramberg (1966).
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VII. ACKNOWLEDGEMENTS.

I wish to express gratitude to Professor F.H. Stewart for the use of research facilities in the Grant Institute of Geology and to Dr. M.R.W. Johnson for his supervision and critical reading of this thesis.

Thanks are due to Miss J. Tarrant for her help in photographic work.

Field work was carried out with the aid of a scholarship from the Quebec Department of Natural Resources for 1964-1965 and a special scholarship from the National Research Council of Canada for 1965-1967 which are gratefully acknowledged.
PLATE I:

A. Brown biotite replacing hornblende along cleavages in the latter. Basal gneiss, 1 km. east of Lom.

B. Chloritization of hornblende and biotite, from basal gneisses, north slope of Lomskollan.

C. Actinolite (light grey) rims around hornblende (dark grey). From basal gneisses 2.5 km. north-east of Lom.

D. Zoned epidote showing highest birefringence at its margins. Basal gneiss, Lom area.

E. Syn-S, muscovite porphyroblasts strained by $F_4$.

F. Retrograded plagioclase containing numerous minute inclusions of epidote and mica. 1.5 km. west of Carmo.
PLATE 2.

A. Extensively cataclased gneiss derived from the Upper Jotun Nappe. Recrystallization has not occurred in the matrix. Marginally granulitized feldspar augen are numerous. Lom area.

B. Partially retrograded pyroxene. The unaltered pyroxene (dark grey) is rimmed by very fine grained white mica. In phyllonites derived from the Upper Jotun Nappe, Sjodalen area.

C. F1 fold of phyllonite layering Lom area.

D. Granulitized and fractured garnet partially altered to chlorite along fractures. Sjodalen area.

E. Syn-S1 epidote porphyroblast, Upper Jotun Nappe phyllonites, Lom area.

F. Post S1 static growth of epidote. Upper Jotun Nappe phyllonites, Lom area.
PLATE 3.

A. Dark green hornblende porphyroclast surrounded by a rim of pale green actinolite. Vindsjokamp-Kyreggi group, 0.7 km. north-west of Kyreggi, Lom area.

B. Phyllonite immediately above the Skardalen thrust, Skardalen, Lom area.

C. Feldspar porphyroclasts disrupted by fractures sub-normal to \( L_1 \) stretching lineation. Vulu-Glomsteinhoe psammites, Glomsteinhoe, Lom area.

D. Lack of preferred orientation of major axes of feldspar porphyroclasts and major ones of deformed quartz grains in section normal to \( L_1 \) stretching lineation.

E. Strong preferred orientation of feldspar porphyroclast and deformed quartz grain major ones parallel to \( L_1 \) stretching lineation. Some specimens as D, but section cut parallel to mesoscopic \( L_1 \).

D and E, from the Vulu Glomsteinhoe psammites are only slightly recrystallized.

F. \( S_1 = L_1 \) fabric in section normal to \( L_1 \) stretching lineation from Lom-Lomskollan psammites. Contrast greater amount of post-strain recrystallization with D, above.
A. $S_1 - L_1$ fabric in section normal to $S_1$ and parallel to $L_1$. From the same spec. as plate 3F.

B, C. Recrystallized fabric in the Hovilelangtjern group. Contrast with A. above, also Plate 3, C, D, E, F.

D. $F_1$ fold core, axial plane marked. Muscovites, post-$S_1$ static, cut across the axial plane. From Lom-Lomskollan psammites, south slope of Lomskollan.

E. Detail of $F_1$ fold core of D. above.

F. $F_1$ fold defined by folded magnetite rich band. Vulu-Glomsteinhoe psammites, Glomsteinhoe.
A. Detail of $F_1$ closure of Plate 4, $F_1$.

B. $F_2$ fold folding $F_1$ schistosity, Ausfjell Striped group. Limbs have been deformed by $F_4$ crinkles.

C. $F_2$ minor fold Ausfjell Striped group. Fresh biotite inclined to the $S_2$ axial plane cleavage has crystallized following $S_1$.

D. $S_2$ crenulation cleavage in the core of $F_2$ minor fold, Lom area.

E. Detail of crenulation cleavage of D. Syn-$S_2$ and post-$S_2$ biotite has crystallized parallel to $S_2$.

F. $F_4$ minor folds, Ausfjell.
PLATE 6.

A. Lensoid syn-$S_1$ quartz grain folded by $F_4$ micro-fold. Biotites in the fold core have re-crystallized to give a pattern of interlocking laths of biotite parallel to the fold limbs.

B. $S_1$ biotites kinked by $F_4$ folds, Ausfjell.

C. Post-$F_4$ muscovite, biotite and chlorite. Soleggje-Rivillan group.

D. $F_{5A}$ minor folds Hovilelangtjern group.

E. $S_{5A}$ crenulation cleavage and syn and post $S_{5A}$ muscovite which has crystallized on $S_{5A}$.

F. $F_{5B}$ minor fold, Soleggje-Rivillan group.
A. $F_2$ minor folds folded by $F_{5B}$ minor folds, graphite schist, Soleggje-Rivillan group.

B. Post-$F_{5B}$ muscovite inclined to $F_{5B}$ axial plane.

C. $S_1 = L_1$, fabric in Sjodalen Sparagmites Section normal to $S_1$ and parallel to $L_1$, stretching lineation.

D. Same specimen as C. Section normal to $S_1$ and $L_1$.

E. $F_1$ fold in Sjodalen Sparagmites. Thin folded bands show well developed annealing texture.

F. Disrupted feldspar porphyroclast, Sjodalen Sparagmites. Fractures, sub-normal to $L_1$, stretching lineation have been infilled with quartz more coarse grained than in the matrix.
PLATE 8.

A. Intensification of $S_1$ foliation by mimetic crystallization of muscovite on $S_1$.

B. $S_2$ crenulation cleavage in the core of a minor $F_2$ fold, Mola-Besstrondfjell group. Syn-$S_2$ muscovite has crystallized on $S_2$.

C. $F_3$ fold core.

D. $S_4$ crenulation cleavage. Note extensive annealing of quartz following $S_4$.

E. $F_4$ minor fold core. Extensive post-$S_4$ annealing of quartz.

F. Growth of small unstrained clear quartz grains from a large deformed grain.
MAP III
GEOLOGICAL MAP OF THE SJODALEN AREA

LEGEND
- Upper Jotun Nappe
- Gravvik Nappe
- Mola Bessetronfell Group
- Sjodalen Nappe
- Thrust
- Sedimentary Contact
- Strike and Dip of S
- Vertical S

FIG 26: CROSS SECTIONS.