THE STRUCTURAL AND METAMORPHIC HISTORY OF THE SULITJELMA REGION, NORWAY, WITH SPECIAL REFERENCE TO THE NAPPE HYPOTHESIS

K. J. HENLEY


Kautsky's (1953) suggestion that Sulitjelma is a region of nappe tectonics is critically examined from the viewpoint of the lithologies, structural histories and metamorphic histories of the supposed nappes. Four main deformation episodes are distinguished, of which the last three are common to all rocks in the region and thus post-date any thrusting. Only one possible major thrust is recognized, the thrusting occurring between the D1 and D2 deformation episodes near the peak of metamorphism. Vogt's (1927) sequence of metamorphic zones and isograds in the calcareous pelites has been considerably modified and it has been established that there is no correspondence between any of the revised isograds and the essentially syn-metamorphic nappe junction.

K. J. Henley, The Australia Mineral Development Laboratories, Frewville, South Australia 5063

Introduction

The first comprehensive account of the geology of the Sulitjelma region was by H. Sjögren in 1900, who interpreted the rock sequence as a simple stratigraphical succession and the cupriferous pyrite ore bodies as epigenetic - brought in by fluids percolating along a brecciated zone. Th. Vogt (1927) in a classic memoir generalized Sjögren's lithological units and extended his mapping to cover a much larger area, but still assumed the rocks to be in simple stratigraphical succession. Vogt described two sequences of metamorphic zones - one in calcareous schists which developed during prograde metamorphism, and the other in gabbroic rocks, which responded to the prevailing metamorphic conditions and developed a retrograde sequence of zones from the original igneous mineralogy. In his interpretation of the origin of the pyritic ore bodies present in the region, Th. Vogt followed his father J. H. L. Vogt in believing them to be magmatic and related to the gabbro intrusion.

In 1953 G. Kautsky published a structural study of an area in Sweden adjacent to the Norwegian border and extending from a latitude just south of Sulitjelma northwards for about 100 km, and from 10 to 35 km wide. Kautsky interpreted the rock sequence as four superimposed nappes, from
the base upwards – the Pieske, Vasten, Salo and Gasak nappes, all belonging to the 'great Seve nappe', which, to the east, itself overthrusts an imbricate mass of Precambrian granitic basement and thin layers of Hyolithes zone Cambrian sediments (the 'Akkajaure Complex' of Kautsky). The lithological succession in Sulitjelma was interpreted as a superimposed sequence of the Pieske, Vasten and Gasak nappes, the Salo nappe here being absent, and Kautsky suggested that the ore had been deposited by solutions along the thrust zone between the Vasten and Pieske nappes.

R. Mason (1967), in a reinterpretation of Vogt's metamorphic facies series for the basic rock of NE Sulitjelma, has shown that a metamorphosed gabbro complex overlies metavolcanic rocks and that the varied metagabbroic assemblages are due to partial attainment of equilibrium, determined by the availability of water during metamorphism rather than by varying P–T conditions. Mason recognized only one thrust zone in the area studied – namely at the base of the gabbro complex and equivalent to Kautsky's Gasak/Vasten nappe junction, no evidence for a Pieske/Vasten nappe junction being found. The zone is marked by a band of well-lineated flaser gabbro, which is present on the southern contact of the gabbro and which is discordant with the lithological units above and below. It was pointed out by Mason that this thrust zone is essentially syn-metamorphic, and not post-metamorphic as held by Kautsky, although Kautsky does suggest that west of the Norwegian/Swedish border some metamorphism post-dates all the thrusting. Subsequently, however, Mason (personal communication) has shown that the flaser gabbro is limited to the southern contact of the gabbro and that the eastern margin of the gabbro in Sweden has true igneous contacts with the underlying amphibolites.

In a recent paper Nicholson & Rutland (1969) have synthesized their own and many others' work in Nordland in a description and interpretation of the geology of a section across the Norwegian Caledonides, between Bodø in the west and Sulitjelma in the east. Many of the observations and conclusions described in the present paper have been accepted and summarized by Nicholson & Rutland in Parts III and IV of their 1969 paper, dealing with the section 'Hellarmo to the Swedish Frontier', and 'Synthesis and Discussion' respectively. Nicholson & Rutland have demonstrated that the whole of the Sulitjelma succession thins dramatically westwards and plunges underneath their Fauske Marble Group, this latter conclusion being contrary to all previous views of the large-scale tectonics of the region which held that the Sulitjelma schists overlay the Fauske Marble Group (e.g. J. H. L. Vogt 1890, Holmqvist 1900, Vogt 1927, and Kautsky 1947 and 1953). Based on the work of Mason (1966, 1967) and Henley (1968) in Sulitjelma and their own work in adjacent areas, Nicholson & Rutland suggest that there is no evidence for Kautsky's Vasten nappe in Sulitjelma, but that there is a major disjunctive nappe junction corresponding to the base of Kautsky's Gasak nappe at about the level of the main ore horizon, and that the Gasak nappe is to be correlated with the Rödingsfjäll nappe further south.
Table 1. The overall lithological nomenclature of the Sulitjelma region according to various authors

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The main evidence presented by Nicholson & Rutland for the suggested nappe tectonics in Sulitjelma can be summarized as follows:

1. Variation in thickness of the pelitic schists between the ore horizon and the Furulund Granite on passing from south to north of Langvann.
2. Variations in lithology above the ore horizon on passing from south to north of Langvann, much of the southern succession not being recognized in the north.
3. The presence north of the gabbro, at the structural level of Kautsky's Gasak thrust, of granite gneiss and granite gneiss breccia. This is interpreted, as does Kautsky (1953), as allochthonous basement.
4. The apparent inversion of the metamorphic zones, first recorded by Vogt (1927) and confirmed by Henley (1968), the zonal inversion perhaps having been caused by thrusting of the hot Gasak (Rödingsfjäll) nappe over the cooler lower rocks.

The present writer (Henley 1968) has studied the rocks south of the main (northern) ore horizon (Fig. 1) in order to throw light on the controversy outlined above by comparing various features of the so-called nappes and assessing any similarities and differences between them. Wilson (1968) has studied in detail some of the rocks north of the main ore horizon and west of the gabbro, and much of his work is relevant to the present discussion. The main features studied may be described under 1) lithological succession, 2) structural history, and 3) metamorphic history, and each has been con-
Fig. 1. Geological map of the Sulitjelma region, based mainly on Sjögren's (1900) map, but incorporating more recent work by G. Kirchner (1955), R. Mason (1966, 1967), M. R. Wilson (1968) and K. J. Henley (1968). On recent official maps, the geographic names – vann (lake) and elven (the river) appear as – vatn and elva, respectively.
sidered separately for the rocks above, below and immediately adjacent to the ore horizon – termed the Upper, Lower and Junction Units respectively. The Units have been sub-divided into Groups and their correlation with the nomenclature of Vogt (1927), Kautsky (1953) and Nicholson & Rutland (1969) is given in Table I. It should be noted that no evidence of Kautsky's Pieske/Vasten nappe junction has been found (see Mason 1967, and above) and the Upper Unit is therefore taken as equivalent to the Gasak nappe and the Lower Unit as equivalent to the Pieske and Vasten nappes.

Fig. 1 is a geological map of the Sulitjelma region which is based mainly on Sjögren's (1900) map but incorporating more recent work by Kirchner (1955), Mason (1966), Wilson (1968) and Henley (1968). The Baldoaivve Group includes all rocks above the Furulund Group and in the SW is approximately equivalent to Vogt's (1927) Baldoaive Schist; the Furulund Group is approximately equivalent to Vogt's Furulund Schist; while the Sjonstå Group corresponds to Nicholson & Rutland's (1969) use of the term. Baldoaivve, Furulund and Sjonstå are all localities in or near the region.

Lithological succession

Lower Unit

The lowest rocks of the Lower Unit which have been studied belong to the Sjonstå Group, which in the east has been metamorphosed at low grade and now consists of green chloritic phyllites with abundant quartz/carbonate segregations (the Muorki Schist) and with minor beds of quartzite and marble. In the west the rocks of the Sjonstå Group have been metamorphosed under high-grade conditions and consist of massive, flaggy psammitic schists above a thick sequence of coarse-grained aluminous pelitic schists, often with garnet (Nicholson & Rutland 1969).

The overlying Furulund Group consists of a thick sequence of closely banded calcareous pelitic schists (the Furulund Schist) often with garnet and/or hornblende, overlain in the NE by an interbedded sequence of schistose amphibolites and calcareous pelitic schists. The highest bed in the succession is a metaporphyratic amphibolite, which is cut out to the west by the flaser gabbro (Mason 1967). The metamorphic grade of the Furulund Schist increases from east to west and the first appearance of garnet and hornblende constitutes a well-defined isograd.

Isolated sills and lenses of metadolerite are common in the Lower Unit, especially in the lower part of the Furulund Schist. They have a completely metamorphic mineralogy and presumably represent formerly continuous sheets of dolerite which were intruded prior to, or contemporaneously with, the main period of metamorphism and were subsequently disrupted during a phase of deformation. Chemically the metadolerites resemble tholeiitic basalts and are distinctly different from the gabbroic rocks in the Upper Unit, containing more \( \text{SiO}_2 \), \( \text{TiO}_2 \), \( \text{FeO} \), \( \text{MnO} \) and \( \text{K}_2\text{O} \) and less \( \text{Al}_2\text{O}_3 \), \( \text{MgO} \) and \( \text{CaO} \) than the gabbroic rocks (Henley 1968).
Upper Unit

In the southern Baldoaivve Group the lowest rocks are aluminous pelitic and semi-pelitic schists and these are overlain by calcareous pelitic schists which often contain hornblende but rarely garnet. The highest bed in the succession is a graphitic schist with or without garnet. Vogt (1927) records a kyanite-staurolite-garnet-mica assemblage from this last rock-type. Granitic and trondhjemitic rocks are common in the upper parts of the succession and a thin sill of granite (the Furulund Granite) occurs near the base. Metadolerites are very rare throughout.

Certain lithological changes take place in passing from the southern to the northern Baldoaivve Group, analogous to the facies changes recorded by Vogt within the southern Baldoaivve Group itself (Vogt 1927, p. 113). Thus, although the lowest rocks of the northern Baldoaivve Group are predominantly aluminous schists with staurolite and/or kyanite, the overlying rocks are mainly rusty or banded psammites (Wilson 1968) and the uppermost rock type is a silvery garnet phyllite (Vogt's Rötind Schist). The Furulund Granite here reaches its maximum development, and amphibolites and marble bands are common. In the east is the large intrusion of the metamorphosed Sulitjelma gabbro, at the base of which is the well-lineated flaser gabbro (Mason 1967). The flaser gabbro thins to nothing in the west and to a few metres near the Swedish border in the east (Mason 1967).

Junction Unit

The rocks in and near the ore horizon are characteristically brecciated and chloritized. This zone of brecciation and chloritization varies in thickness from less than a centimetre to 10-15 metres, the thickest development occurring around Furuhaugen, Sagmoen and north of Langvann. Sjögren (1900) mapped the zone as 'chloritized brecciated amphibolite', Vogt (1927) referred to it as 'amphibolite', believing it to be a retrogressively metamorphosed sill-like extension of the gabbro, and Kautsky (1953) described parts of it as 'agglomerate' and considered that it represented the junction between his Pieske and Vasten nappes.

The characteristic features of this zone are certainly unusual. The zone locally cuts across the adjacent schists (Fig. 17) and contains a variety of altered and fragmented rock-types; staurolite schists, kyanite schists, hornblende-mica schists and garnet-mica schists occur in all stages of chloritization, and chloritized amphibolites are common. An unusual rock type often present is a white-weathering plagioclase-quartz rock with subordinate biotite, chlorite and epidote, which occurs as discontinuous bands and fragments in a chloritic matrix. This rock-type possibly represents a chloritized trondhjemite, or the last stage in the sequence amphibolite → chloritized amphibolite → chlorite/felspar rock.

The cupriferous pyrite ore bodies are closely associated with the chloritization, and pyrite impregnation is not uncommon at this level. The ore is mainly pyrite with minor chalcopyrite, sphalerite and pyrrhotite, and large porphyro-
blasts of pyrite overgrow the other finer grained sulphides (Carstens 1941), suggesting the influence of the main metamorphism on the ore bodies (Vokes 1968).

Structural history

The large-scale structure of the Sulitjelma region has been well described by Vogt (1927) and will only be summarized here. Essentially there are two major folds which cross to give the rocks their overall saddle-shaped arrangement (J. H. L. Vogt 189) – the N–S-trending Baldoivve Synform and the WNW–ESE-trending Langvann Antiform (Fig. 2). The axial trace of the Baldoivve Synform crops out west of the area investigated, but from Vogt’s (1927) profiles and map it appears to have a steep wester-dipping axial plane. The axial plane of the Langvann Antiform is vertical and crosses the axial trace of the Baldoivve Synform north-west of the present area. The attitude of the schistosity surfaces in Sulitjelma (illustrated diagrammatically in Fig. 2) reflects this major folding. The present outcrop pattern (Fig. 1) is mainly determined by these large-scale structures, which were formed relatively late in the history of the region (D4) and will not be considered further. In the following sections only the syn- and immediately post-metamorphic structures (mainly minor structures) are considered.

![Diagrammatic map summarizing the structures (mainly minor) of that part of the Sulitjelma region investigated by the writer. The location of the axial trace of the D4 structures follows Vogt (1927) with some modification.](image-url)
**Lower Unit**

The metasediments of the Lower Unit all have a well-developed penetrative schistosity defined by the sub-parallel orientation of mica flakes. In the Furulund Group this schistosity is accompanied by a penetrative mica lineation caused by the intersection of mica flakes with the schistosity surfaces (the L-S fabric system of Flinn 1965), but in the eastern Sjönså Group the lineation is not obvious. In the schistose amphibolites at the top of the Furulund Group sub-parallel hornblende prisms define the schistosity-lineation. Hornblende porphyroblasts in the Furulund Schist are often oriented sub-parallel to the mica lineation, although all variations between random and sub-parallel occur, frequently within a single outcrop. In some rocks quartz 'eyes' or 'tails' to garnet porphyroblasts may be present and these lie parallel to the mica lineation. The attitudes of the schistosity surfaces and penetrative lineations are shown diagrammatically in Fig. 2, and Fig. 3 is a stereographic projection of the penetrative lineations recognized and measured in the field.

Several generations of minor (i.e. < 20 m amplitude or wavelength) folding have affected the rocks of the Lower Unit. The earliest observable folds are isoclinal but there has been some dispute as to whether the isoclinal folding should be subdivided. Henley (1968) originally divided the isoclinal folding into two phases – a first phase of folding on SW- to WSW-trending axes (D₁) and a second phase of folding on W-trending axes which was coeval with the development of the penetrative mica lineation (D₂). The main evidence for the separation was the common occurrence in the well-exposed Balmielven stream section of bedding/schistosity intersections associated with isoclinal folds trending WSW, the axial planar schistosity having a well developed mica lineation trending oblique to the bedding/schistosity intersections (i.e. E–W). It was suggested that the WSW-trending folds developed...
Fig. 5. Style of $D_1$ folds in the Lower Unit. Stippled = psammitic beds; unstippled = pelitic and semi-pelitic beds; black = quartz segregations. An axial-planar schistosity is invariably present. A. $F = 4^\circ/267^\circ$; south-east of Langvann. B. $F = 9^\circ/293^\circ$; east of Balmielven. C. $F = 0^\circ/242^\circ$; Balmielven. The penetrative lineation in the axial-planar schistosity plunges at $0^\circ/280^\circ$. D. $F = 0^\circ/277^\circ$; east of Balmielven. E. $F = 0^\circ/257^\circ$; south east of Langvann. F. $F = 9^\circ/242^\circ$; Balmielven.

with an early fine-grained schistose fabric, now only preserved in the cores of garnet porphyroblasts, and that subsequently, during a period of E–W extension of the rocks and a coarsening of the rock fabric, widespread E–W-trending isoclinal folding occurred. However, as rightly pointed out by Wilson (1968) and Nicholson (personal communication) on the basis of the work of Flinn (1962) and Ramsay (1962), and admitted by myself (Henley 1968), during three dimensional progressive deformation there is no necessity for fold axes to parallel the maximum axis of the strain ellipsoid (in this case E–W) and both sets of isoclinal folds could justifiably be considered as coeval. Thus Nicholson & Rutland (1969) and Wilson (1968) distinguish only one phase of isoclinal fold development in Sulitjelma ($D_1$), followed by a phase of flattening parallel to the schistosity and extension parallel to the mica lineation (the $D_{2b}$ phase of Henley 1968) during which some rotation of porphyroblasts occurred.
Fig. 6. Equal area, lower hemisphere projection of $D_{3a}$ fold axes (dots) and poles to axial planes (triangles) in the Furulund Group of the Lower Unit.

Fig. 7. Style of $D_{3a}$ folds in the Furulund Group of the Lower Unit. Stippled = psammitic beds; unstippled = pelitic and semi-pelitic beds. An axial-planar cleavage, which varies from strain-slip to completely penetrative, is generally present in the pelitic beds but is absent in the psammitic beds. The folds are all overturned to the east and in the lower-grade part of the biotite zone near Lomivann kink-bands of the same generation are common (e.g. G.). A. $F = 22^\circ/352^\circ$; west of Lomivann. B. $F = 27^\circ/360^\circ$; west of Lomivann. C. Kink bands; west of Lomivann. D. $F = 35^\circ/352^\circ$; west of Lomivann. E. $F = 34^\circ/280^\circ$; east of Balmielven. F. One of the largest $D_{3a}$ folds in the Furulund Group; south-east of Langvann. G. $F = 19^\circ/028^\circ$; near exit of Lomielven from Lomivann.
I accept therefore that all the isoclinal folds may be coeval and are the earliest structures present in the Lower Unit. The deformation phase during which the isoclinal folds developed is referred to as D₁, following Wilson (1968) and Nicholson & Rutland (1969), and during this time a distinct schistose fabric formed in the rocks. The attitude and style of these earliest folds cannot be determined since the later phase of more homogeneous east-west extension with flattening parallel to the schistosity (D₂) has altered both these features, swinging axes which originally lay at an angle to the east-west direction towards parallelism with it, and compressing originally more open folds into their present isoclinal form. At present, however, the folds have axes which trend mainly E–W (Fig. 4), although as stated previously axes trending WSW–ENE are common in the Balmielven stream section and have been observed elsewhere, while south of Ny Sulitjelma Wilson (1968) has recorded several examples of N-S-trending axes. The style of the D₁ folds, which have the regional schistosity parallel to their axial planes, is illustrated in Fig. 5. An interesting type of fold is illustrated in Fig. 5 C where small second-order folds are superimposed on a large first-order fold, giving a characteristic flame-like appearance. Such folds are present in the Balmielven stream section and have been recorded by Wilson (1968) from the Furulund Schist north of Langvann. They suggest considerable flattening of originally more open folds.

The second phase of deformation (D₂) was of a more homogeneous type than the first. The presently observed L-S fabric of the rocks developed at this time, modifying the D₁ schistosity and folds, and with the penetrative mica lineation representing the principal elongation direction (Z), and the schistosity representing the principal plane of flattening (ZY) of the triaxial strain ellipsoid. Other features which are believed to be genetically related to the D₂ deformation are the elongated pebbles in the stretched conglomerate at the east end of Lomivann (Vogt 1927, Nicholson 1966); the elongated

Fig. 8. Equal area, lower hemisphere projection of D₃b fold axes (dots) and poles to axial planes (triangles) in the eastern Sjönstå Group of the Lower Unit.
Fig. 9. Style of $D_{3b}$ folds in the eastern Sjönstå Group of the Lower Unit. Black = quartz segregations. An axial-planar strain-slip cleavage is only rarely present and quartz segregations are abundant. The folds are all overturned to the north-west. A and B. South-west of Lomivann.

fragments in several horizons of amphibolitic breccia in the schistose amphibolites at the top of the Furulund Group near Ny Sulitjelma (Mason 1966) and in a ‘mud-flake conglomerate’ in the central Furulund Group; the presence of quartz ‘eyes’ or ‘tails’ to garnet porphyroblasts; the sub-parallel orientation of hornblende porphyroblasts; and the elongate-lens form of the pyritic ore bodies. All these features are characterized by an E–W elongation or orientation, parallel to the penetrative mica lineation in the surrounding schists, and generally with flattening parallel to the schistosity.

The stretched conglomerate at the east end of Lomivann is perhaps the most spectacular illustration of the flattening and E-W elongation characteris-
tic of $D_2$. This rock type has been described and illustrated by Vogt (1927, p. 60 and pls. XIII and XIV) who quotes a pebble with dimensions $50 \times 15 \times 5$ cms. Nicholson (1966), in a redescription of the east Lomivann area, favours an agglomeratic origin for this rock-type and describes several other horizons of fragmented rocks, the fragments also being oriented E-W.

Although the $D_2$ mica lineation trends approximately E-W in most of the Lower Unit rocks there is a gradual change in the orientation to NW-SE on nearing the gabbro (Vogt 1952; Henley 1968; Wilson 1968). It was suggested (Henley 1968) that this change was due to the presence of the solid competent mass of gabbro during the $D_2$ extension/flattening, causing a local change in the flow pattern of the surrounding more ductile metasediments, and this is the view of Wilson (1968).

Other features of the Lower Unit rocks which are probably related to $D_2$, but which I have not studied, are the boudinage of amphibole-rich beds and of quartz veins and segregations in the Furulund Schist described by Wilson (1968).

Following the $D_2$ deformation, minor open folding of several attitudes and styles affected different parts of the Lower Unit, with little or no areal overlap, and thus with no means of deciding their time relations apart from the fact that they are post-$D_2$. They are thus jointly referred to as $D_3$ and subdivided in $D_{3a}$, $D_{3b}$ and $D_{3c}$. The $D_{3a}$ folds are confined to the lower half of the Furulund Group – approximately east of the garnet/hornblende isograd. The folds are open, and may be described as stacked (i.e. half a dozen or so folds of the same size and sense of overturn occurring together (Nicholson 1968)), with approximately N-S-trending axes and a well-developed axial-planar strain-slip cleavage in the more pelitic beds which is, however, absent in the psammitic beds. The axial planes dip approximately NW and Fig. 6 is a stereographic projection of the axial planes and fold axes. The style of the folds is illustrated in Fig. 7.

These $D_{3a}$ folds are absent in the upper part of the adjacent eastern

![Fig. 10. Equal area, lower hemisphere projection of $D_{3c}$ fold axes (dots) and poles to axial planes (triangles) in the Furulund Group of the Lower Unit.](image)
Fig. 11. Style of $D_{3c}$ folds in the Furulund Group of the Lower Unit. Stippled = psammitic beds; unstippled = pelitic and semi-pelitic beds; black = quartz segregations. An axial-planar cleavage, which varies from strain-slip to completely penetrative, is sometimes present in the pelitic beds but is absent in the psammitic beds. A. $F = 12^\circ/127^\circ$; near Furuhaugen. B. $F = 7^\circ/315^\circ$; south of Furuhaugen. C. $F = 7^\circ/265^\circ$; early isoclines refolded by $D_{3c}$ folds; south of Furuhaugen. D. $F = 3^\circ/285^\circ$; Sagmoen. E. $F = 9^\circ/294^\circ$; Furuhaugen. F. $F = 0^\circ/268^\circ$; western end of Langvann. G, H and I. Small folds associated with F.

Sjönstå Group, where they are replaced by NW-SE trending open folds with SW-dipping axial planes, referred to here as $D_{3b}$. An axial-planar strain-slip cleavage is sometimes present and quartz segregations are often found along the axial planes. A stereographic projection of the $D_{3b}$ fold axes and axial planes is shown in Fig. 8 and the style is illustrated in Fig. 9. The relation of
D₃ₐ folds to D₃₄ folds is not clear but it seems possible that the folds represent the two parts of a conjugate fold system, and this is supported by the fact that where D₃ₐ and D₃₄ overlap, the D₃₄ folding is always present as crenulations on the shorter limbs of the D₃ₐ folds (Nicholson 1968).

D₃₆ folds are limited to the western part of the Furulund Group. They vary from crenulations of the schistosity to folds up to 20 cm in amplitude, and are open, with axes oriented approximately E-W. An axial-planar strain-slip cleavage is often present and the axial planes generally dip to the south. A stereographic projection of the fold axes and axial planes is shown in Fig. 10 and the style of the folds is illustrated in Fig. 11.

**Upper Unit**

As in the Lower Unit, the metasediments of the Upper Unit all have a well-developed penetrative schistosity/mica lineation. In that part of the Baldoaivve Group studied by the writer (the southern part) the lineation trends E-W, parallel to the lineation in the underlying Furulund Group (Fig. 12). It should be noted, however, that this almost ubiquitous presence of a penetrative lineation in the Upper Unit rocks is in contradiction with Kautsky’s description of the Gasak nappe rocks as having no lineation (Kautsky 1953, p. 223).

Folding is not common in the southern Baldoaivve Group, but two phases of minor folding have been recognized – an early phase of tight folds with axial planes parallel to the regional schistosity (Fig. 13 A-C) and a later phase of folds which vary from tight to open and which deform the schistosity (Fig. 13 D-F).

The earliest folds, referred to the D₁ deformation episode, are well exposed in the Villumelven stream section and are of a ‘similar’ style. No truly isoclinal folds have been recognized in the southern Baldoaivve Group, although Wilson (1968) has recorded some in the northern Baldoaivve Group. The well-developed axial-planar schistosity is identical with the regional schistosity and granitic boudins, often in strings, frequently occur along the axial planes.
Fig. 13. Style of structures in the southern Baldoivve Group of the Upper Unit. Black = quartz segregations; plus signs = granitic material. A. \( D_1 \) fold with axial-planar schistosity; calcareous schist; Villumelven. B. Boudin of pegmatite parallel to the axial-planar schistosity of a tight \( D_1 \) fold. \( F = 12^\circ/270^\circ \); Villumelven. C. A similar occurrence adjacent to B. D. \( D_3 \) fold with shearing and quartz segregation along axial plane; calcareous schist; near Villumelven. E. \( D_3 \) fold in calcareous schist. \( F = 29^\circ/270^\circ \); near Villumelven. F. \( D_1 \) fold with sheared-off limb in calcareous schist. \( F = 5^\circ/259^\circ \); north of Villumvn.

(Fig. 13 B, C), implying that during \( D_1 \) a well developed schistose fabric had been developed. The folds have predominantly E-W axes (Fig. 14) in the southern Baldoivve Group, but in the northern Baldoivve Group Wilson (1968) has recorded a wide scatter of early fold axes, many of which are associated with the early N-S-trending large-scale Duoldagop Synform
Fig. 14. Equal area, lower hemisphere projection of $D_1$ fold axes in the southern Baldoaivve Group of the Upper Unit.

(Mason 1966, Wilson 1968). Wilson attributes much of the fold axis scatter to later large-scale open folding.

The second, more homogeneous, deformation episode ($D_2$) involved flattening parallel to the schistosity and extension parallel to the mica lineation (E-W), as in the Lower Unit. The evidence for this phase is less clear-cut than in the case of the Lower Unit, where extension features such as elongate fragments are common, but is based on the ubiquitous development of the penetrative mica lineation in virtually all rocks except the massive gabbro; the sub-parallel E-W orientation of staurolite, kyanite and hornblende porphyroblasts which post-date the $D_1$ schistose fabric, and the compression of the schistosity around these porphyroblasts; the boudinage of granite veins referred to above; and the development of an intense mineral lineation fabric in the flaser gabbro (Mason 1967) and Furulund Granite (Wilson 1968) adjacent to the gabbro, both of which were intruded after $D_1$.

The evidence from the flaser gabbro and nearby eastern bodies of Furulund Granite is especially interesting since here there appears to have been much more strain than in the rest of the Upper Unit. Mason (1967) suggested that the lineation of the flaser gabbro was due to thrusting, but has since modified this view as he has not found the flaser gabbro elsewhere round the base of the gabbro, the contact on the eastern margin being magmatic (Mason, personal communication). Wilson (1968) has shown that the intensity of lineation in the eastern bodies of Furulund Granite is much greater than in the western body, and attributes this to unusually high strain suffered by rocks adjacent to the relatively undeformed gabbro, the cause, suggested by Henley (1968), of the intense lineation in the flaser gabbro.

Minor post-schistosity folding, referred to the $D_3$ deformation episode, is rare in the southern Baldoaivve Group. The folding is generally on E-W axes and may be open or tight, with some dislocation along the axial planes as shown in Fig. 13. In the northern Baldoaivve Group Wilson (1968) has recorded several types of post-schistosity minor folds which he tentatively
suggests are associated with a late, large-scale E-W-trending synform that has folded the gabbro and country rock and produced the 'eyed' interference pattern of outcrop in the Duoldagop area.

**Junction Unit**

Two varieties of schistose fabric are present in the rocks of the Junction Unit – the often partially chloritized penetrative D₁-D₂ schistosity/lineation of the constituent rocks, and a 'secondary schistosity' produced by laminae and felts of chlorite and mica between rounded fragments of such rocks, the appearance of this 'secondary schistosity' thus being rather nobby and irregular.

The orientation of the D₂ mica lineation in the constituent rocks is parallel to that in the adjacent Upper and Lower Units (i.e. E-W) but in the more brecciated parts a slickenside type of lineation is often present on the surfaces of the individual fragments. This lineation varies in trend from SW-NE to NW-SE (Fig. 15). The micaceous surfaces of the 'secondary schistosity', however, show no mineral lineation; neither do the fragments between the mica felts show any preferred orientation.

Folding in the Junction Unit is of two types, which are probably genetically related and which are confined to this level. Minor folds and crenulations with N-S-trending axes and variably dipping axial planes (Fig. 16) are frequent in the more schistose chloritized beds of the Junction Unit. These folds are open and vary in amplitude from mere crenulations of the schistosity to folds with amplitudes of two or three metres (Fig. 17). They occur in the chloritized rocks and also in the immediately underlying Furulund Schist. Larger scale N-S-trending folds with steeply dipping axial planes occur sporadically along the Junction Unit from Sagmoen westwards. Such folds

**Fig. 15.** Equal area, lower hemisphere projection of slickenside lineations in the Junction Unit.

**Fig. 16.** Equal area, lower hemisphere projection of minor folds (dots) and poles to axial planes (triangles) in the Junction Unit.
Fig. 17. Style of structures in and near the Junction Unit. Circles = tectonic breccia; plus signs = Furulund Granite. A. Infold of schists into tectonic breccia, cutting out of beds against tectonic breccia, and folding on N–S axes; east of Villumelven. B. Beds of white quartzite (stippled) and garnetiferous schists cut out against tectonic breccia; west of Villumelven. C. Asymmetric N–S trending downwarp of tectonic breccia and underlying Furulund Schist; the overlying Furulund Granite sill is unaffected; northwest of Sagmoen. D and E. Minor N–S folds in Junction Unit between Jakobsbakken and Sagmoen. F. $F = 25^\circ/358^\circ$; garnetiferous schist above ore horizon east of Ny Sultjelma. G. N–S minor fold with incipient axial-planar strain-slip cleavage; west of Villumelven.
(e.g. Fig. 17 C) are characteristically disharmonic, warping down the chloritized schists of the Junction Unit several metres into the underlying Furulund Schist without disturbing the overlying schists and Furulund Granite of the Upper Unit. Wilson (1968) has described similar unusual folds of the Junction Unit in the Bursi area north of Langvann, where he also records three strips of Furulund Schist infolded with the tectonic breccia.

The age relations of the structural features of the Junction Unit are not clear. The brecciation and development of the slickenside lineations and secondary schistosity are certainly post-D2 (referred to the Upper and Lower Units), as is the N-S minor folding. The Junction Unit is folded by the large-scale D4 Baldoaivve Synform and thus originated sometime between D2 and D4, but precise correlation with D3 of the Upper or Lower Units is not possible.

The tectonic significance of the structural features of the Junction Unit is also not clear. Overall, the features do not suggest extensive post-metamorphic thrusting, although some local dislocation has obviously occurred. To the north-west of the Sulitjelma region, however, the tectonic breccia is not present (Nicholson and Rutland 1969) and even within the region there is rapid thinning of the tectonic breccia southwards and eastwards.

**Metamorphic history**

*The metamorphic zones*

Detailed field work and laboratory study of the mineralogy of the metasediments and metadolerites of the Sulitjelma region has led to some revision of Vogt's (1927) original metamorphic zonal sequence and clarified the recent controversy (e.g. Strand in 'Geology of Norway', 1960, p. 255) on the nature of the isograds and whether they correspond to nappe junctions.

Vogt (1927) described a sequence of zones from east to west as follows: (1) chlorite zone; (2) biotite zone; (3) garnet zone; and (4) oligoclase zone. As a result of the recent work in the region it has been established that the biotite isograd has no validity in the area mapped, biotite having been found sparsely in the metasediments all round the western end of Lomivann by the writer, and in the metasediments and metavolcanics at the eastern end of Lomivann by Nicholson (1967 and personal communication). However, although no sharp biotite isograd can be drawn, the sparsity of biotite and abundance of chlorite suggest that this part of the region probably corresponds in grade to the lower biotite zone. It must be noted that Vogt realized the arbitrary nature of his biotite isograd, for he wrote (1927, p. 202) 'The boundary between these schists [viz. biotite schists which have been retrogressively altered to chlorite-muscovite schists but still containing relics of biotite] and the chlorite-muscovite schists which have never carried biotite is, however, very difficult to locate, and the line drawn must therefore be interpreted as purely conjectural' (translation; my insertion in brackets).
Compositionally, the biotites from the various metasedimentary and metabasic rocks are fairly similar, with the ratio $\frac{Fe}{Fe+Mg}$ mainly between 0.35 and 0.45. Comparison of biotite composition with host-rock composition has shown that the ratio in biotite is mainly determined by the ratio of the host-rock rather than by the metamorphic grade. There is, however, an irregular decrease in the Mn content of biotite with increasing grade in garnetiferous schists due to the removal of Mn into the increased amounts of garnet at higher grade (Henley 1968).

West of the biotite isograd Vogt mapped a garnet (almandine) isograd, which over the central part of the area investigated is fairly accurate. However, in the north-east of the area garnetiferous schists have been found immediately beneath the schistose amphibolites at the top of the Furulund Group, extending eastwards at least as far as the eastern end of Lomivann (Nicholson 1966), and the revised isograd is shown in Fig. 18.

Although Vogt recorded the incoming of hornblende in the metasediments soon after garnet, he did not map a separate isograd. However, in mapping the garnet isograd in detail it became clear that there was a definite hornblende isograd – essentially coincident with the garnet isograd – and the joint line of first appearance is referred to hereafter as the garnet/hornblende isograd. Thus, at the garnet/hornblende isograd schists with garnet, garnet plus hornblende, or hornblende alone, first appear and are common throughout the higher grade area to the west.

The garnets which first form are rich in almandine but contain significant amounts of the spessartine and pyrope components. Virtually all the garnets analysed show compositional zoning from core to margin – the ‘cryptic zoning’ of Harte & Henley (1966). With increasing metamorphic grade there is, with some exceptions, a gradual decrease in the spessartine content of the garnet margins from 14–17% at the garnet/hornblende isograd to 2–6% at high grades. This decrease is due to increasing amounts of garnet formed at the higher grades, rather than any stability limitation of garnet composition. In addition to these almandine-rich garnets, a single occurrence of a bed of schist with spessartine- and grossular-rich garnet (25.7% spessartine, 40.5% grossular, 33.8% almandine) has been recorded from the lower biotite zone in the SW corner of Lomivann (Henley 1968) and the probable continuation has been noted by Nicholson (1966) in the SE corner of Lomivann.

The amphiboles which first form in the metasediments are typical common hornblendes containing over 15% $Al_2O_3$. There is significant change in hornblende composition with increasing metamorphic grade, the main determinant of hornblende composition being the composition of the host-rock. No actinolite has been observed in the metasediments, although it occurs commonly in the metadolerites in the biotite zone, often mantled by, or intergrown with, hornblende.

The evidence bearing on the oligoclase isograd, which was drawn by Vogt...
Fig. 18. Summary of the variation in metamorphic grade across the Sulitjelma region according to Vogt (1927) and the writer. The occurrences of staurolite and kyanite are from Vogt (1927), Wilson (1968) and Henley (1968). Plagioclase compositions in the schistose amphibolites, metaporphryitic amphibolite and metagabbro are from Mason (1966); other determinations are from Henley (1968).
at a considerable distance to the west of his garnet isograd, is both direct and circumstantial. The direct evidence is that oligoclase has been found in the metasediments as far east as the SE end of Langvann and that no albite is present west of this occurrence (Fig. 18). Unfortunately there is very little identifiable plagioclase in the metasediments and the number of determinations is not large. In particular, the only observed occurrence of albite is from the SW corner of Lomivann, and thus, on the evidence from the metasediments alone, the true oligoclase isograd must lie somewhere between here and the SE end of Langvann.

The indirect evidence, however, enables the likely position of the isograd to be defined more accurately. This evidence is the variation in anorthite content of plagioclase in the metadolerite sills and lenses for which many more data are available (Fig. 18). In these metadolerites the plagioclase changes from $An_{1.5}$ in the lower biotite zone, through $An_{12.15}$ at the garnet/hornblende isograd, to $An_{23.30}$ west of the garnet/hornblende isograd. There is a marked absence of plagioclase in the composition range $An_{15.28}$, a feature which is similar in character to that noted from other areas (e.g. Ambrose 1936, Card 1964, Lyons 1955) and which has been discussed by Brown (1962) in terms of the peristerite solvus. West of the garnet/hornblende isograd the plagioclase in the metadolerites shows no consistent increase in anorthite content but varies between $An_{23}$ and $An_{30}$, whereas in the metasediments the plagioclase composition varies between $An_{23}$ and $An_{42}$.

On the basis of the work of De Waard (1959) and others, which has established that the sharp change in composition of plagioclase from albite or sodic oligoclase to calcic oligoclase occurs at the same metamorphic grade in metasediments as in metabasic rocks, the evidence from the Sulitjelma metadolerites suggests that the (calcic) oligoclase isograd in the metasediments corresponds approximately to the garnet/hornblende isograd. Confirmation of this is provided by Vogt himself who wrote (1927, p. 195) 'The transition to the next zone with oligoclase is unsharp, or more correctly, difficult to define. Comparatively soon after the appearance of hornblende the first very small grains of oligoclase appear. With increasing metamorphism the grain size of the plagioclase increases and it can often be questionable where to set the boundary' (translation).

In the investigation of the Sulitjelma plagioclases both optical (universal stage) and X-ray methods (Smith 1956) of determination were used. It was established for 10 different plagioclases from the metadolerites (of various metamorphic grades) for which both methods were used that the plagioclase was in the low-temperature state, and hence it may be assumed that all the plagioclase in both the metasediments and metadolerites is in this state. The optical measurements involved the determination of the twin-law and composition of at least three separate grains within a thin-section. In both metasediments and metadolerites the twinning is almost exclusively simple – albite twinning in albritic plagioclase and predominantly pericline (or acline?) twinning in oligoclase and andesine. The anorthite contents of the different
Fig. 19. Textures in garnet-mica schists of the Lower Unit. Unless otherwise stated the schistosity shown (S_e) is the composite D_1-D_2 regional schistosity, and sections were cut perpendicular to the D_2 penetrative lineation.

64K-140. Euhedral MnO- and CaO-rich garnet in schist in the lower biotite zone; western end of Lomivann.

63K-171. Elongate garnets in thin graphitic schist immediately underlying schistose amphibolites north of Lomivann.

63K-206. Garnets with S_i at a marked angle to S_e; section cut perpendicular to D_1 fold and showing the junction between a pelitic and a semi-pelitic bed; Balmielven.

64K-176. Garnets with cores containing S-shaped S_i at marked and varying angles to S_e.
grains within a single thin-section are not significantly different from each other, suggesting the attainment of local equilibrium, and this is also indicated by the absence of zoning in the plagioclase of the metasediments and the scarcity of zoning in the plagioclase of the metadolerites (slight, normal, or reversed zoning is occasionally present).

West of the garnet/hornblende isograd the rocks of the Lower Unit cannot be further subdivided into metamorphic zones, assemblages with garnet, hornblende, or garnet plus hornblende being stable and ubiquitous in the calcareous schists to the western limit of the area studied. Thus, in terms of the assemblages in aluminous schists, it is uncertain what is the maximum grade reached in the Lower Unit.

The Upper Unit, however, contains aluminous schists, enabling the metamorphic grade to be more clearly defined than in the Lower Unit. At the base of the southern Baldoaivve Group is a thin bed of staurolite-garnet-mica schist, while in the graphitic mica schist at the top of the Group Vogt (1927, p. 217) has recorded the occurrence of a kyanite-staurolite-garnet-biotite assemblage. In the northern Baldoaivve Group schists with staurolite, kyanite, or staurolite plus kyanite are more common and define the metamorphic grade unequivocally.

There is no evidence to suggest that there is a sharp change in metamorphic grade on passing from the Lower Unit to the Upper Unit, such as might be expected if the two Units were separated by a post-metamorphic thrust as held by Kautsky. The absence of staurolite- and kyanite-bearing schists in the Lower Unit cannot be attributed to a grade effect due to the absence of rocks of suitable composition, while the evidence, albeit somewhat controversial, of the composition of plagioclase in equilibrium with clinozoisite and of the MnO contents of the garnet margins, indicates identical metamorphic grades immediately either side of the Junction Unit.

One notable feature of the metamorphic zoning to which attention was drawn by Vogt (1927, pp. 196–197) is the apparent inversion of the zones. The writer concurs with Vogt in this interpretation since north and north-west of Lomivann higher grade rocks (with garnet or oligoclase) definitely overlie lower grade rocks (with biotite or albite), the garnet isograd dipping gently northwards beneath the gabbro approximately parallel to the schistosity.

and with margins which cut across $S_e$ and contain few or no inclusions; quartz 'eyes' common; north of Langvann.
64K-290. Garnets with S-shaped $S_1$ at a marked angle to $S_e$; garnet/hornblende isograd south of Lomielven.
64K-159. Euhedral garnets with radial inclusion arrangement analogous to the 'cross' in chiastolite; from schist band interbedded with schistose amphibolites north-east of Ny Sulitjelma.
64K-325. Garnets with two apparent phases of growth (cf. 64K-176); south-east of Furuhaugen.
64K-121. Garnets with incipient quartz 'eyes' and $S_1$ at a marked angle to $S_e$; garnet/hornblende isograd south of Lomielven.
Fig. 20. Textures in garnet-hornblende-mica schists of the Lower Unit. Unless otherwise stated the schistosity shown (S₀) is the composite D₁-D₂ schistosity, and sections were cut perpendicular to the penetrative D₂ mineral lineation. The thick lines denote garnet; the medium lines denote hornblende.

63K-97a. Section cut parallel to the D₂ lineation and perpendicular to the schistosity. The garnets contain S-shaped S₁ at varying angles to S₀ and the hornblendes lie approximately parallel to the schistosity; north of Furuhaugen.

64K-97b. The same rock but cut perpendicular to the D₂ lineation. The garnets still contain S-shaped S₁ which, together with the evidence from 63K-97a, indicates that the axes of garnet rotation do not lie parallel or perpendicular to the D₂ lineation.
Nicholson & Rutland (1969) have suggested that the metamorphic inversion was caused by the thrusting of a hot upper nappe over a lower colder one, but another possibility, suggested by Vogt (1967, p. 196), is that the inversion was caused by the presence of the hot mass of gabbro which was intruded into the Upper Unit near the peak of metamorphism, providing extra heat to raise the temperature of the underlying rocks to above that due to the regional metamorphism alone.

**Metamorphic history of the Lower Unit**

In the Lower Unit metasediments the presence of porphyroblasts of garnet, hornblende, muscovite, and occasionally biotite, often with well-developed inclusion trails (S₁) and quartz 'eyes' or 'tails', and the relation of the porphyroblasts to the external rock fabric (Sₑ) and minor structures, has enabled a detailed sequence of mineral growth to be established with reference to the deformation episodes.

The textural features of the rocks which have enabled such a sequence to be established may be summarized as follows. In sections cut parallel or perpendicular to the Dₑ mica lineation garnet porphyroblasts show various textures (Figs. 19 and 20). Garnets with one apparent phase of growth may have S-shaped or linear S₁ in continuity with Sₑ near the garnet/hornblende iso­grad but generally at a marked angle to Sₑ elsewhere (Fig. 19, 63K-206, 63K-290 and 64K-121), or may contain no well-developed S₁ (Fig. 19, 64K-140; Fig. 20, 64K-137). When two apparent phases of growth are present the garnet cores often contain fine-grained S-shaped S₁ at marked and varying angles to Sₑ even within one thin-section, surrounded by a margin which cuts slightly across Sₑ and contains a few larger inclusions or no inclusions (Fig. 19, 64K-176 and 64K-325). It is clear that the axis of garnet rotation, if there is a single axis for all garnets in a rock, does not lie parallel or perpendicular to the Dₑ mica lineation and that the rotational movement during garnet growth mainly pre-dates the Dₑ deformation. Wilson (1968) has investigated the orientation of garnet rotation axes in the Furulund Schist, using the proposed model for garnet growth of Powell & Treagus (1967) to establish when the S₁ indicates a section which has been cut parallel to the rotation axis. Wilson's results suggest that in many cases the garnets in any one thin-section have approximately the same rotation axis, and that this differs significantly in orientation from the rotation axes of garnets in

64K-137. Hornblende-rich schist from the western Sjønstå Group. The garnets are sub­to anhedral, often cracked, and have quartz 'eyes'. The hornblende occurs as small prisms defining the schistosity, which is microfolded.

64K-329. Garnet and hornblende porphyroblasts containing linear or slightly flexed S₁ in continuity with Sₑ; near the garnet/hornblende isograd south of Lomielven.

63K-54. Garnets with S-shaped S₁ at marked and varying angles to Sₑ, and with quartz 'eyes'. In the same thin-section are hornblendes which grow across Sₑ with little or no disturbance, clearly indicating that the garnets have nucleated and grown earlier than the hornblendes; near Furuhaugen.
Fig. 21. Textures in hornblende-mica schists of the Lower Unit. Unless otherwise stated the schistosity shown \((S_e)\) is the composite \(D_1-D_2\) schistosity, and sections were cut perpendicular to the penetrative \(D_2\) mineral lineation.

63K-276. Hornblendes with linear \(S_i\) growing across the schistosity with little disturbance. The \(D_2\) deformation has partially rotated some of the crystals and bowed \(S_e\) about them all; west of Kjeldvann.

63K-161. Hornblendes with linear and S-shaped \(S_i\) at varying angles to \(S_e\); south-east of Furuhaugen.


63K-212. Hornblendes wrapped around by the axial-planar strain-slip cleavage of a \(D_{3a}\)
other rocks. However, Wilson did find that the rotation axes of garnets in different rocks generally plunged in the NW quadrant, in directions apparently unrelated to any minor structures in the surrounding rocks. It is suggested here that these garnets grew in the later stages of the \( D_1 \) deformation, rotating on axes which were directly related to the folding, and that subsequent elongation and flattening of the rocks during \( D_2 \) rotated the early fold axes towards the E-W azimuth while leaving the garnet rotation axes unaffected (cf. Ramsay, 1962).

Virtually all the illustrations in Figs. 19 and 20 show the flattening (\( D_2 \)) of \( S_e \) around the garnet porphyroblasts. Quartz 'eyes' or 'tails' may be present around the garnets in some rocks and their orientation (E-W) suggests an origin during \( D_3 \). They are analogous to the pressure shadows which have been observed around pyrite crystals in slate (Harker 1939, p. 156).

As mentioned previously, hornblende porphyroblasts are common in the Lower Unit metasediments, where they generally lie sub-parallel to the schistosity and mica lineation, although all stages between random and sub-parallel arrangement occur, often in adjacent hand specimens. The hornblende porphyroblasts generally grew slightly later than the garnet porphyroblasts and in thin-section typically contain inclusions of quartz, clinozoisite and ilmenite in S-shaped, linear or random arrangement (Figs. 20 and 21) with the schistosity slightly flattened around them. In most cases \( S_i \) are continuous with \( S_e \) and quartz 'eyes' are absent. Sections cut parallel and perpendicular to the \( D_2 \) mica lineation both show hornblende porphyroblasts with S-shaped \( S_i \) and it is clear that in general there is no single rotation axis for all the porphyroblasts in one rock. In some thin-sections hornblende porphyroblasts which have been sectioned parallel to their length show S-shaped \( S_i \), indicating growth during rotation about an axis lying oblique to the Z-crystallographic axis (Fig. 21, 64K-326).

The orientation of the hornblende porphyroblasts sub-parallel to the penetrative mica lineation and the various textural features of the porphyroblasts indicate growth before and during the \( D_2 \) deformation episode, with the prismatic shape of the crystals affecting their response to the flattening and extension of the surrounding rock matrix. Hornblende porphyroblasts which grew across the schistosity, unless lying exactly perpendicular to the plane of flattening (i.e. the schistosity), were rotated towards parallelism with the mica fold. The hornblends contain linear \( S_i \) approximately parallel to the relict regional schistosity within the microlithons. Section cut perpendicular to \( D_{3a} \) fold axis (\( F = 12°/360° \); south of Ny Sulitjelma).

64K-326. Thin-section cut parallel to \( D_2 \) penetrative lineation. The hornblende porphyroblast is oriented approximately parallel to this lineation and has been rotated during growth into its present position by the \( D_2 \) flattening/extension; near Furuhaugen.

64K-289. Hornblends wrapped around by the axial-planar strain-slip cleavage of a \( D_{3c} \) fold. The hornblends contain linear \( S_i \) approximately parallel to the relict regional schistosity within the microlithons. Section cut perpendicular to \( D_{3c} \) fold axis; near Bursi.
lineation/schistosity, and continuing growth during this deformation gave rise to the unusual feature of hornblendes containing S-shaped $S_i$ with axes of rotation oblique to the hornblende length. The variation in hornblende orientation observed in the field was thus due to porphyroblasts nucleating and growing continuously during $D_2$.

Certain textural evidence that, at least in part, hornblende grew later than garnet is provided by 63K-54 in Fig. 20, where garnets, with quartz 'eyes' and containing S-shaped $S_i$ at an angle to $S_e$, occur in the same thin-section as hornblende porphyroblasts which grow across the schistosity with only a slight bowing of the latter (indicating post-hornblende flattening). The difference in time of growth of garnet and hornblende is possibly related to the stabilization of manganiferous almandine at lower temperatures than hornblende, the garnet cores often having relatively high MnO contents (up to 12% MnO).

The $D_3$ deformation clearly post-dates hornblende and garnet growth, as shown by the bending of the strain-slip cleavage which is axial planar to $D_3$ folds around the porphyroblasts (Fig. 21, 63K-212 and 64K-289) and the preservation of $S_i$ in continuity with the immediately adjacent $S_e$ within the strain-slip microlithons, but at high angles to the newly-produced cleavage. This applies to both $D_{3a}$ and $D_{3c}$ folds but the absence of porphyroblasts in the Muorki Schist precludes the use of such textures to establish the age of the $D_{3b}$ folding. However, the $D_{3b}$ folds deform the $D_1$-$D_2$ schistosity, therefore post-dating it, and as previously suggested are probably coeval with $D_{3a}$.

The last porphyroblasts to grow in the Lower Unit were of muscovite, which occurs in 'books' growing randomly across all the minor structures formed during $D_1$ to $D_3$ and thus post-dating them.

Biotite in the Lower Unit metasediments generally forms the schistosity, muscovite not commonly occurring in this form of sub-parallel oriented flakes. However, certain rocks contain porphyroblasts of biotite with planar or S-shaped $S_i$, which in one case were shown to have grown between $D_1$ and $D_2$, and in another case between $D_2$ and $D_3$ and continuing to overlap the start of $D_3$. Biotite was probably stable and crystallized or recrystallized from early $D_1$ to after $D_3$ in time, the evidence for its crystallization during $D_3$ being the presence of an axial planar cleavage to $D_3$ folds composed of newly-recrystallized, unstrained biotite flakes.

Of the other minerals in the Lower Unit metasediments, clinozoisite forms small prisms oriented parallel or sub-parallel to the $D_2$ mica lineation, while quartz, calcite and plagioclase form anhedral, unstrained grains. Sparse ilmenite flakes lie parallel to the schistosity and in some rocks, particularly in the biotite zone, sub-parallel oriented chlorite flakes define the schistosity. Chlorite also occurs as an incipient alteration product of biotite and, much more rarely, as porphyroblasts which grow across the $D_1$-$D_2$ schistosity.

The minerals present in the metadolerites of the Lower Unit are now completely of metamorphic origin. Hornblende, actinolite, saussuritized plagioclase, epidote, sphene and ilmenite characteristically occur intergrown in a
random manner and impart a massive fabric to the rocks. It is clear that the
original dolerite was intruded prior to $D_2$ (and possibly even prior to $D_1$)
and that during $D_2$ the presumably continuous sills and dykes of dolerite were
deformed and metamorphosed, attaining their present lens-like form and
metamorphic mineralogy.

A summary of the relation between mineral growth and deformation in the
Lower Unit metasediments is shown in Fig. 22. Fig. 22 illustrates the varia­
tion in metamorphic grade in the Lower Unit schists, the maximum grade
being attained before and during $D_2$ with the rocks still in biotite grade con­
ditions during $D_3$.

Metamorphic history of the Upper Unit
In the Upper Unit metasediments, porphyroblasts of garnet, hornblende,
staurolite, kyanite and muscovite occur, but in general well-developed $S_i$ are
rare, the inclusions mainly being randomly arranged. Quartz ‘eyes’, which are
not uncommon around the garnets in some schists of the Lower Unit, are rare
in the Upper Unit. Typical Upper Unit textures are illustrated in Fig. 23.

Garnet is not common in the southern Baldoaivve Group, but where it
occurs it generally forms either spongy, anhedral crystals with numerous
inclusions (Fig. 23, 64K-168) or subhedral to euhedral crystals with few or
no inclusions (Fig. 23, 64K-293, 64K-294). In all cases $S_e$, the composite
$D_1$–$D_2$ schistosity, terminates against the garnet with a slight bowing, indi­
cative of flattening after garnet growth. In the staurolite schist at the base of
the southern Baldoaivve Group elongate garnets occur which are wrapped
around by the schistosity (Fig. 23, 64K-256) but which may enclose a bowed
fabric of elongate quartz, indicating garnet growth during flattening. Garnets
with cores containing slightly S-shaped or straight $S_i$ lying at marked angles
Fig. 23. Textures in the Upper Unit schists. Unless otherwise stated the schistosity shown (S_e) is the composite D_1–D_2 schistosity, and sections were cut perpendicular to the penetrative D_2 mineral lineation. The thick lines denote garnet; the medium lines denote hornblende.

64K-293b. Garnet-mica schist cut parallel to the D_2 penetrative lineation and schistosity; Villumelven.

64K-293a. The same rock but cut perpendicular to the D_2 lineation.

64K-168. Spongy garnets, full of large inclusions of quartz and calcite, growing across S_e. Garnet-hornblende-mica schist; north-east corner of Villumvann.


64K-256. Staurolite-garnet-mica schist at base of Upper Unit. The garnets vary from elongate to equant and are wrapped around by S_e; large muscovite pods (Mu) contain small relict cores of staurolite; between Furuhaugen and Sagmoen.

64K-294a. Garnet-mica schist taken from the nose of a D_1 fold. The schistosity (S_e) is axial-planar to the fold and is completely penetrative. The garnets grow across S_e which is slightly bowed around them. Section cut perpendicular to the fold axis; Villumelven.

64K-294b. The same rock, showing an elongate garnet porphyroblast lying parallel to the axial-planar schistosity.

to S_e have been observed (Fig. 23, 64K-293) but are unusual; when two apparent phases of growth are present the cores usually contain randomly oriented fine inclusions and the margins a few large inclusions or no inclusions at all.
Fig. 24. Summary of mineral growth and recrystallization in relation to deformation in the Upper Unit schists. There is some uncertainty about the relations during D1, and also for clinozoisite, plagioclase and ilmenite.

It is clear that the garnets mainly post-date the D1 schistose fabric of the rocks, which has then subsequently been flattened around the solid porphyroblasts during D2, some growth continuing during D2 in certain rocks.

Where D1 folds are present the garnets grow across their axial planar schistosity (Fig. 23, 64K-294), thus post-dating the folds. Indications of rotational movements during garnet growth are exceptional, in contrast to the Lower Unit where they are common.

Hornblende porphyroblasts in the southern Baldoaivve Group are much stumper than in the Lower Unit (length to width 3 or 4 to 1 as compared to 6 to 1 in the Lower Unit), S1 are rare or absent, and the abundant inclusions are randomly orientated. The porphyroblasts are oriented subparallel to the D2 mica lineation and the schistosity is flattened around them (Fig. 23, 63K-115). They thus show similar textural relations to the garnets and are interpreted as having grown after D1 but before D2.

The sparse prismatic porphyroblasts of staurolite and kyanite which have been observed in the southern Baldoaivve Group mainly have similar textural relations and orientation to the hornblendes described above, except that inclusions are not so common. Occasionally, however, staurolite porphyroblasts occur with planar S1 in continuity with a flattened S0, again indicating static growth.

Biotite is the predominant mica in most of the calcareous schists of the Baldoaivve Group, but in the aluminous schists muscovite is common. Both occur as sub-parallel flakes which define the L-S fabric of the rocks. Paragonite has been found in a garnet-hornblende-mica schist at the base of the southern Baldoaivve Group, and its co-existence with muscovite has been used to determine the temperature of recrystallization of the staurolite-grade rock as 550° to 570°C (Henley 1970). In the Furulund Granite sill an
Fig. 25. Textures in the Junction Unit rocks.

64K-255. Chloritized garnet-mica-plagioclase rock showing chlorite flakes (Chl) wrapping around and forming from relict garnet (Ga) porphyroblasts in a matrix of plagioclase, muscovite and chlorite; between Sagmoen and Furuhaugen.

64K-171. Chloritized amphibolite showing hornblende (Hb) altering to chlorite along cracks and cleavages, in a matrix of plagioclase, epidote and chlorite (not distinguished); near Furuhaugen.

63K-248. Chloritized staurolite-mica schist showing chlorite flakes wrapping around a staurolite (St) porphyroblast which has been almost completely altered to a fine-grained aggregate of sericite, chlorite and iron ore; south of Jakobsbakken.

63K-333. Chloritized kyanite schist showing chlorite flakes wrapping around a kyanite porphyroblast (Ky) which has been marginally altered to fine-grained sericite; biotite (Bi) flakes are also present; Jakobsbakken.

L–S fabric is present and this suggests that the Granite was emplaced prior to $D_2$, probably in the static period between $D_1$ and $D_2$, since locally Wilson (1968) has found apophyses of granite from the main body north of Langvann cutting across the surrounding schists. A similar but less well developed L–S fabric is present in the extensive granites and trondhjemites in the upper part of the southern Baldoaivve Group.

In addition to ‘schistose’ muscovite, randomly oriented muscovite porphyroblasts are common in the calcareous schists of southern Baldoaivve Group, where they grow across the schistosity.
Of the other minerals in the Upper Unit metasediments, clinozoisite occurs as small prisms, while quartz, calcite and plagioclase form anhedral unstrained grains. Chlorite is rare in the Upper Unit.

With regard to the meta-igneous rocks of the Upper Unit, the mica fabrics in the Furulund and Baldoaivve Granites have been mentioned previously. The sill-like extension south of Langvann of the Furulund Granite has a flaser structure, with large microcline and oligoclase crystals in a finer-grained, partially comminuted matrix, the large crystals often showing cracking and fracturing. The textures of these granites are believed to have developed during $D_2$, when boudinage of granite veins in the southern Baldoaivve Group probably also occurred. The textures of the flaser gabbro have been described by Mason (1966, 1967), who noted a well-developed hornblende lineation in both the coarse lenses and their surrounding fine-grained amphibolitic matrix and the presence of cataclastic textures. The hornblende lineation lies parallel to the mica lineation in the surrounding schists and probably originated during $D_2$.

A summary of the relations between mineral growth and deformation in the Upper Unit metasediments is shown in Fig. 24, the maximum grade of metamorphism being attained between $D_1$ and $D_2$.

Metamorphic history of the Junction Unit

The constituent rocks of the Junction Unit are similar to rocks in the adjacent Upper and Lower Units except that they are fragmented and chloritized. Typical chloritization textures are illustrated in Fig. 25. The chloritization clearly post-dates the $D_2$ deformation (of both Upper and Lower Units) but it is not certain whether the abundance of chlorite in this zone of dislocation was caused by retrogression under low-grade conditions, or by local influx of water under moderate or even high-grade conditions. The presence of coarse unstrained quartz in the matrix of some of the tectonic breccias perhaps supports the latter case, as does the occasional presence of porphyroblasts of muscovite which grow randomly across minor open N–S folds and crenulations of the chloritic secondary schistosity.

Discussion

Comparison of the lithologies of the Upper and Lower Units indicates that, whereas virtually all the metasediments in the Lower Unit may be described as calcareous schists, aluminous schists with staurolite and kyanite are not uncommon in the Upper Unit, where they occur interbedded with calcareous schists. Granites and trondhjemites are common in the Upper Unit but completely absent in the Lower Unit.

The structural and metamorphic histories of the Upper and Lower Units are closely similar in many respects, at least since before $D_2$ times. Thus in both a schistose fabric was generated during $D_1$ together with folds, which
in the Upper Unit may be described as tight and in the Lower Unit isoclinal. There is evidence to suggest, however, that the D\textsubscript{1} folds in the Lower Unit were much more open at this stage than they appear at present, having been modified severely during the later D\textsubscript{2} deformation. The metamorphic grade increased during D\textsubscript{1} in both Units such that garnet porphyroblasts began to crystallize. In the Lower Unit, growth of these early garnets was coeval with rotational movements which may have been associated with the D\textsubscript{1} folding. S-shaped S\textsubscript{1} in garnet cores, often at marked and varying angles to S\textsubscript{2}, are common in the Lower Unit and Wilson (1968) has established that in many cases garnet rotation axes plunge in the NW quadrant. It is suggested that the axes preserve the original D\textsubscript{1} fold axis orientations, while the fold axes themselves were later rotated into sub-parallelism with the direction of maximum strain during D\textsubscript{2} (mainly E–W). Garnet growth in the Lower Unit continued after D\textsubscript{1} until the closing stages of D\textsubscript{2}, the later garnet often forming margins around cores of the earlier garnet. Hornblende growth in the Lower Unit began between D\textsubscript{1} and D\textsubscript{2} and continued to the end of D\textsubscript{2}, the maximum growth occurring somewhat later than that of the garnet. Again inclusions are abundant, and often in S-shaped or planar arrangement in continuity with S\textsubscript{2}.

In contrast to the garnet and hornblende porphyroblasts in the Lower Unit, those in the Upper Unit rarely have well-developed S\textsubscript{1}, and garnets with few or no inclusions are relatively common. It appears that in most cases the porphyroblasts in the Upper Unit, including staurolite and kyanite, grew statically across the D\textsubscript{1} schistose fabric and, in the case of garnet, extended into D\textsubscript{2}.

There is therefore some evidence to suggest that, in terms of both style of folding and change in metamorphic grade in relation to deformation, the Lower and Upper Units had somewhat different histories during and slightly after D\textsubscript{1}. However, the structural and metamorphic histories of the two Units are indistinguishable from at least D\textsubscript{2} onwards. Thus during D\textsubscript{2} the rocks of both Units underwent flattening parallel to the schistosity and extension in an E–W direction, modifying the D\textsubscript{1} schistosity and giving rise to the presently observed penetrative L–S mica fabric. Locally, near the competent mass of the Sulitjelma gabbro, the strain was higher than elsewhere and caused the development of the intense lineation in the flaser gabbro and eastern Furulund Granite bodies. The strict parallelism of the D\textsubscript{2} linear structures in both Units, even where, such as near the gabbro, these structures lie at a marked angle to the main E–W extension direction, is conclusive evidence that the Units have been in their present relative positions since D\textsubscript{2} times.

Subsequent deformation affected both Units in the same manner. Thus later open minor folds (D\textsubscript{3}), which deform the D\textsubscript{1}–D\textsubscript{2} schistosity and post-date the porphyroblastesis, are found in both Units. However, during D\textsubscript{3}, except perhaps for the present biotite zone in the Lower Unit where chlorite grade conditions may have prevailed, the rocks of both Units were under biotite
grade conditions. Following \( D_9 \), widespread muscovite porphyroblastesis occurred in the calcareous schists of both Units and large-scale folding (\( D_4 \)) of the rocks took place, resulting in the formation of the Baldoaivve Synform and Langvann Antiform.

Analysis of the structural and metamorphic histories of the Upper and Lower Units shows conclusively that, apart from possible local dislocation at the level of the Junction Unit, post-metamorphic thrusting is absent in Sulitjelma. There is no sharp change in metamorphic grade at the suggested level of Kautsky's (1953) nappe junctions and any thrusting must have taken place before \( D_2 \). There is some evidence to suggest that prior to \( D_2 \) the Upper and Lower Units had different structural/metamorphic histories, and although lithological evidence is by no means conclusive, it is, perhaps, significant that granitic rocks and aluminous schists are confined to the Upper Unit.

The pattern of the inverted metamorphic zones also requires explanation, for although there is no sharp change in metamorphic grade between the Units there is a definite increase in grade up the lithological succession north of Lomivann, and it appears that here higher grade Upper Unit rocks overlie lower grade Lower Unit rocks, with the garnet/hornblende isograd lying approximately parallel to the junction but within the Lower Unit. Nicholson & Rutland (1969) interpret this zonal inversion as originating with the thrusting of the hot rocks of the Gasak/Rödingsfjäll nappe over the cooler underlying rocks of the Seve-Köli nappe - a sort of contact metamorphism similar to that recorded by Haller & Kulp (1962) in east Greenland. On the basis of the present investigation such thrusting must have taken place between \( D_1 \) and \( D_2 \), at or near the peak of metamorphism in the Upper Unit, and when the Lower Unit rocks were already undergoing regional metamorphism to garnet grade in the western part of the Sulitjelma region.

However, at least one other explanation is possible for the zonal inversion north of Lomivann, namely that the inversion was caused by long-term accession of heat to the Lower Unit from the overlying Sulitjelma gabbro. The gabbro was intruded between \( D_1 \) and \( D_2 \) into Upper Unit rocks which were already metamorphosed to kyanite grade, since Mason (1966) has recorded xenoliths of staurolite-kyanite schist within the gabbro. The initial heat from the gabbro caused a narrow thermal aureole which was partly destroyed during \( D_2 \), but it seems possible that the long term dissipation of heat from the gabbro after its intrusion might have heated the underlying rocks sufficiently to bring them up to garnet grade. This explanation is an elaboration of Vogt's (1927) explanation of the strip of 'biotite schist' (actually garnetiferous mica schist) immediately underneath the gabbro, and has the merit that it is consistent with the presence of igneous contacts on the eastern margin of the gabbro (Mason, personal communication).

As mentioned in the introduction, in addition to the zonal inversion, Nicholson & Rutland (1969) suggest that the increase in thickness of the rocks between the ore horizon and the Furulund Granite on passing from the southern to the northern Baldoaivve Group, and the lack of correlation of
many of the lithologies of the northern Baldoaivve Group with those of the southern Baldoaivve Group, may be related in some way to their suggested nappe tectonics for the region. In my view this evidence is inconclusive since rapid thickness and facies changes are not uncommon in this part of Nordland (Vogt 1927, Nicholson & Rutland 1969). More conclusive is the evidence of the mass of granite gneiss and granite gneiss breccias north of Sorjusvann at the level of Kautsky's Gasak thrust. Both Kautsky (1953) and Nicholson & Rutland (1969) interpret these rocks as allochthonous Precambrian granitic basement with overlying in situ breccia, and it is difficult to explain them in any other way.

On balance the combined local and regional evidence suggests that there is a syn-metamorphic nappe junction in Sulitjelma as proposed by Nicholson & Rutland (1969), at about the level of the ore horizon, the thrusting having occurred between D₁ and D₂ and any distinctive linear features that may have developed in the thrust zone having been obliterated during D₂. The chloritization and brecciation associated with the Junction Unit may then, as suggested by Nicholson & Rutland, be associated with late-stage local movements along the earlier syn-metamorphic nappe junction, though the reason for such later movements is obscure.

The presence of the pyritic ore bodies at the level of the suggested thrust is unlikely to be fortuitous and it may be that the ore was emplaced along the nappe junction, either from epigenetic solutions or from remobilized syn-genetic sulphides, the nappe junction at this time being a dislocation zone of easy deposition. Following emplacement, metamorphism of the ore bodies occurred and the subsequent elongation and flattening took place during D₂.

Redistribution of elements took place at this time or later by hydrothermal solutions or by direct remobilization (Vokes 1968, p. 58), giving rise to such features as the unusual antimony-rich paragenesis recorded by Ramdohr (1938) from Jakobsbakken and possibly the chloritization so characteristic of the Junction Unit.

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