Structural evolution of the Bergsdalen Nappes, Southwest Norway

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Fossen, H. 1993: Structural evolution of the Bergsdalen Nappes, Southwest Norway. *Nor. geol. unders. Bull. 424*, 23-49.

The Bergsdalen Nappes are Caledonian thrust sheets which experienced pervasive, although heterogeneous ductile deformation during the upper greenschist to lowermost amphibolite-facies, Caledonian, thrusting history. The linear D1 fabric and associated asymmetric meso- and microstructures indicate a top-to-the-ESE sense of shear during thrusting (D1). Little evidence is found for Precambrian ductile deformation in these nappes. However, the penetrative set of fabrics formed during D1 were consistently reworked and overprinted by a later, ductile event. This D2 event caused a slight retrogression of the D1 assemblages, and abundant evidence is found for a non-coaxial D2 history. The sense of vorticity changed by nearly 180° from D1 to D2, so that the Bergsdalen Nappes and the overlying units experienced a top-to-the-WNW translation at this stage. The influence of the post-thrusting D2 event on the Bergsdalen Nappes is found to be considerably higher than previously thought, and can be traced into the basement rocks to the north and west, i.e. the Western Gneiss Region and the Øygarden Complex. Kinematic indicators suggest that Hardangerfjorden and the eastern margin of the Bergen Arc System mark extensional shear zones which penetrate the basement. These shear zones are held responsible for the southeastward rotation of the Bergsdalen Nappes and the Western Gneiss Region in the mapped area, and are therefore the last recorded expression of the ductile D2 deformation, which is related to post-Caledonian, Devonian, extensional tectonics.

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Introduction

In southwestern Norway, two large slivers of basement rocks, named the Bergsdalen Nappes (Kvale 1946), are incorporated in the phyllitic décollement zone between the Western Gneiss Region (parautochthonous basement) and the overlying Jotun Nappe. These nappes were subjected to detailed petrographic and structural analyses in the 1930's and 40's by Kvale (1946, 1948) who mapped pervasive linear and planar fabrics which he related to Caledonian thrusting. The increased understanding of fabrics and structural development since Kvale's work calls for a reinterpretation of this structurally intriguing region. This paper gives a summary of the results obtained from recent mapping and reinterpretation of the structures of the Bergsdalen Nappes and the associated phyllites, and includes a discussion of kinematic observations in the surrounding units.

Previous and present work

A large part of the study area has been mapped by Kvale and other geologists, and the geologic map shown in Plate I includes data from the maps prepared by Kvale (1946), Gray (1978), Kvale & Ingdahl (1985), and Ingdahl et al. (1990), together with the author's own mapping carried out during the summers of 1990-91. New mapping has been concentrated in the western half of the Bergsdalen Nappes where previous mapping is scarce or absent, but all parts of the region have been visited and have been the subject of renewed structural investigation. Similarly, some of the lineation measurements shown in Plate II stem from maps by Gray (1978), Kvale & Ingdahl (1985), and Ingdahl et al. (1990). Their measurements have only been included from areas where the obvious lineation has been found to be the D1 lineation (see below), and serve the purpose of obtaining a better continuity of the linear pattern. However, the majority of the lineations in Plate II are recorded by the present author, and all structural data in the other plates and figures in this article are new data unless otherwise stated.

The Bergsdalen Nappes

The Bergsdalen Nappes are sheets of detached Precambrian rocks which occur in the décollement zone between the Precambrian basement of the Western Gneiss Region and the overlying Caledonian Jotun Nappe and other far-travelled nappes (Fig. 1, Plate I). They form two major tectonic units, the Lower Bergsdalen Nappe (LBN) and the Upper Bergs-

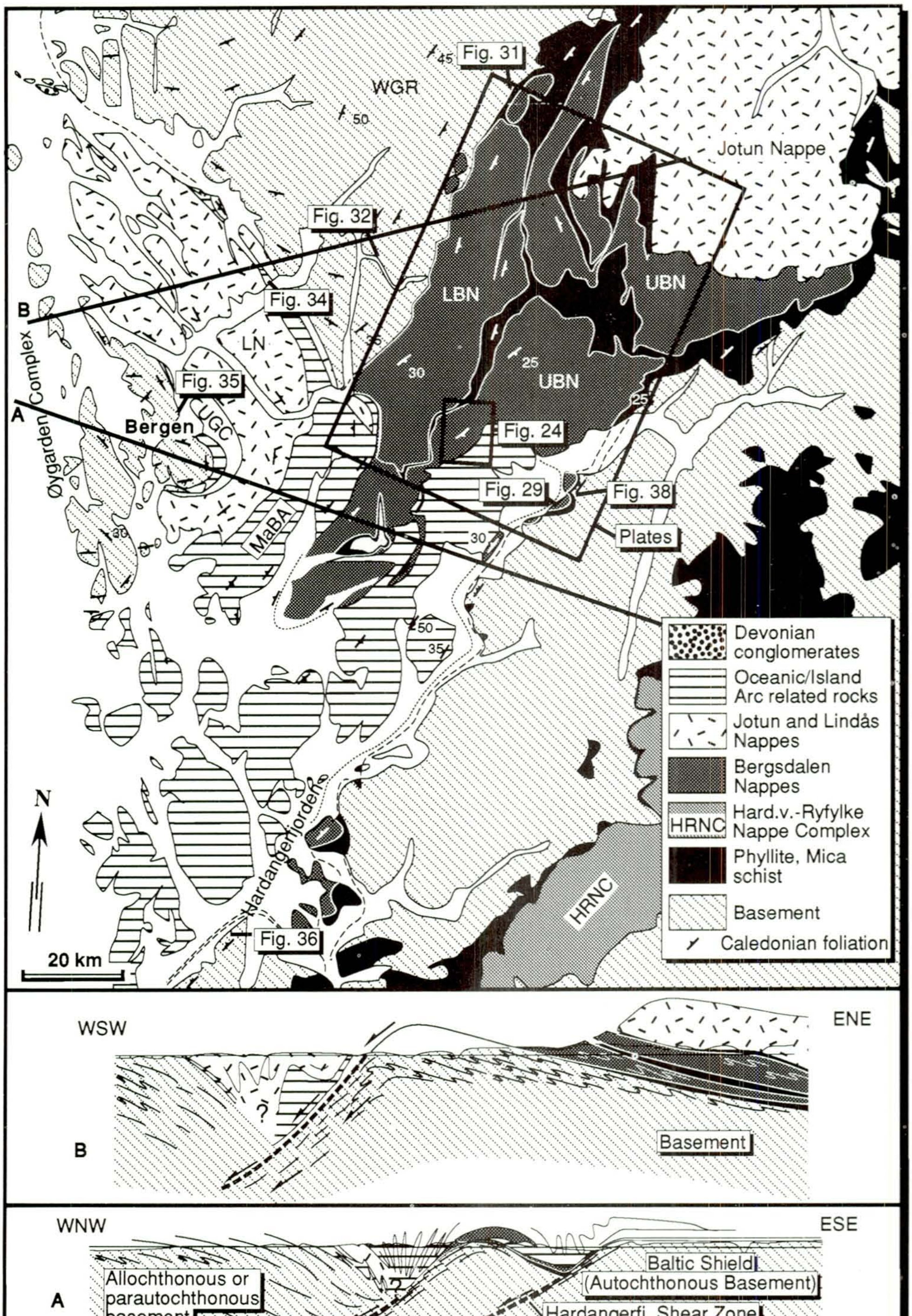

Fig. 1. Geological setting of the Bergsdalen Nappes in southwestern Norway. The areas covered by Plate I and Fig. 24 are outlined, and some of the figure locations are indicated. LBN=Lower Bergsdalen nappe, LN=Lindås nappe, MaBA= Major Bergen Arc, UGC=Ulriken Gneiss Complex, UBN=Upper Bergsdalen nappe, WGR=Western Gneiss Region.

dalen Nappe (UBN), which are separated by an almost continuous zone of phyllite. There are also internal discontinuities of tectonic character which have been interpreted as thrusts (Kvale & Ingdahl 1985), but these appear to be associated with relatively small displacements. At present, the Bergsdalen Nappes are bounded to the southeast by the extensional Hardangerfjord Shear Zone (Fossen 1992) or Faltungsgraben (Goldschmidt 1912), to the southwest by the partly ophiolite and island arc-related rocks of the Hardangerfjord Group (Færseth 1982) and the Bergen Arc System (Kolderup & Kolderup 1940, Færseth et al. 1977, Fossen 1989), and to the north by the Western Gneiss Region. To the northeast the Bergsdalen Nappes taper out in the phyllites beneath the Jotun Nappe.

The Upper and Lower Bergsdalen Nappes are composed of very similar lithologies, and also seem to have a common pre-Caledonian history. Essentially, a supracrustal sequence consisting of basic to intermediate metavolcanites, metarhyolites, quartzite, conglomerates, quartz schists, and quartz-mica schists were intruded by gabbroic magma and, finally, by a number of granitic bodies and associated granite dikes. Several of the granitic bodies have been dated by the Rb-Sr method, and give whole-rock ages between 1274 and 953 Ma (Pringle et al. 1975, Gray 1978). In addition, the metarhyolite has been dated at 1219 ± 111 Ma in the Torfinnsvatnet area southwest of Voss (Gray 1978), suggesting a close genetic relationship between the rhyolite and the granite. Hence, all the rocks in the Bergsdalen Nappes are Precambrian in age. However, the intercalated phyllites and mica schists are more likely to be at least partly Lower Paleozoic in age (Kvale 1946, 1960).

Table 1. Rb/Sr whole rock analyses from the Bergsdalen Nappes

Hodnaberg granite, Hamlagrøv., UBN (Brueckner 1972)	1004 ± 90 Ma, IR=0.7068
Bukkafjellet granite, Bergsdalen, LBN (Pringle et al. unpubl.)	953 ± 16 Ma, IR=0.7143
Evanger granite, Bergsdalen, LBN (Pringle et al. 1975)	1274 ± 48 Ma, IR=0.7637
Metarhyolite, Torfinnsv., UBN (Gray 1978)	1219 ± 111 Ma, IR=0.7381
Fosse granite sheet, Eksingedalen, LBN (Gray 1978)	971 ± 71 Ma, IR=0.7084

Detailed petrographic descriptions of the various lithologies are given in Kvale (1946), Gray (1978) and Ragnhildstveit (1987), and will not be repeated here. It may be noted, however, that the most obvious lithological difference between the UBN and the LBN is that there are more quartz(-mica) schists and less quartzite in the UBN. The many similarities suggest that the two were detached from the same part of the pre-Caledonian basement, probably somewhere near the Cambro-Ordovician margin of Baltoscandia. The supracrustal series in the Bergsdalen Nappes has been correlated with the Proterozoic Telemark Group (Kvale 1946), which is preserved as an infolded and integral part of the autochthonous Baltoscandian Shield to the southeast (Dons 1960). However, the Proterozoic structural development of the rocks of the Bergsdalen Nappes seems far simpler than that suggested for the Telemark Group (see below).

Age of deformation

Since the rocks of the Bergsdalen Nappes are Proterozoic in age, the deformation structures may be either Proterozoic or Paleozoic or both. Multiple Proterozoic deformation structures occur in the nearby Ullensvang Group in the autochthonous basement (Torske 1982). The Ullensvang Group contains metasedimentary and metavolcanic rocks similar to those of the Bergsdalen Nappes. However, the deformation of the Ullensvang Group has apparently not affected the similar, supracrustal rocks of the Bergsdalen Nappes to any significant degree. Numerous Proterozoic granitic dikes and heterogeneous strain makes it possible to draw this conclusion. Where deformation structures such as lineations, folds, and planar fabrics are developed, the granitic dikes appear to have experienced the same deformation as does the host rock. Although they have been carefully searched for, undeformed granitic dikes that crosscut ductile deformation structures have not been found in the study area. Where deformation in the host rock is strong, the granite dikes are transposed and foliated. However, in low strain areas, weakly deformed granitic dikes are found to crosscut layering and sedimentary structures such as cross-bedding (Fig. 2). In contrast, the Proterozoic deformation structures in the Ullensvang Group are normally cut by granitoid dikes. It should also be noted that the deformation in the Ullensvang Group appears to be related to forceful intrusion of granitic plutons (Torske 1982) and may therefore be of relatively local extent. The only unequivocal evidence so far

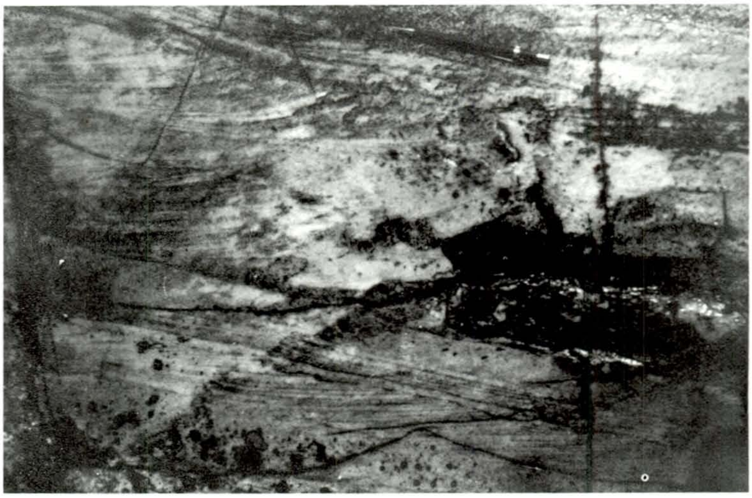

Fig. 2. Primary cross bedding in quartzite south of Holmavatn, LBN (UTM coordinates LN327053). Note pen for scale (14cm).

Fig. 3. Quartz vein system cut by a Precambrian granitic dike from the Bukkafjell Granite. Locality north of Svartavatn/Holmavatn, LBN (LN336077). Lense cap is 6.5 cm in diameter.

for Proterozoic deformation in the supracrustal rocks of the Bergsdalen Nappes is local zones of *en echelon* arranged sigmoidal quartz veins cut by granitic dikes (Fig. 3). The penetrative set of planar and linear fabrics, which also affects the Proterozoic vein systems, are thus likely to be either Caledonian, as inferred by Kvale (1948), or post-Caledonian.

The phyllites

Intercalated with the Bergsdalen Nappes are dark phyllites and mica schists which have experienced intense deformation, and primary structures are obscured. These rocks vary from dark, grayish phyllites with multiple cleavages to mica schists, both typically with milky-white quartz pods and rods (Fig. 4). In places, more calcareous or quartzitic layers are found within the phyllites, probably reflecting a primary lithological stratification. The phyllites and mica schists can be traced continuously eastward under the nappe stack to the fossiliferous, Lower Paleozoic foreland sediments deposited unconformably on the Baltoscandian Shield (e.g. Hossack & Cooper 1986). It is therefore reasonable to assume that at least parts of these phyllites and mica schists are Lower Paleozoic in age (Kvale 1946), although no fossils have been found in these highly deformed rocks to support this assumption. Work in the Hardanger-Ryfylke area (Riis 1977, Solli et al. 1978) has indicated the presence of Precambrian as well as Lower Paleozoic phyllites and mica schists, based on Rb/Sr age determinations of interbedded, Precambrian, gneissic meta-andesites. Similar gneisses have not been found in the phyllites embracing the Bergsdalen Nappes, except for gneisses apparently detached from the Bergsdalen Nappes, and the volume of Precambrian rocks in the phyllite zone is therefore unknown.

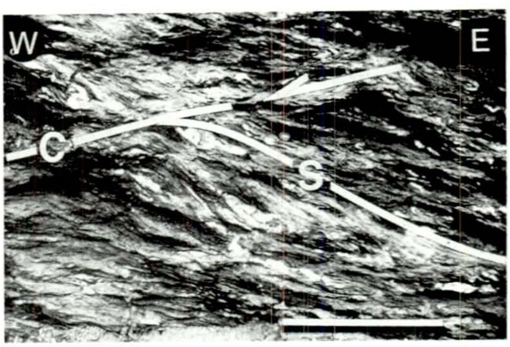

Fig. 4. Typical phyllite near Bulken, west of Voss (LN505251). Note F2-folded quartz veins and the foliation (S) which is affected by gently W-dipping shear bands (C) formed during D2 shearing. Looking NNE. Scale bar = 1 meter.

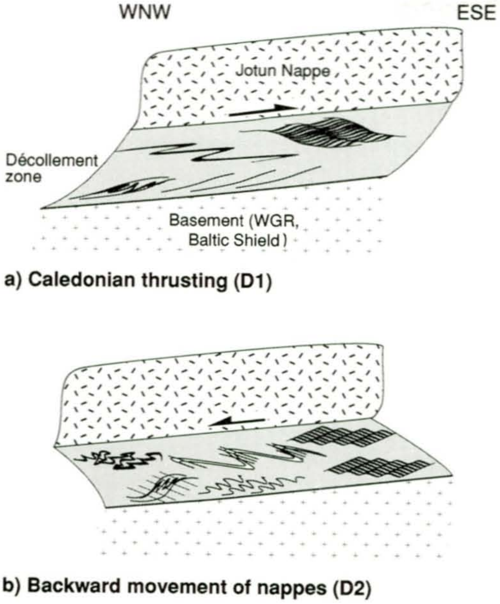

Fig. 5. Schematic illustration of the two tectonic events in the study area. The décollement zone in the study area consists of the Bergsdalen Nappes in addition to phyllites and mica schists.

Structural development

The structural-kinematic analyses carried out in this study suggest a separation of the Caledonian structures into two distinct groups. An early set of penetrative structures which indicate top-to-the-ESE translations is consistently overprinted by top-to-the-WNW structures (Fig. 5). This pattern has also been recognized elsewhere in the Scandinavian Caledonides, e.g. in the Nordfjord-Sogn area (Séranne et al. 1987, Chauvet & Séranne 1988) and in the nappes of northern (Rykkelid 1992, Fossen & Rykkelid 1992) as well as southern (Andresen 1974, Milnes & Koestler 1985, Olesen 1986, Fossen 1992) Norway. The top-to-the-ESE deformation and the later backward movement event are so distinct in terms of relative age and asymmetry of their related structures that they have been named D1 and D2, respectively. The first event is interpreted as related to the main Caledonian nappe translation in Ordovician(?)-Silurian times. The second event was basically a reversal of the D1 movement to form shear zones and asymmetric fabrics indicating top-to-the-WNW transport. Both events involved extensive syn-kinematic recrystallization under greenschist to lowermost amphibolite-facies conditions, and their heterogeneous nature has resulted in areas of low D2/high D1 strains and vice versa.

D1 structures

A variety of gently SE-dipping to sub-horizontal linear and planar fabrics formed in the Bergsdalen Nappes during the D1 event. These are best studied in areas of weak or no D2 reworking. A regionally penetrative foliation and lineation developed in all lithologies during D1. The foliation varies from a weakly developed shape fabric in the least deformed lithologies (typically the larger dioritic and granitic bodies) through gneissic to truly mylonitic in high-strain zones (Fig. 6 a). In the quartzites, the first foliation to form is seen to be axial planar to F1 folds which fold the bedding.

Where (proto)mylonitic gneisses are developed, the fabric varies from strongly planar to strongly linear, although L(S) tectonites appear to be more common than S(L) tectonites. The foliation and lineation are defined by mm- to cm-wide white or reddish bands or ribbons of quartz and feldspar enveloped by more irregular, finer-grained, and completely recrystallized feldspar, quartz and mica (Fig. 6 b). Feldspar porphyroclasts deformed by both brittle

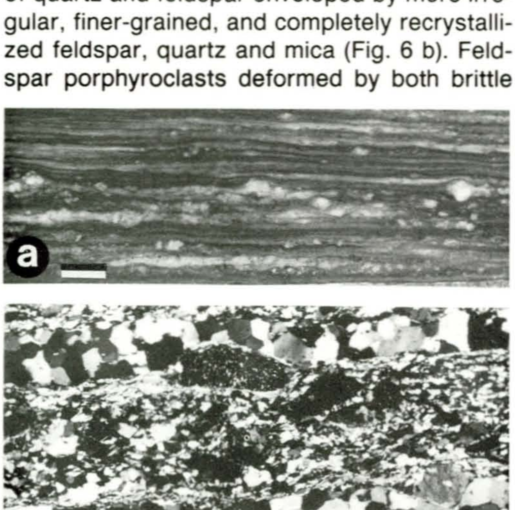

Fig. 6. a) Mylonitized granite, LBN near Torfinnsvatnet south of Voss (LN493197). Sense of shear is dextral (top-to-the-east). The sense of shear is not immediately obvious in such intensely D1-mylonitized rocks of the Bergsdalen Nappes. Scale bar = 1cm. b) Domains of quartz alternating with more fine-grained domains of mostly quartz, feldspars and micas. Same sample as above. Scale bar = 1mm.

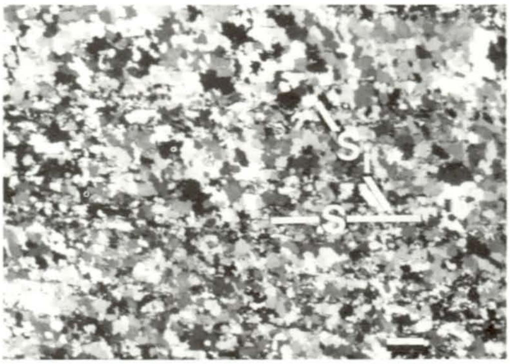

Fig. 7. Quartzite, with mesoscopically visible foliation (S) sub-horizontal, and an oblique grain shape fabric (Si) indicating a top-to-the-east (left) sense of shear. Krampane, near Hamlagrøv., LBN (LN393268). Scale bar = 1mm.

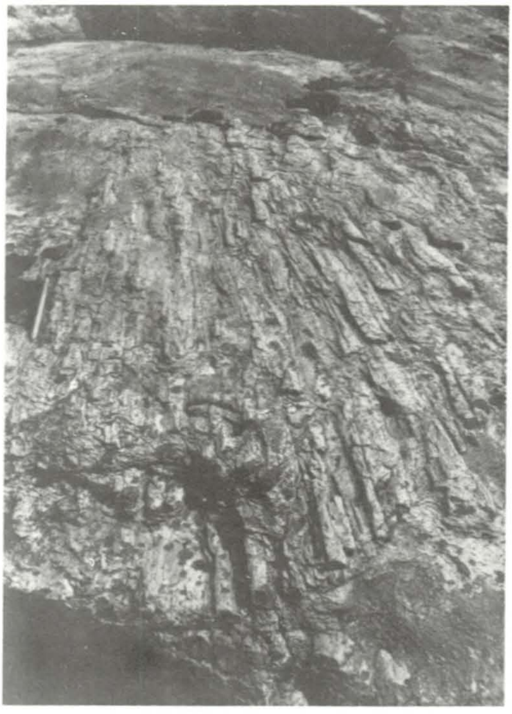

Fig. 8. Pebbles in quartzite conglomerate, stretched into rods (constrictional strain). North of Holmavatnet, LBN (LN336078), looking east. Note pencil for scale.

and crystal-plastic deformation, and small tails of recrystallized feldspar grains occur. In many of the mylonitic granitoids, feldspar porphyroclasts are commonly broken into aggregates consisting of floats of the original feldspar surrounded by more fine-grained, dynamically recrystallized feldspar grains. In some of the feldspar-rich deformed granitoids, the feldspar grains show no sign of plastic deformation, but are bounded by sharp, straight boundaries along which slip appears to have occurred. The deformation mechanism responsible for this texture is thought to be one of grain boundary sliding, apparently triggered by the absence of abundant quartz in these feldspar-rich granitoids.

Many feldspar porphyroclasts in granites and rhyolites continue laterally into mm-thick bands of nearly pure quartz and small amounts of sericite, interpreted as sinks of SiO2 transferred from the mica-rich layers by diffusion (e.g. Robin 1979). Thus, a variety of processes (solution transfer, recrystallization, grainsize reduction of porphyroclasts, grain boundary sliding, etc.) co-acted to form the present, pervasive L-S fabric in the Bergsdalen Nappes.

The S1 foliation in the quartzites of the Bergsdalen Nappes is composed of a meso- to microscopic compositional banding which in many cases can be shown to represent transposed bedding. In high-strain zones, however, the foliation may be the result of metamorphic processes, particularly in quartz (-mica)schists where the common foliation is a pressure-solution cleavage. Although a mesoscopic foliation may be easily detectable in the quartzites, the shape fabric defined by the preferred orientation of quartz grains is not very pronounced (Fig. 7), indicating that stretching of quartz grains appears to have been efficiently counteracted by dynamic recrystallization during deformation. Evidence for recrystallization by both grain boundary migration and subgrain rotation is found in the quartzites. The steady-state grain size in single-phase portions of the D1 recrystallized quartzites is about 2-300 μm, indicating D1 shear stresses of about 7-17 MPa (cf. Ord & Christie 1984), and the teextures are comparable with those in regime 3 of Hirth & Tullis (1992).

Associated with the pervasive D1 foliation is a variety of lineations (cf. Fossen 1993). The presence of a ribbon or shape fabric lineation in deformed granitoids has been mentioned above. A different kind of stretching lineation is seen in the deformed conglomerates (Fig. 8). Alignment of elongated minerals other than feldspar aggregates is also common, e.g. amphiboles in amphibolites. Deformed quartz veins occur in all rocks in the region, and a fine (mm-scale) striping or corrugation on the interface between the vein and the host rock defines a lineation in addition to the rod-structu-

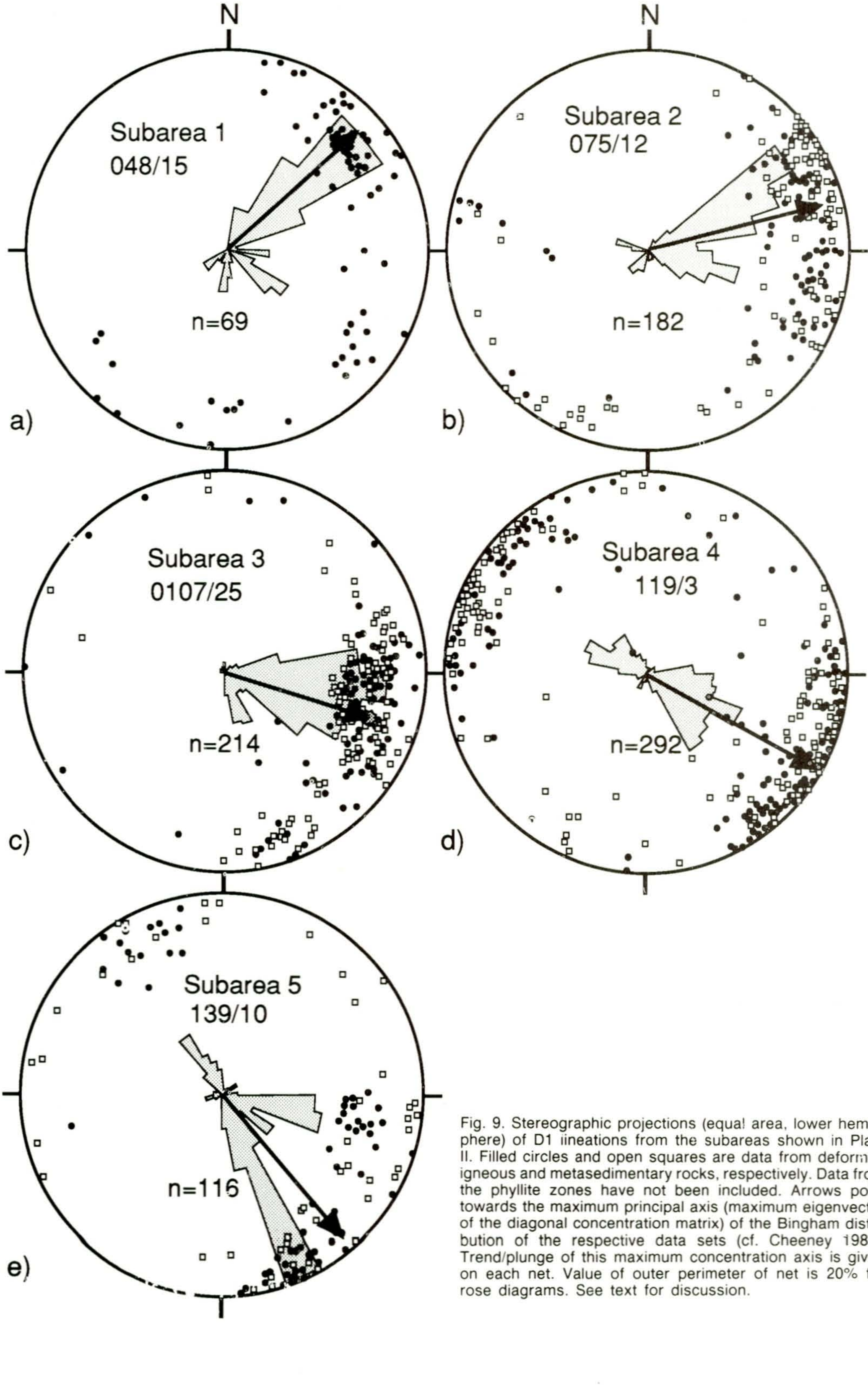

Fig. 9. Stereographic projections (equal area, lower hemisphere) of D1 lineations from the subareas shown in Plate II. Filled circles and open squares are data from deformed igneous and metasedimentary rocks, respectively. Data from the phyllite zones have not been included. Arrows point towards the maximum principal axis (maximum eigenvector of the diagonal concentration matrix) of the Bingham distribution of the respective data sets (cf. Cheeney 1983). Trend/plunge of this maximum concentration axis is given on each net. Value of outer perimeter of net is 20% for rose diagrams. See text for discussion.

re that is characteristic of strongly deformed quartz veins. The intersection between a compositional layering in metavolcanic rocks or primary bedding in the quartzitic metasediments and the penetrative S1 foliation gives rise to a well developed intersection lineation, which is parallel to the related F1 fold hinges. An intersection lineation is also developed where the penetrative S1 foliation is folded by somewhat later D1 folds. This structural overprinting, interpreted as resulting from a progressive D1 event, is quite common in the well deformed quartz(-mica) schists, particularly in the UBN.

The parallelism of the various D1-related linear structures mentioned above is a striking feature of the area. However, a systematic spatial variation in the trend of the overall gently plunging lineations is clear from Plate II. Going from the WGR (basement) where the lineations have a variable, but mostly northeasterly plunge (Fig. 9 a), the trend becomes gradually more E-W through the LBN (Fig. 9 b-c) and finally shows a NW-SE trend (Fig. 9 e). Together, the lineations form a consistent linear pattern which varies from E-W in the western (lower) part of the nappes, to more SE-NW towards the east (Plate II). There is no significant difference in the linear pattern developed in metasedimentary rocks (mostly quartzites and quartz-mica schists) and meta-igneous rocks (granitoids, metarhyolite, quartz diorite and metadacite) (Fig. 9). The spatial variation in linear trend may reflect a primary D1 pattern. If the linear pattern is taken to roughly reflect the transport direction, as argued by Kvale (1960), and if the reorientation of the D1 lineations by D2 deformation is generally-small, then the variation may indicate a change from orogen-parallel (sinistral) movements in subarea 1 (WGR) through oblique movements in the LBN (subareas 2 and 3) to transverse (SE directed) movements in the UBN (subareas 4 and 5).

The presence of D1 structures in the phyllites and mica schists is less obvious than in the Bergsdalen Nappes, as they have been more strongly affected by the later D2 deformation. However, it is locally quite clear that a strong domainal cleavage seen in some of these rocks is of D1 age. In rare areas preserved from D2 shearing, e.g. close to the phyllite-basement contact along the main highway from Voss to Granvin (just off the eastern limit of Plate I) or in the phyllites close to the Jotun Nappe in the northern part of Plate I, one can see asymmetric folds and S-C structures related to thrusting.

Kinematic indicators

The D1 structures have been distinguished from later deformation structures by systematic kinematic analysis. A range of kinematic indicators is associated with the D1 deformation fabrics, some of which can be recognized in the field or by studying cut hand samples, whilst others require microscopic investigations. The type and character of the kinematic structures depend on lithology as well as strain. In weakly deformed igneous rocks, particularly in the large quartz-diorite body in the LBN (Plate I), discrete Ramsay-Graham (1970) type shear zones are developed. The angular relationship between the shear zone-related foliation and the shear zone boundaries reveals the sense of shear, and the associated lineation indicates the slip direction. In some cases granitic dikes or other markers are deflected by such shear zones, and provide additional evidence for the sense of movement.

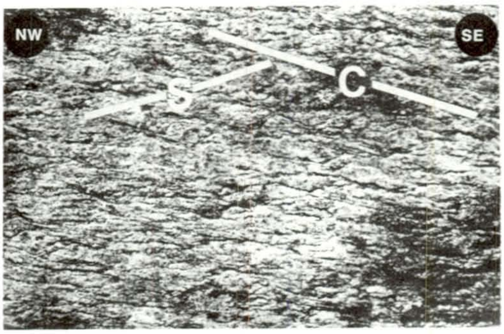

Fig. 10. S-C structures in granite at the base of the Jotun Nappe close to the contact with the UBN, indicating top-to-the-SE sense of shear (D1). Width of figure is about 20cm. Skreieggi (LN566429).

Moderately sheared plutonic rocks show classical S-C structures of the type described by Berthé et al. (1979) (Fig. 10), indicating (S)E translations, although parts of the granites, particularly in the LBN near Eksingedalen, show symmetric fabrics which may indicate coaxial strain. Truly mylonitic granitoids show broken feldspar porphyroclasts, or porphyroclasts with asymmetric tails or pressure shadows. The weak grain-shape fabric which occurs in quartz-rich bands is oblique to the

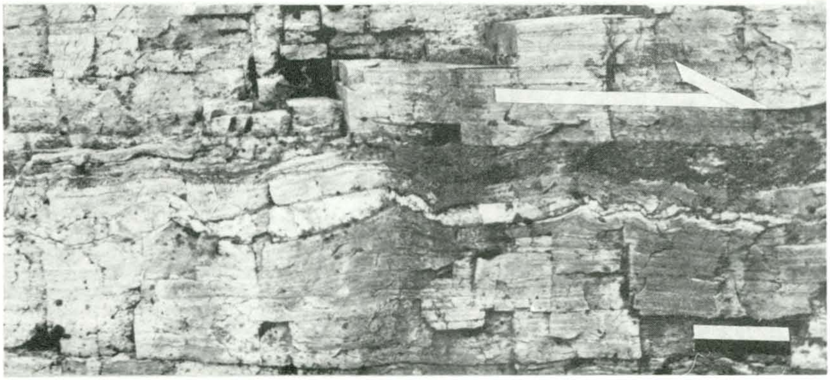

Fig. 11. Asymmetric boudins of vein quartz in quartz schist, indicating a top-to-the-SE sense of shear (D1). Scale bar = 10cm. West of Torfinnsvatnet (LN448175).

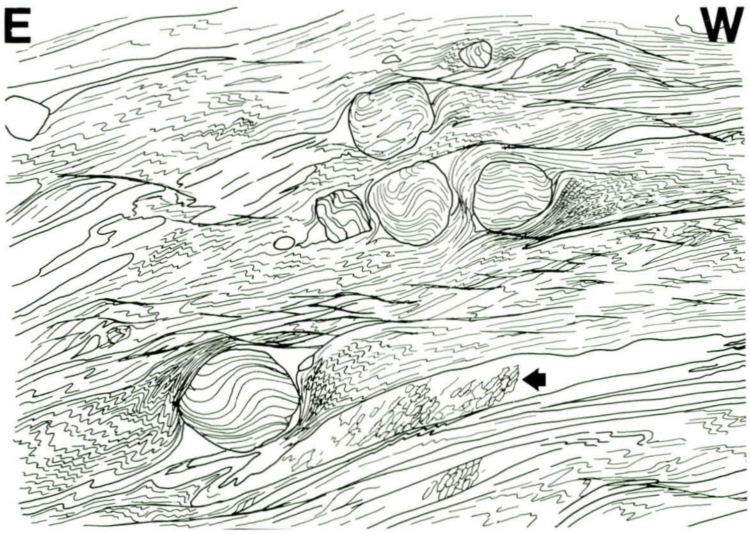

Fig. 12. Drawing of thin-section of garnet-mica schist, Børdalsfjellet, Kvamskogen, LBN (LN285006). Note that the sense of rotation indicated by curved inclusion trails in the garnets (top-to-the-E) is opposite to the shear sense indicated by micro folds, asymmetric strain shadows around the garnets, oblique grain shape fabric (arrow) in quartz bands, and shear bands in the matrix. The inclusion trails are generally discontinuous with, and thus older than the foliation in the matrix. Width of figure is 7.3 cm.

mylonitic foliation, and consistent with the (S)E translations (Fig. 7).

In the more micaceous lithologies in the Bergsdalen Nappe, shear bands indicating a top-to-the-E shear sense have locally been preserved from later D2 deformation, as have foliation fish or asymmetric boudins (Fig. 11). In the garnet-mica schists, the garnets contain curved inclusion patterns which may be interpreted as rotational fabrics (Rosenfeld 1970, however, see also Bell 1985), in which case the majority of the patterns indicates a top-to-the-E sense of shear (Fig. 12). The garnets occur as porphyroclasts in the matrix which shows abundant evidence for top-to-the-W (D2) reworking (shear bands, microfolds, asymmetric grain shape, etc., see below).

The orientation of quartz *c*-axes has been reported to be anomalously high at one locality in a quartzite in the LBN (Kvale 1945). More thin-sections were analyzed from different parts of the quartzites in the LBN to see if this is a general feature, and if there is an asymmetry to the microfabrics relative to the foliation. Samples were collected from localities where the later D2 strain appeared to be relatively low. In general, the fabric diagrams (Fig. 13) show a considerable variety, and may be categorized into three distinct groups of patterns: Single point maxima (samples 6 and 114), single-girdle patterns (samples 110 and 231), and cross-girdle patterns. The first group is best developed in the area sampled by Kvale (1945), where a maximum of 34.1% was

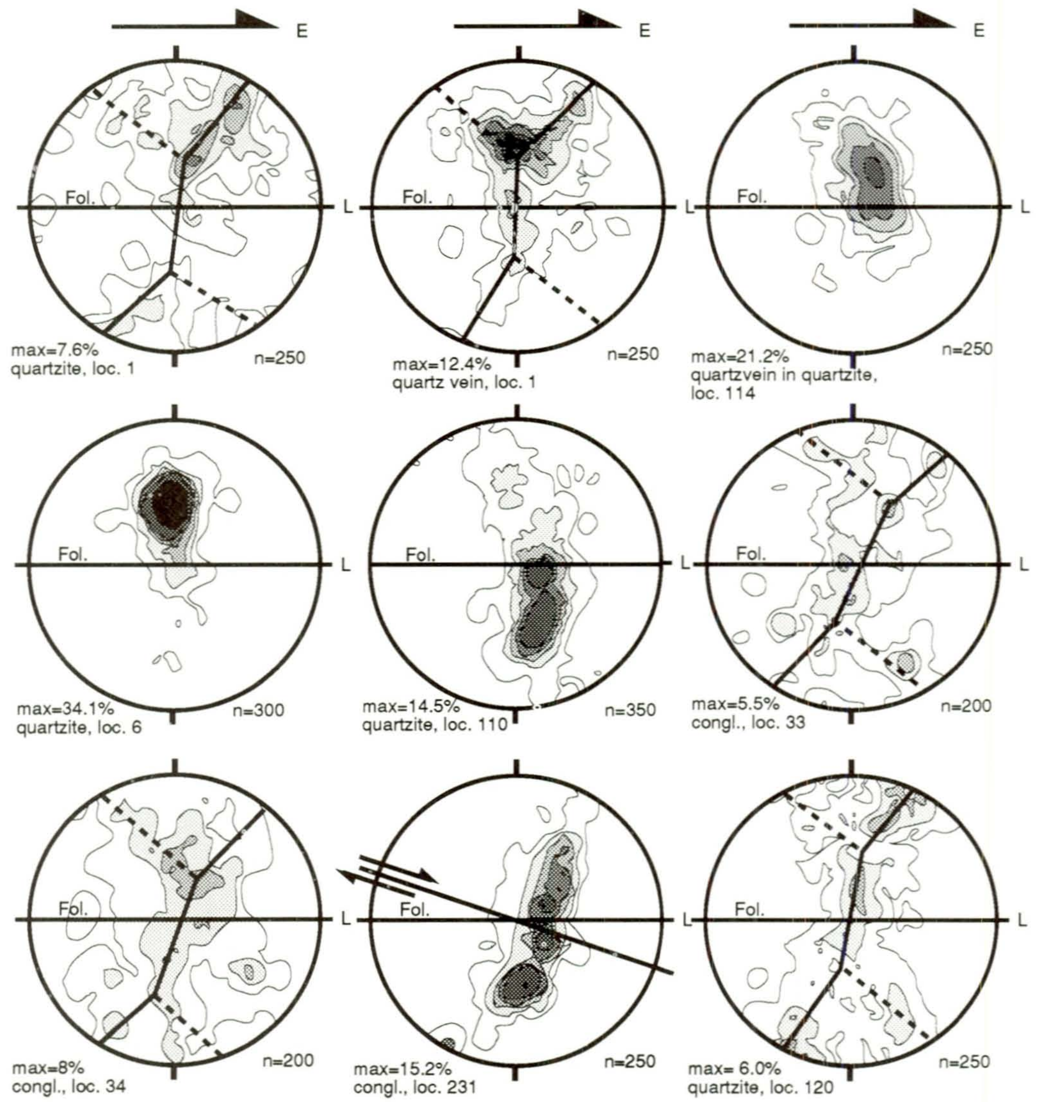

Fig. 13. *c*-axis fabrics from quartzites in the LBN. Most are consistent with the top-to-the-E sense of shear indicated by independent kinematic indicators, and are interpreted to be D1 in age. Contour intervals: 0.5-2, 2-4, 4-6, 6-8, 8-15, 15-25, 25-35%. See Plate II for locations and text for discussion.

reproduced (Fig. 13, loc. 6). In other areas maxima are still relatively high (6-21%), and the reason for this and for the variation in the position of the maximum is not known.

A particularly interesting observation is the asymmetry of some of the crystallographic fabrics with respect to the foliation. This is a common feature in shear zones, and the relationship between the sense of asymmetry and the sense of shear is well established (see Simpson & Schmid 1983 for a review). Some of the single-girdle fabric diagrams (particularly loc. 231) resemble closely fabrics expected from an active basal glide plane together with first-order rhombs and first-order prisms, i.e. *c*-axis sub-maxima along a single girdle. The asymmetry of this girdle with the foliation is generally taken to indicate the shear direction, assuming that the *c*-axis maximum forms normal to the shear plane (Etchecopar 1977). Sample 231 thus indicates a top-to-the-E sense of shear.

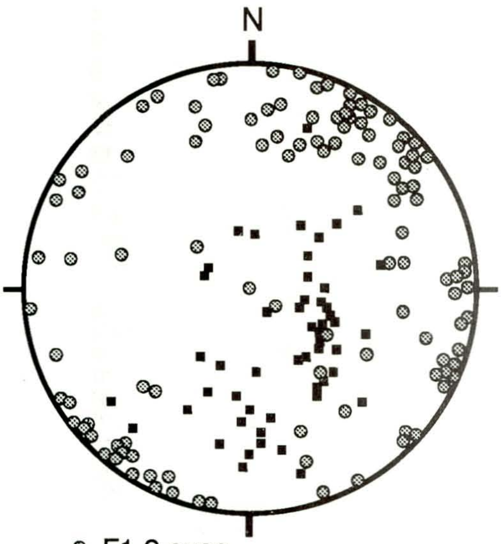

Fig. 14. Stereographic projection of late D1 folds and their axial plane cleavage from the UBN. Data mostly from the Torfinnsvatnet and the Volavatnet areas.

Most of the crossed-girdle patterns are also slightly asymmetric with respect to the foliation. The skeletal outline tentatively indicated in Fig. 13 illustrates the asymmetry (particularly in samples from locs. 33 and 34) and conforms with a top-to-the-E shear sense. The fabric of sample 114 is the only one that may indicate a top-to-the-west sense of shear, possibly due to reworking during D2.

A weak grain shape asymmetry seen under the microscope (Fig. 7) is also asymmetric with respect to the foliation, and indicates a shear sense consistent with that inferred from the c-axis diagrams above. This is also in agreement with the impression that the stretching of the conglomerate pebbles and the formation of the general linear pattern occurred during the top-to-the-E (D1) shear deformation.

Most of the D1 folds have axes parallel with the lineation, but those that are not are asymmetric with a vergence consistent with the top-to-the-(S)E sense of shear. They vary from open to isoclinal, and they fold the D1 foliation and/or the primary layering, indicating that they formed progressively during D1. For a more comprehensive discussion of D1 lineations and folds, see Fossen (1993).

A distinct phase of folding which occurs only in the UBN affects the linear and planar D1 structures described above, and are here named F1.2 folds. The axes of these folds are sub-horizontal and mostly NE-SW-trending, and the folds have sub-horizontal to gently NW-dipping axial planes (Fig. 14). A spaced axial planar cleavage (S1.2) is locally developed (Fig. 15). The folds are open to close, and their SE vergence is consistent with the sense of vorticity determined for the D1 deformation. They are overprinted by the D2 deformation described below, and are best explained as having formed at a late stage during thrusting.

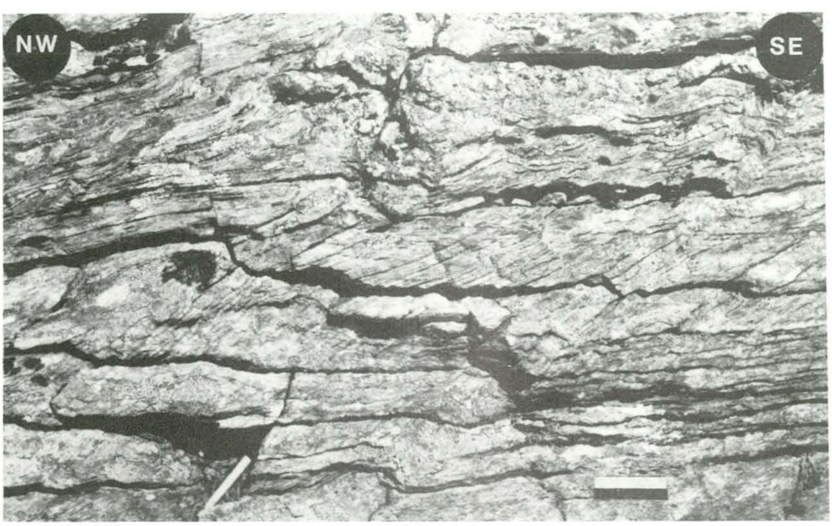

Fig. 15. Spaced cleavage related to SE-verging folds of late-D1 age in metarhyolite in the UBN. Kvådalsfjellet near Torfinnsvatnet (LN490144), looking northeast.

Strain and strain history during D1

The state of finite strain has been determined from a number of localities in quartzite conglomerates in the LBN. These conglomerates are fairly monomictic with quartzite pebbles set in a quartzitic matrix, making them well suited for strain determinations. Conglomerates also occur in the UBN, but these also contain other clast types, particularly granitic pebbles. More important is the more micaceous matrix in the polymictic conglomerates in the UBN, which produced a viscosity contrast between the various pebbles and the matrix, and made these rocks receptive to later folding, disturbing the primary D1 strain. These conglomerates have therefore not been used for quantitative strain analyses, but they qualitatively seem to have experienced constrictional strains similar to the conglomerates in the LBN. The latter (Fig. 16) indicate that the finite strain was clearly constrictional, and flattening strains have not been encountered in the conglomerates. Unfortunately, the quartzite conglomerates have a fairly restricted occurrence compared to the extent of the Bergsdalen Nappes, and may therefore not be representative of the whole area. However, an impression of the general strain geometry can be obtain by discriminating between strong L-fabrics (constrictional), strong S-fabrics (flattening) and equally well developed S and L fabrics (Flinn k-value close to 1). Using this approach, it appears that L(-S) fabrics dominate the Bergsdalen Nappes, although S-dominated fabrics do occur in several places.

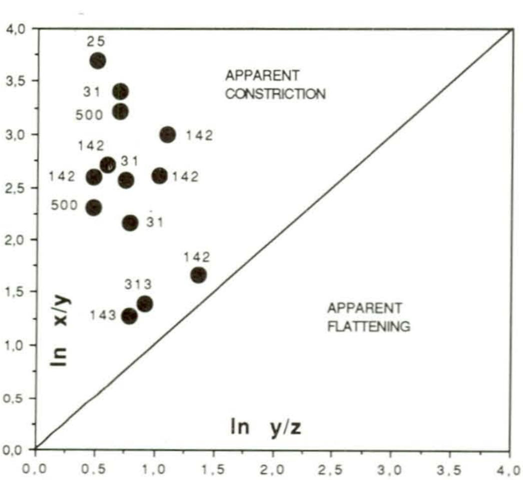

Fig. 16. Strain data (logarithmic Flinn diagram) from quartzite conglomerates, LBN. See Plate II for locations and text for discussion.

The constrictional strain may have been formed by a single, simple deformation, or may be the result of a more complex strain path due to separate pulses of deformation or to non-steady state flow during deformation. Since any significant information about the strain path is lacking, no conclusion can yet be drawn as to the strain history. However, a simple model combining simple shear (thrusting) and horizontal pure shear with maximum extension in the thrusting direction has been suggested (Fossen 1993), which explains the constrictional strain as well as the rapid rotation of fold axes towards the lineation. Independent of the model favored, constrictional strain in a relatively significant portion of the Bergsdalen Nappes must have been accommodated by other types of strain in other parts and, particularly, in the surrounding phyllites and mica schists.

M1 metamorphism

The metamorphic grade during D1 is constrained by the growth of white mica, biotite, amphibole, and garnet, and by the coexistence of oligoclase and albite in sheared lithologies (Kvale 1946), indicating lowermost amphibolite-facies conditions. This is consistent with the fairly equidimensional grain shape in the dynamically recrystallized quartzites, which requires temperatures of at least upper greenschist-facies conditions. Growth of kyanite in the northwestern part of the LBN (Eksingedalen, Gray 1978) indicates at least moderately high pressures in this part of the region. The metamorphic conditions are thus very similar to those in the nearby Lower Paleozoic rocks in the Major and Minor Bergen Arc (M2 in Fossen 1988 a).

D2 structures

The regional set of fairly penetrative, linear and planar fabrics that was established during D1 was consistently deformed by a tectonic event (D2) which involved non-coaxial deformation with a roughly opposite sense of shear. Abundant evidence, particularly the consistent overprinting relationship, the difference in structural asymmetry, and features such as shortened D1 boudins (Fig. 17), suggest that the D1 and D2 events are temporally distinct. The character of the D2 deformation depends

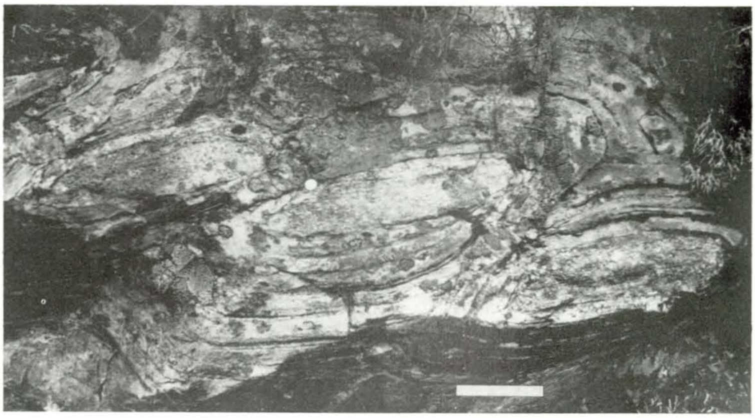

Fig. 17. D1-boudinaged granitic dike that was contracted during D2. LBN, south of Kvitingsvatnet (LN2999051).

on the local intensity of D2 and the pre-existing structures and rheology, but the most characteristic D2 structures are asymmetric folds and S-C structures.

D2 folds

The most widespread effect of the D2 deformation on the Bergsdalen Nappes is the development of asymmetric folds with a consistent NW vergence (overturned to the NW). Most of these folds have NE-SW-trending axes (Fig. 18), but local deviations exist, particularly in the southwestern part of the region (Plate III) where the axial trend is more E-W and their vergence is to the N. The axial surfaces of these folds dip 20-45° to the SE, and a crenulation cleavage is locally developed (Fig. 18), particularly in the hinge zones in mica-bearing lithologies.

The F2 folds vary in size from millimeter scale in micaceous lithologies (Fig. 12) through meter-large mesoscopic folds (Fig. 19) up to kilometer scale, asymmetric folds in deformed granitic sheets (Fig. 20, Plate I b). Several of the small-scale folds appear to be directly related to neighboring shear band structures, forming pairs of reverse and normal slip crenulations, respectively (Dennis & Secor 1987). F2 folds observed on outcrop-scale appear to be of two kinds. One is demonstrably related to larger-scale fold structures, i.e. parasitic folds. Such second-order folds are common in or near the hinge zone of their parent structure, and have a SE vergence where developed on the short, steep limb of first-order folds. The other type of F2 folds observed in the field is that of asymmetric, open to isoclinal folds which rapidly die out across the layering. They are common where there is an alternation between competent and incompetent layers in a zone between thicker and more homogeneeous, competent layers. Abundant evidence for layer-parallel slip (Fig. 21) together with the intrafolial nature of these folds indicate that they are the result of localized top-to-the-WNW shear. Many of these structures resemble "contractional composite structures" as described from the Øygarden Complex to the west of the Bergsdalen Nappes (Rykke-

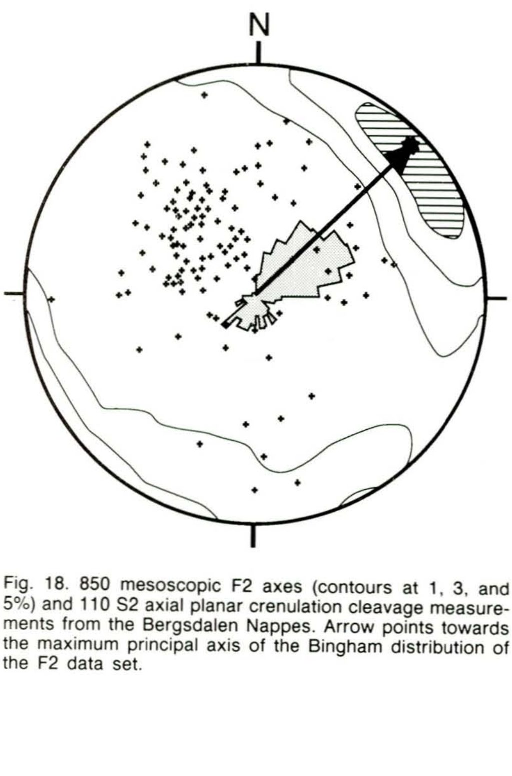

Fig. 18. 850 mesoscopic F2 axes (contours at 1, 3, and 5%) and 110 S2 axial planar crenulation cleavage measurements from the Bergsdalen Nappes. Arrow points towards the maximum principal axis of the Bingham distribution of the F2 data set.

Fig. 19. Mesoscopic F2 folds on the steep short limb of a major F2 fold, Holmavatnet, LBN (LN327062), looking southeast.

Fig. 20. Large-scale F2 folds in the LBN, looking southwest. a) photograph, b) drawing of the same view. Quartzite is shown as dotted layer, underlain by amphibolitic metagabbro. Near Gråfjellet west-northwest of Kvitingsvatnet (LN260083).

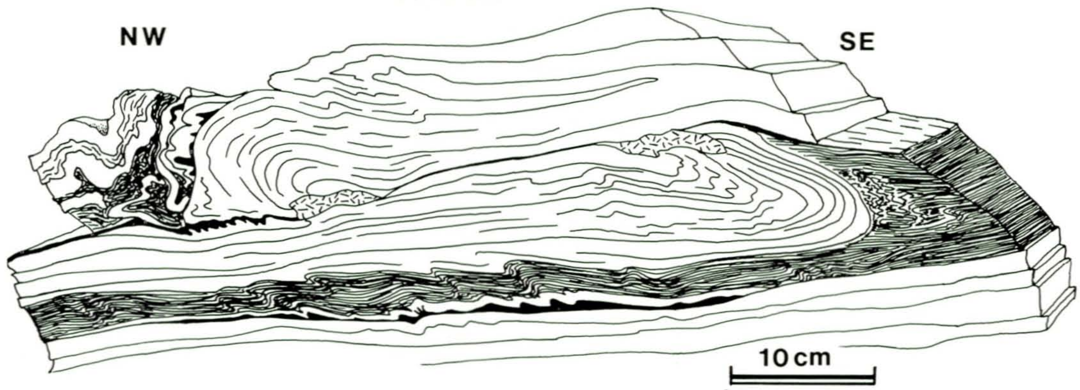

Fig. 21. Intrafolial F2 fold with discrete slip along the inverted short limb. LBN, near Frøland (LN246979).

lid & Fossen 1992), and several of these shear-related folds are developed above associated shear bands.

The larger-scale F2 folds are typically seen where sheets of competent, foliated rocks, typically quartzites and sheared granites, occur between less competent, more micaceous lithologies. The folds, several of which are visible on air photographs, have fairly constant NE-SW-trending axes. One of these large F2 folds was mapped by Gray (1978) as a klippe of the LBN in the Western Gneiss Region in the northwestern part of Plate I. It is thought that most of these large-scale folds are analogous to the intrafolial, mesoscopic folds, and that they developed as a result of the competent sheets being in the contractional field of the incremental strain ellipse during the backsliding (D2). Both the large and the small-scale folds consistently fold the composite D1 fabric (Fig. 22) and their top-to-the-E micro- and mesoscopic kinematic indicators.

Many of the vein quartz pods in the phyllites were also folded during D2, although these show a larger variation in orientation than do the F2 folds within the Bergsdalen Nappes (Fig. 23).

A particularly interesting feature is the local tendency of F2 folds of a certain axial trend to interfere with F2 folds of slightly different trend (self-interference). Fig. 24 shows an example of this from the UBN. Here a series of F2 folds with ENE-WSW axial trend meets another set of NNE-SSW-trending F2 folds (both with approximately NW vergence) to make a classical fold interference pattern. A classical interpretation would be that the two sets of folds represent two distinct phases of folding, probably the ENE-WSW-trending

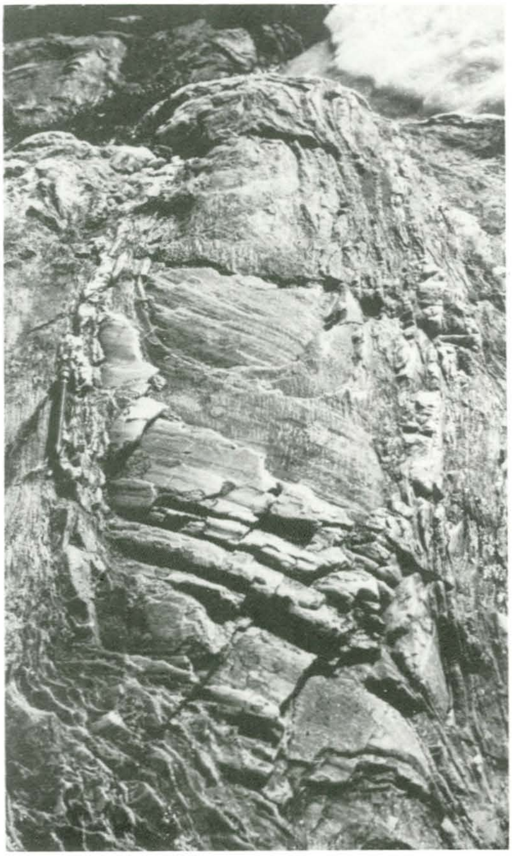

Fig. 22. Hinge zone of F2 fold, showing folded D1 lineation. A weak D2 crenulation lineation is developed parallel to the fold axis. Note pencil for scale. UBN, Geitafjellet (LN417074), looking northeast.

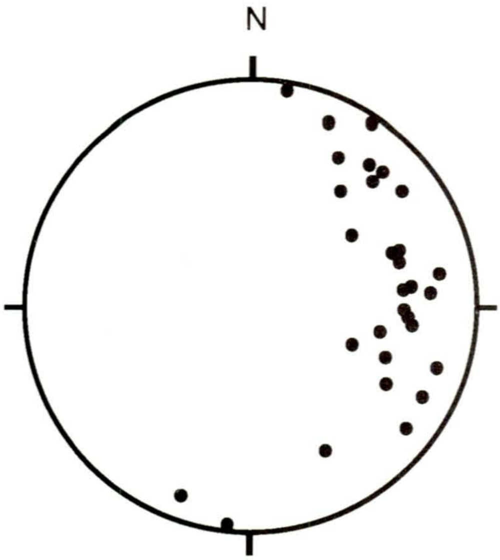

Fig. 23. Axes to folded vein quartz in phyllites from road sections near Bulken, west of Voss.

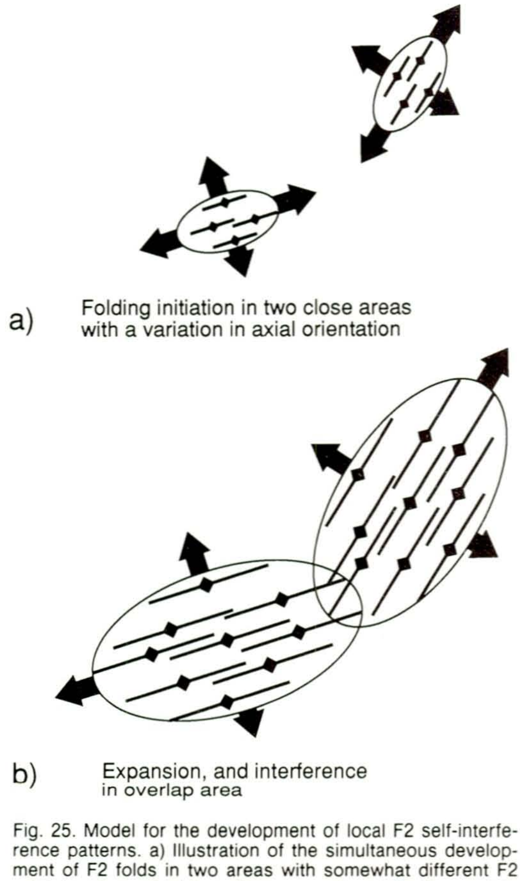

Fig. 25. Model for the development of local F2 self-interference patterns. a) Illustration of the simultaneous development of F2 folds in two areas with somewhat different F2 axial trends, and b) how they expand and finally meet to form a local interference pattern. Scale approximately as in Fig. 24.

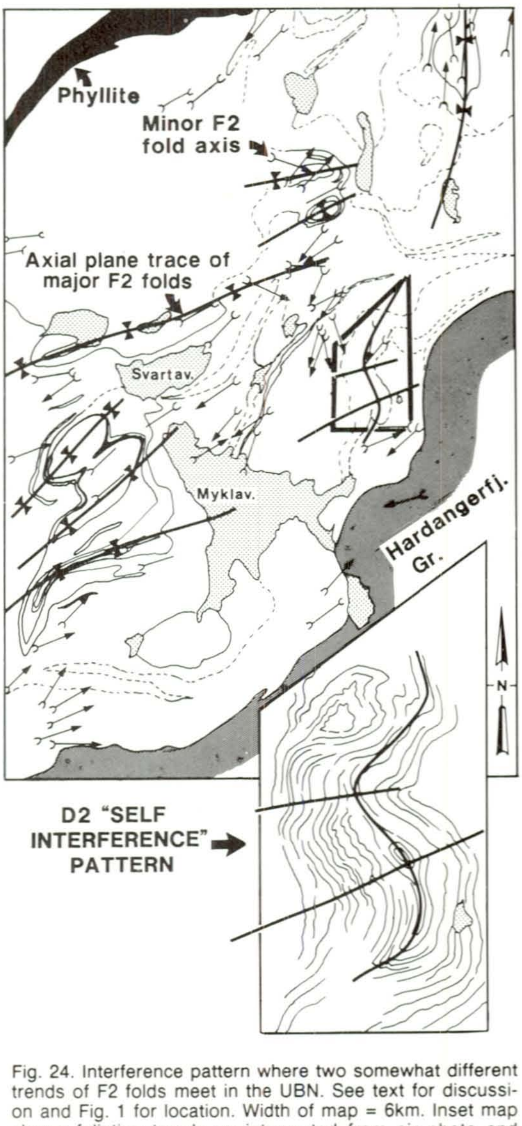

Fig. 24. Interference pattern where two somewhat different trends of F2 folds meet in the UBN. See text for discussion and Fig. 1 for location. Width of map = 6km. Inset map shows foliation trends as interpreted from air photo and field observations in area of fold interference.

folds being the younger set. However, if one accepts the general model that all the NW-verging structures are the result of a NW (backward) movement of the Jotun and related nappes above the Bergsdalen Nappes (cf. Fossen 1992), the age difference between the two sets of folds must be very small. Furthermore, the fact that the two sets of folds occur in distinct areas, and that interference is only encountered at the boundary between these two areas must be explained. The model favored here is that NW-verging folds initiated more or less simultaneously in different parts of the UBN, and expanded laterally (Fig. 25). If the folds initiated in two different domains with slightly different trends, fold interference would be expected where the two trends met, but only where they met, since the anisotropy generated by the two fold trends would hinder further expansion. The fact that the different F2 folds are all similar as far as geometry, kinematics, and time relationship to other structures go, favors this self-interference model.

Shear bands, S-C structures and other mylonitic D2 structures

Shear bands consistent with a top-to-the-WNW sense of shear occur in micaceous lithologies in the Bergsdalen Nappes. They form small, discrete shear zones which affect the D1 fabric in conglomerates and quartz schists (Fig. 26a). Shear bands are even more common in the phyllites and mica schists between the Bergsdalen Nappes, and occur in the field as centimeter to meter-long surfaces spaced on the order of less than a centimeter up to two meters (Figs. 4, 26 b). Except for the region along Hardangerfjorden, the shear bands tend

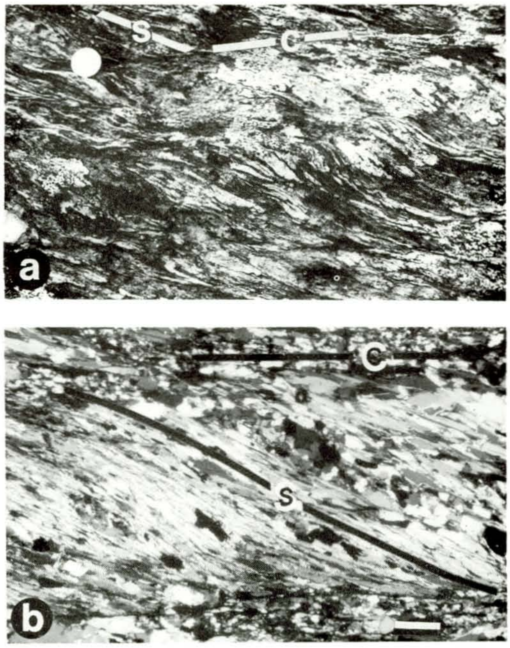

Fig. 26. Shear bands of D2 age developed a) in quartz-mica schists in the LBN between Evanger and Eksingedalen (LN445408), and b) in phyllite between LBN and UBN, Kvamskogen (LN337983). Sense of shear is sinistral (top-to-the-WNW) in both cases.

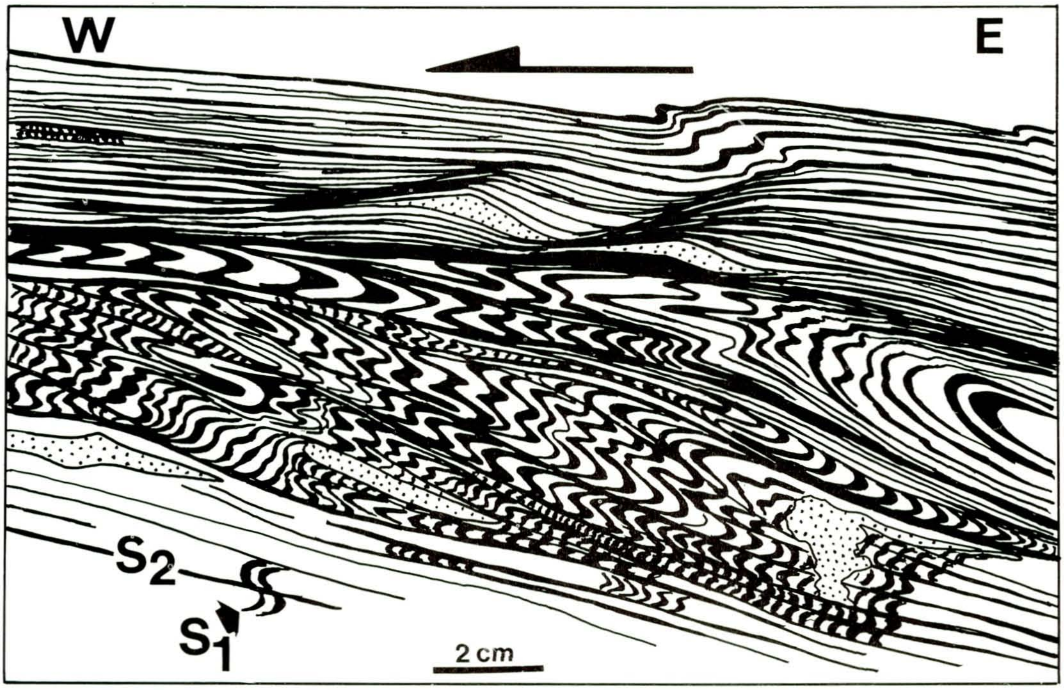

Fig. 27. S2 crenulation cleavage merging into mylonitic D2 foliation with shear bands consistent with a top-to-the-W sense of shear. Dotted objects are vein quartz pods. Phyllite zone between UBN and LBN north of Hamlagrøvatnet (LN470203).

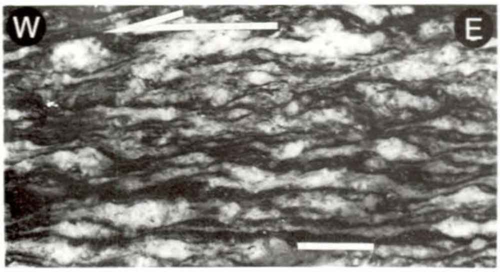

Fig. 28. Granite sheared during D2, showing top-to-the-W S-C structures. The planar, continuous shear bands are characteristic for the D2 deformation in such lithologies. LBN, north of Evanger (LN475360). Note that the foliation, which is horizontal in the photograph, dips about 30° to the SE in the field.

to be sub-horizontal, whereas the foliation dips to the SE at about 30-40° (Fossen 1992 fig. 10a). The foliation in the phyllites is typically a D1 fabric which in many places has been considerably modified by D2 deformation. However, in some places the foliation can be shown to be a true D2 cleavage formed by crenulation and metamorphic segregation of the D1 fabric and minerals (Fig. 27).

It is interesting to note that shear bands and asymmetric, NW-verging folds of the type described above commonly coexist without a clear overprinting relationship. Competent layers are typically folded, whereas the surrounding micaceous layers show an abundance of shear bands. This is taken as evidence for simultaneous development of folds and shear bands during D2.

Typical S-C structures of D2 age are developed in some of the granitoid rocks of the Bergsdalen Nappes, e.g. in the northern tip of the granite north of Evanger (Evanger Granite) (Plate I, Fig. 28). Asymmetric boudins (Hanmer 1988) of competent layers are common in the banded, mica-bearing gneisses (metadacites with granitic dikes), and can for instance be observed along road sections west of Evanger. The boudins are everywhere tilted more steeply to the SE than the general foliation, indicating that they formed during the top-to-the-WNW D2 shearing. Similar to the shear bands, the asymmetric boudins are associated with the asymmetric, NW-verging F2 folds which refold the D1 folds and fabrics.

Asymmetric grain shape fabrics are developed in quartz aggregates where the D2 deformation is strong. This is most commonly observed in micaceous rocks, particularly in the mica schists (Fig. 12), but occurs also in quartzo-feldspatic rocks where the D2 shear deformation has been strong, e.g. along the southwestern side of the Hardangerfjord (Fig. 29). The grain size of recrystallized quartz during D2 seems to be smaller than the one during D1 (compare Figs. 7 and 29), indicating higher flow stresses during D2 (cf. Ord & Christie 1984).

D2 translation direction

The exact direction of shear during D2 is not everywhere obvious, as the orientation of preexisting structures and anisotropy may have influenced the orientation of the D2 structures. However, the translation direction may be inferred with confidence in areas of high D2 shear strain where the lineation is clearly

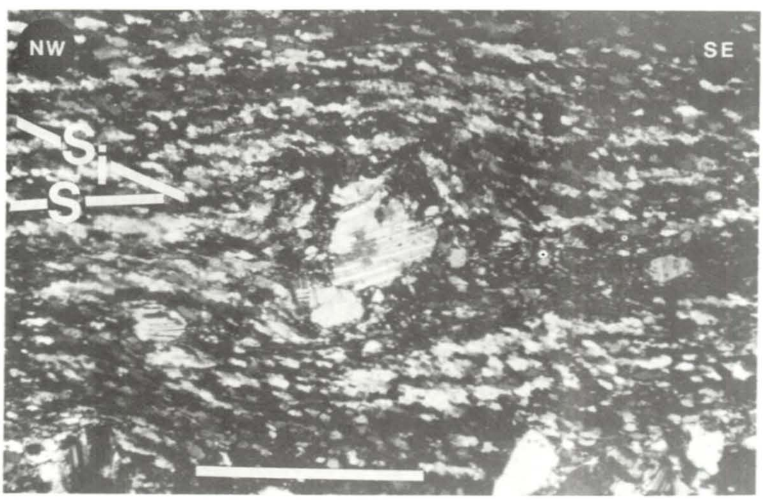

Fig. 29. Oblique quartz fabric, granitic mylonite gneiss, Hardangerfjord Shear Zone (LM531910). Sense of shear is sinistral (top-to-the-NW). Scale bar = 1mm.

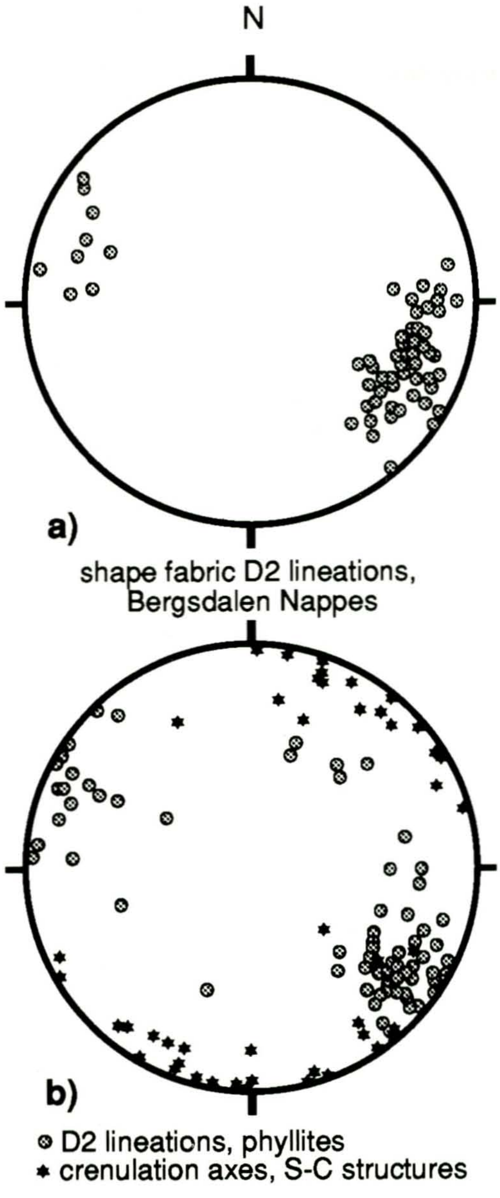

Fig. 30. a) D2 lineations in plutonic rocks in the Bergsdalen Nappes. b) D2 lineationsand axes to S-C crenulations in phyllites and mica schists between and beneath the Bergsdalen Nappes.

a result of D2 deformation, and not merely a reworked D1 lineation.

In places where the asymmetric (top-to-the-west) mylonitic structures discussed above indicate that the D2 deformation caused incipient mylonitization in granitic lithologies in the Bergsdalen Nappes (e.g. Fig. 28), the associated stretching lineations are interpreted to closely indicate the direction of movement during D2. This lineation, which is best developed in the tips of some of the granites in the LBN, has an ESE-WNW trend (Fig. 30 a).

Another reliable source of information is the fibre and mineral lineations developed in strongly sheared phyllites with penetrative fabrics indicating a top-to-the-west sense of shear. Evidence for intense top-to-the-west shear (see above) justifies the interpretation of this lineation as a D2 fabric, particularly where it is developed on the shear bands (C-surfaces) themselves. This lineation has a trend very similar to that of the D2 stretching lineation (Fig. 30 b). Together with the D2 stretching lineation these lineations are plotted on Plate III as solid, bold arrows.

The orientation of the crenulation axes (the line of intersection between the foliation, S, and the shear planes, C) of the S-C structures in the phyllites and mica schists would, ideally, be expected to form perpendicular to the slip direction. The crenulation axes measured in the field (Fig. 30 b) are consistent with this general expectation, and with a top-to-the-(W)NW sense of shear during D2. In Plate III, open arrows indicate the normal to the trend of these crenulation axes. In general, the data presented on Plate III and Fig. 30 indicate that the direction of D2 shearing was relatively constant (to the west-northwest) throughout the area.

M2 metamorphism

The metamorphic grade during D2 clearly was somewhat lower than during D1. This is evident from the retrogression associated with the formation of D2 fabrics, particularly in the phyllites and mica schists where M1 garnets and amphiboles are cracked and locally replaced by biotite and chlorite. Indications of retrogression in the granitic and quartzitic rocks in the Bergsdalen Nappes are restricted to the stronger D2 quartz shape fabric, which may indicate lower recrystallization rates and thus lower temperatures, and possibly also the finer recrystallized quartz grain size of the D2 fabrics than of the D1 fabrics. However, the latter feature may also be related to higher strain rates and localization of deformation unrelated to a lowering in temperature from D1 to D2. The many similarities in D1 and D2 microtextures and structures indicate that the PT conditions were not very different, and middle to upper greenschist grade is suggested for M2.

Later folding

Open folds with gently E-plunging axes and vertical axial planes occur sporadically in the Bergsdalen Nappes, but are best developed in the Eksingedalen area where they have previously been described by Gray (1978). They appear to fold the D2 structures described above, and are the last expression of ductile deformation recognized in the region. They are geometrically and chronologically very similar to meso- and macroscopic folds in the Bergen Arc System which are likely to be related to the large-scale arc structure itself (Fossen 1988 b), and a correlation is tentatively proposed with folds of very similar geometry in the Devonian basins to the north (Høister 1971, Skjerlie 1971).

Extension to surrounding areas

The two-fold, Lower Paleozoic deformation history recorded in the Bergsdalen Nappes is also recorded in the surrounding geological units. To the northwest, the D2 folds can be traced into the underlying Western Gneiss Region (Fig. 31), where also porphyroclast systems, small-scale imbricate structures, intrafolial folds and other shear sense indicators indicate heterogeneous reworking of Caledonian (D1) and earlier, Precambrian structures (Fig. 32). The effect of D2 is weak or absent in a portion of the Western Gneiss Region lying northwest of the Bergsdalen Nappes (Fig. 33) but increases in intensity towards the west where the generally SE-dipping foliation is deflected into the extensional Nordfjord-Sogn

Fig. 31. Folded gneiss at the contact between the Western Gneiss Region and the phyllites underlying the LBN, Skjerjavatnet, Eksingedalen (LN450551), looking northeast.

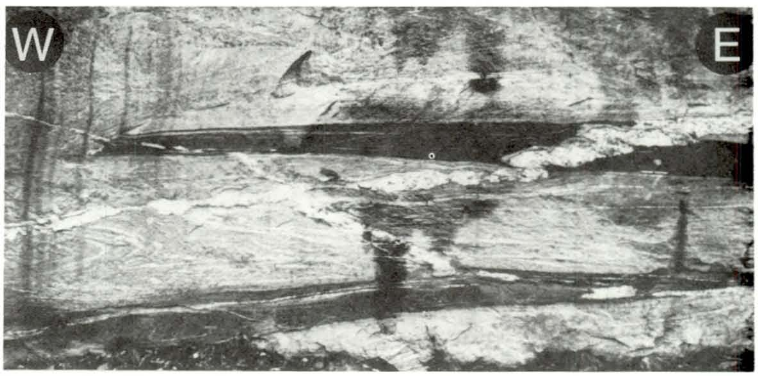

Fig. 32. Evidence for late top-to-the-W shearing in Precambrian migmatites in the Western Gneiss Region, Straume. Westerly-directed shear is localized to dark bands, and granitic dikes are folded or boudinaged depending on their orientation with respect to the incremental strain ellipse. Height of photo is about 1 meter. See Fig. 1 for location.

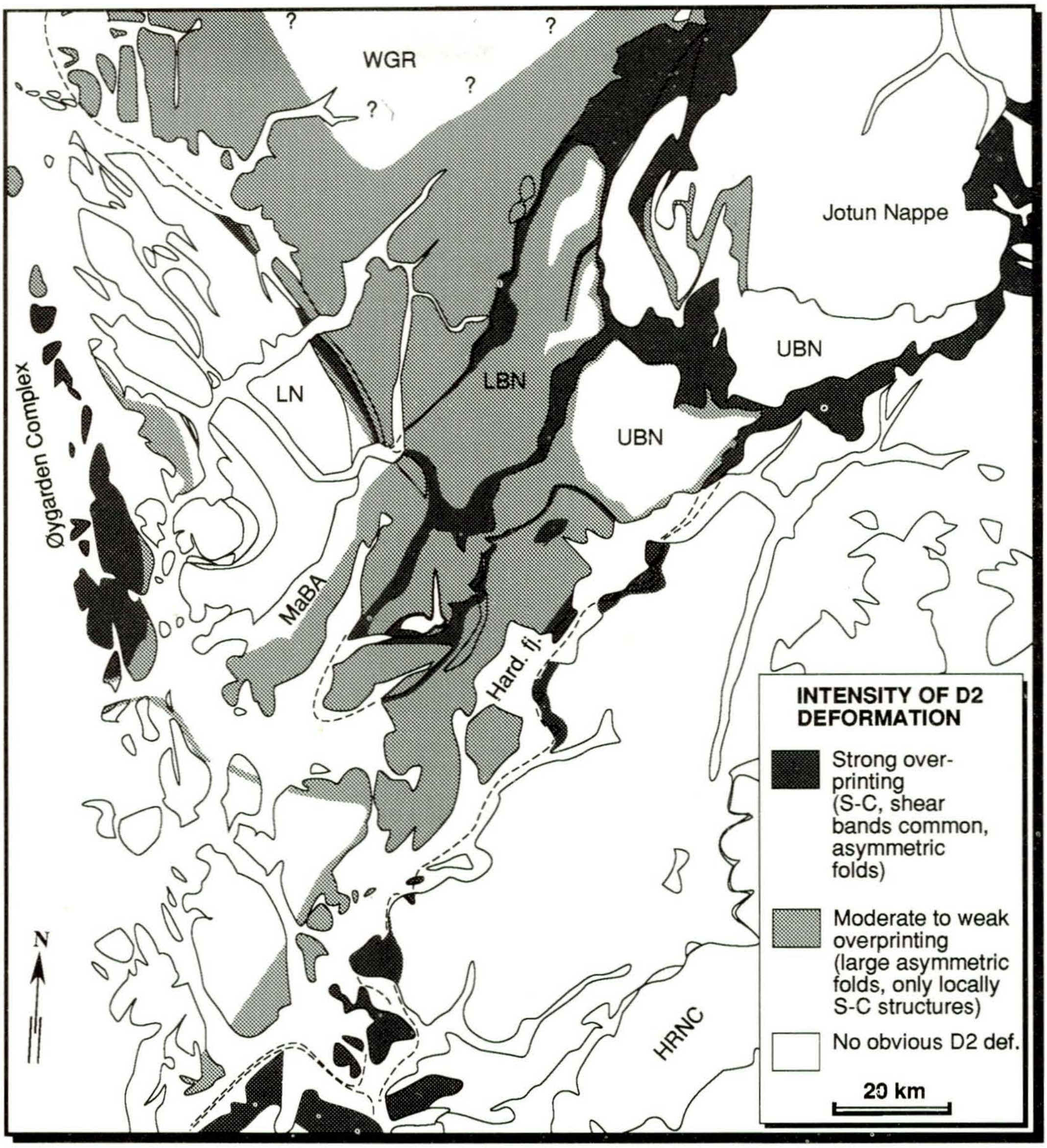

Fig. 33. Intensity of the D2 deformation as interpreted from qualitative fabric studies in the field and in the laboratory.

Detachment (Norton 1986, 1987, Milnes et al. 1988). Furthermore, a structural history and fabric development very similar to the one described in this paper is reported from the upper plate rocks in the Nordfjord-Sogn Detachment zone north of the study area (Séranne & Séguret 1987, Chauvet & Séranne 1988, Séguret et al. 1987, Séranne et al. 1989, Swensson & Andersen 1991).

D2 structures, including intense shear band development, are even more pronounced in the décollement zone to the east of the Bergsdalen Nappes than in the Bergsdalen area, and are found in the phyllites all the way to the southeastern side of the Jotun Nappe (Milnes & Koestler 1985, Fossen 1992). Asymmetric D2 folds overturned to the NW have also been described in the Haukelid region south of the study area (Andresen 1974).

The boundary between the Bergsdalen Nappes and the Bergen Arc System is marked by a change in dip of lithological contacts and

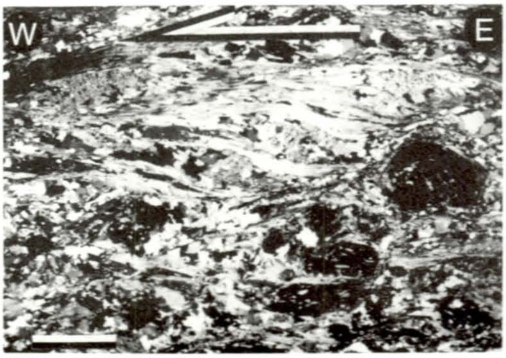

Fig. 34. S-C structures in amphibole-mica schists in the Major Bergen Arc, between Hindnesfjorden-Osterfjorden, Lindås (see Fig. 1 for location). Sense of shear is sinistral (top-to-the-W). Scale bar = 2mm.

the Caledonian fabrics in the Bergsdalen Nappes from the general SE dip to moderate to steep dips towards the west (Fig. 1). Associated with this contact are asymmetric structures indicating down-to-the-W shear (Fig. 34), suggesting that this contact is genetically related to the extensional Hardangerfjord Shear Zone and the Nordfjord-Sogn Detachment.

Going into the Bergen Arc System, the sense of shear gets fairly complex in the Lindås Nappe, while top-to-the-E asymmetric mylonite structures related to thrusting (D1) dominate the Ulriken Gneiss Complex and Minor Bergen Arc (Fossen 1988 b). The D1 (thrusting) deformation can be traced into the Øygarden Complex, which is interpreted as Precambrian basement considerably reworked by Paleozoic deformation (Bering 1984). Unambiguous top-to-the-E S-C structures occur along the contact with the Minor Bergen Arc and some kilometers into the Øygarden Complex (Figs. 33, 35). However, toward the west there is a change in sense of shear from top-to-the-E to top-to-the-W, and in the western portion of the Øygarden Complex, the top-to-the-W shear sense is completely dominant (cf. Fossen & Rykkelid 1990, Rykkelid & Fossen 1992). Compelling evidence for the relative age of the two different shear events has not been found, and although it was indicated by Fossen & Rykkelid (1990) that the top-too-the-E shearing might be the older, the striking similarity between the top-to-the-W fabrics in the Øygarden Complex and the Bergsdalen Nappes may indicate that they are both related to D2 shearing.

The Bergsdalen Nappes and the phyllites reappear along the southeastern side of Hardangerfjorden. Although relics of the D1 deformation occur, the present fabrics are to a large extent the product of strong D2 shearing. Sense of shear indicators include various types of S-C structures (Fig. 36) and asymmetric, intrafolial folds (Fig. 37), asymmetric porphyroclasts, and asymmetric quartz *c*-axis fabrics (Fig. 38) and are found in the nappes,

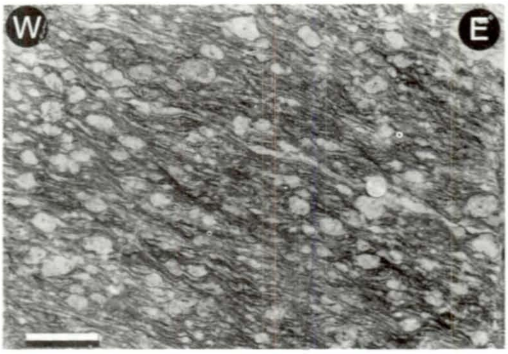

Fig. 35. S-C structures in augen gneiss, indicating a dextral (top-to-the-E) sense of shear (D1). Askøy (KN928036).

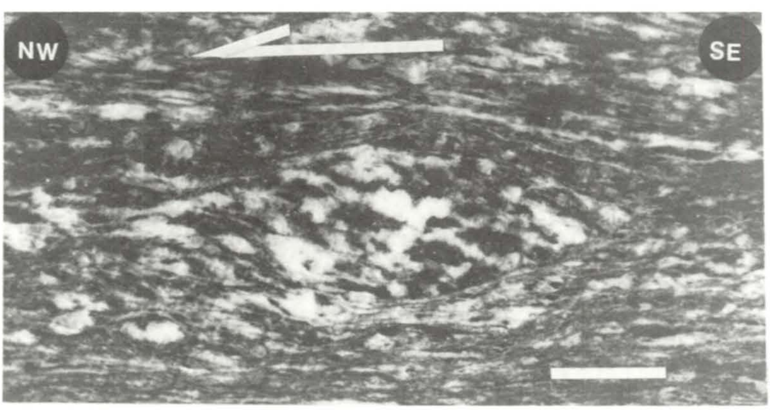

Fig. 36. S-C structure in D2 protomylonite in basement below Hardangerfjord Shear Zone, Tittelsnes (see Fig. 1 for location). Sense of shear is sinistral (top-to-the-WNW). Scale bar = 2cm.

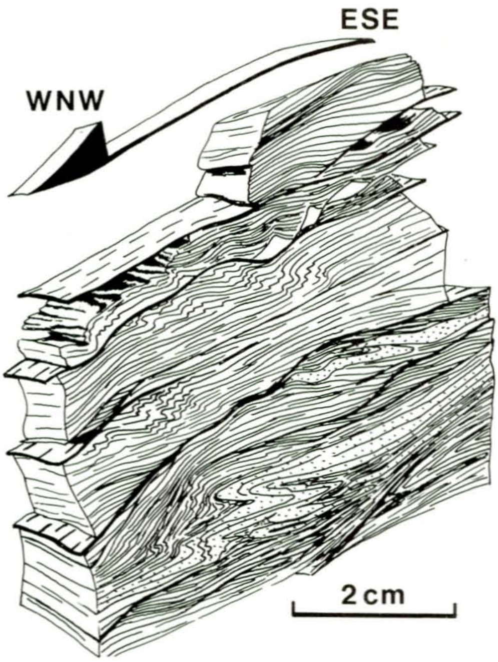

Fig. 37. Illustration of coexisting D2 folds and shear bands in the Hardangerfjord Shear Zone, 2 meters below the phyllite-basement contact (LM555930), indicating top-to-the-WNW shear.

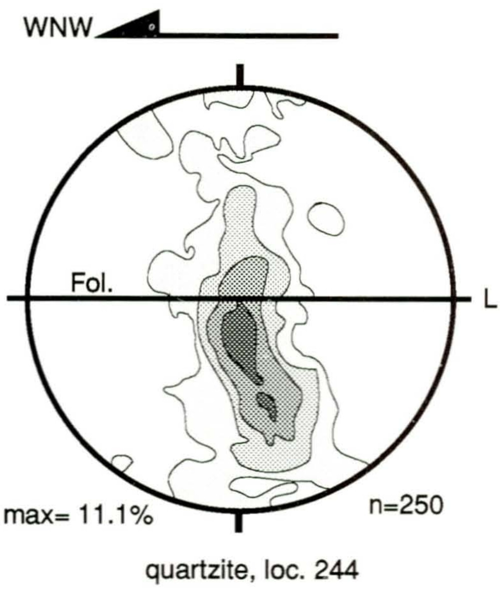

Fig. 38. *c*-axis orientation in quartzite of the Jonstein conglomerate, Ullensvang Group, Hardanger (LM555928). The single girdle is asymmetric with respect to the foliation, and interpreted as having formed during top-to-the-WNW shearing. See Fig. 1 for location.

the phyllites and the basement (Baltoscandian Shield). The fabrics indicate a top-to-the-WNW (extensional) sense of shear along Hardangerfjorden. Seen in the light of recent deep-seismic reflection profiling and interpretations of Hardangerfjorden as a major, possibly crustal-scale shear zone (Hurich & Kristoffersen 1988), it thus appears that the Hardangerfjorden lineament acted as a ductile, extensional shear zone after D1. The Hardangerfjord Shear Zone and the Nordfjord-Sogn Detachment are important features which may help to explain the abrupt change in dip of the décollement zone across the Hardangerfjord. Either a listric geometry of the Hardangerfjord Shear Zone, a footwall uplift related to the giant Nordfjord-Sogn Detachment, or a combination of the two may explain the 10-30° dip of the Bergsdalen Nappes and the southern portion of the Western Gneiss Region. Hence, the Hardangerfjord Shear Zone was active as an extensional D2 shear zone also after the top-to-the-WNW shearing in the Bergsdalen Nappes ceased.

Regional linear patterns

It appears from Fig. 39 that the D1 linear pattern in the Bergsdalen Nappes varies gradually from E-W in the west to a more SE-NW orientation towards the foreland. This variation could be due either to a gradual change in shear direction during D1 or to large-scale, late-D1 or D2 deformation of the linear pattern. Variations from thrust sheet to thrust sheet may indicate that each sheet has a somewhat different deformation history. On a slightly larger scale, the transverse linear pattern in the Bergsdalen Nappes seems to be continuous with the D1 linear pattern in the Øygarden Complex. However, this mostly ESE-trending pattern is clearly discontinuous with the longitudinal lineations in the Lindås Nappe and the outboard terranes of the Major Bergen Arc/Hardangerfjord Group. The recognition of these two broad units of transverse versus longitudinal linear patterns may possibly indicate the presence of two distinct kinematic levels within the orogen, the first (Bergsdalen Nappes, Jotunn Nappe, Øygarden Complex) indicating orogen-normal collisional movements, and the latter and tectonostratigraphically higher unit recording more orogen-parallel movements.

The movement pattern during D2, as determined from D2 lineations (Fig. 40), shows much

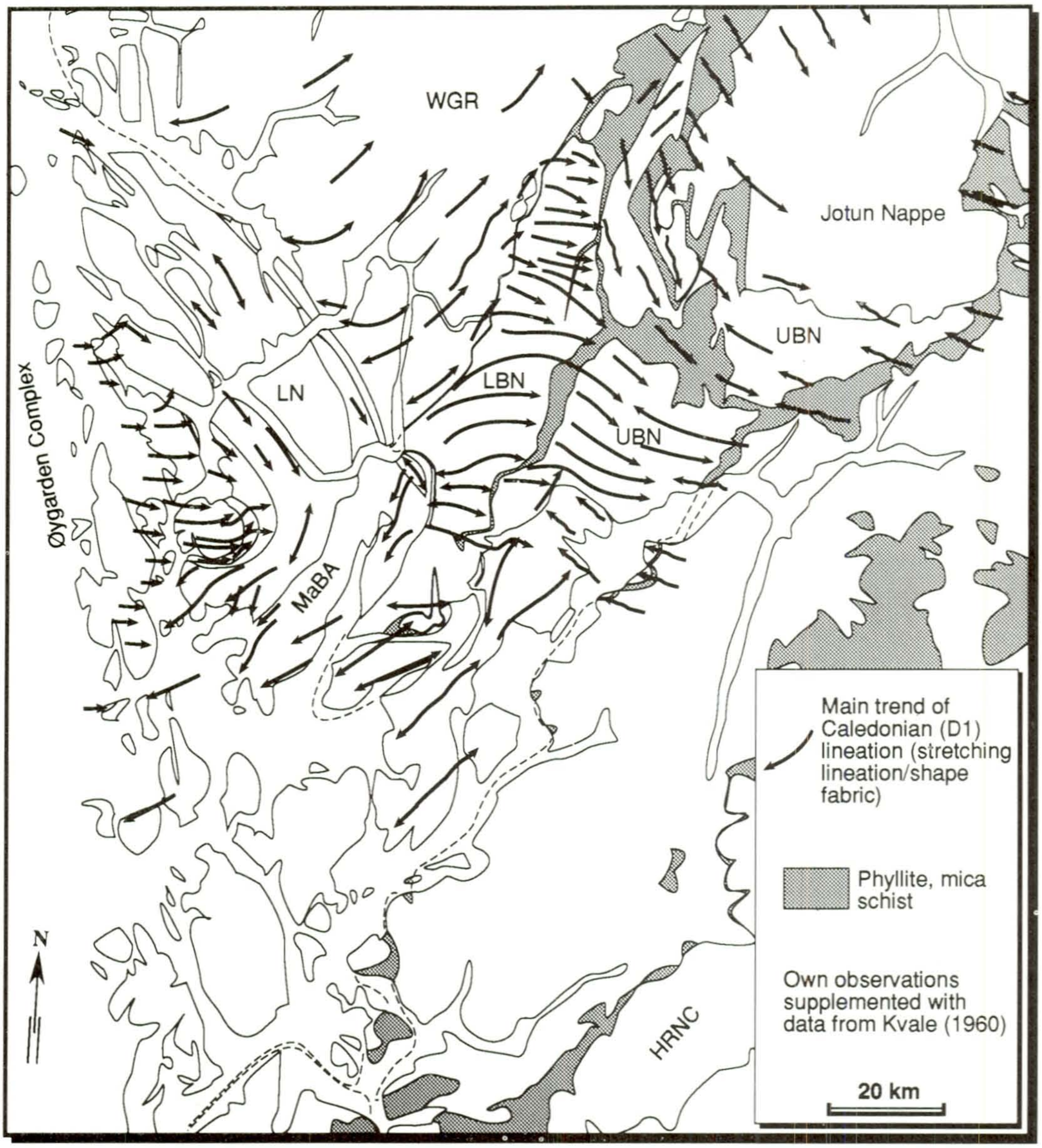

Fig. 39. Map of lineations related to D1 (thrusting) deformation, as determined from overprinting relations and related kinematic indicators. Arrows indicate direction of plunge.

smaller variations than the D1 pattern, probably reflecting a more complex D1 movement history. However, the L2 pattern does show a slight change from a consistent NW-SE trend in the Bergsdalen Nappes and along Hardangerfjorden, to a somewhat more E-W trend in the Øygarden Complex. An E-W trend is also reported from the tectonites below the Devonian basins immediately north of the Bergen Arc System (Chauvet & Séranne 1988), but the reason for this change is not known.

Conclusions

Expressions of two very distinct events of deformation have been found from structural mapping and kinematic analysis in the Bergsdalen Nappes. The first (D1) involved top-to-

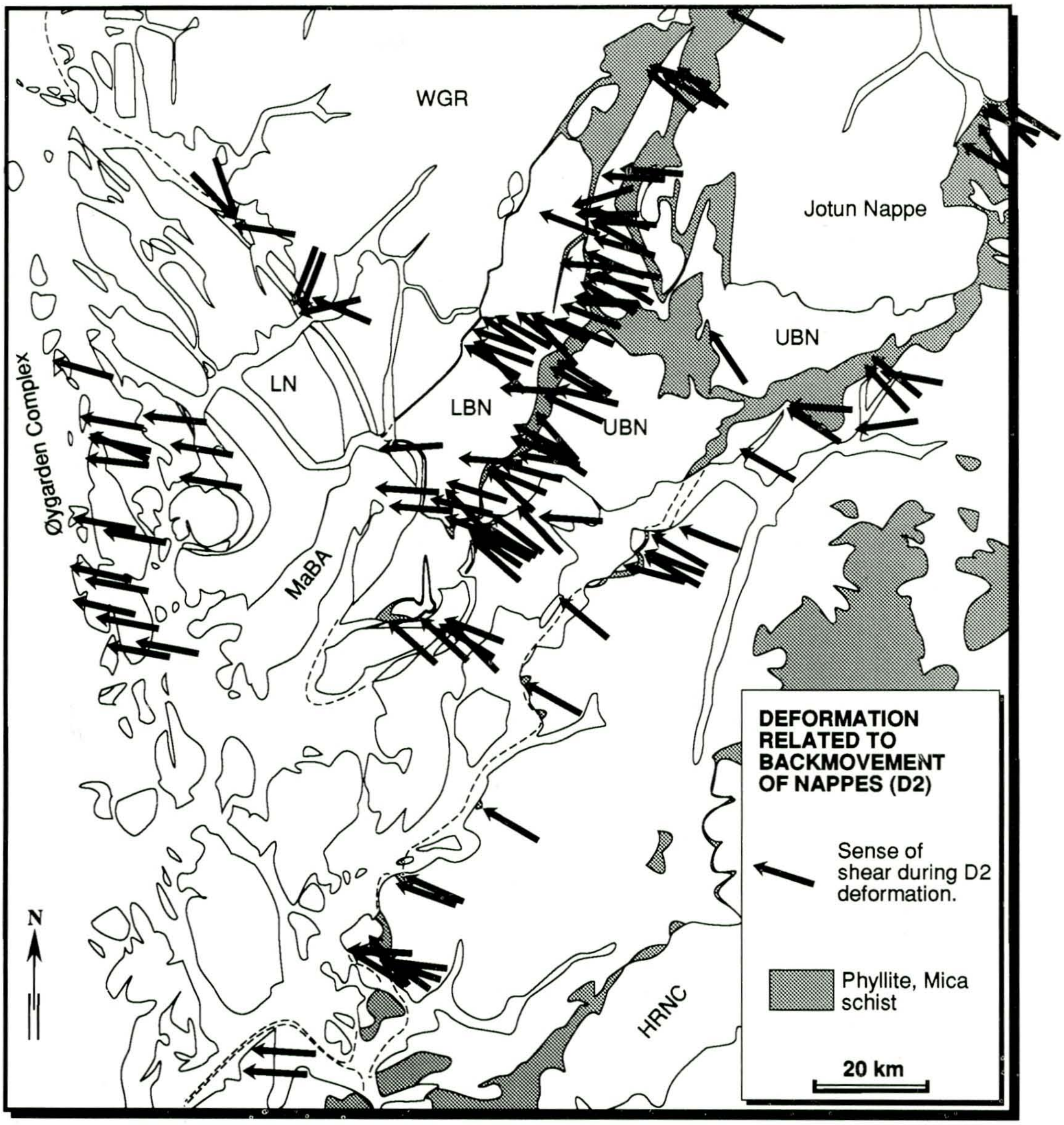

Fig. 40. Map of lineations related to D2 (extensional) deformation, as determined from D2 kinematic indicators. Arrows indicate sense of shear, not direction of plunge.

the-E shear deformation under uppermost greenschist to lowermost amphibolite-facies metamorphic conditions, and is related to the well known thrusting of the overlying Jotun Nappe Complex and other nappes. Pervasive, and locally mylonitic, fabrics formed in the Bergsdalen Nappes during this event, which caused impressive constrictional strains in at least parts of the Bergsdalen Nappes. The succeeding D2 event, which was slightly retrogressive in nature, was less intense than D1. However, although being neglected in earlier work from the region, D2 involved significant non-coaxial deformation related to WNW-directed movement of the overlying Jotun Nappe, and had a significant influence on both the lithological outcrop pattern and the structures present.

The two-fold Lower Paleozoic development in the Bergsdalen Nappes may tentatively beextended to account for the main structural features seen in the surrounding region. Both D1 and D2 were strongly non-coaxial (characterized by asymmetric structures), and the latest expression of D2 is found along the Hardangerfjord Shear Zone which, together with the related Nordfjord-Sogn Detachment to the northwest, is suggested to be responsible for the southeastward rotation of the Bergsdalen Nappes.

Acknowledgements
This study forms part of the authors PhD thesis at the University of Minnesota. The author is grateful to PhD advisers T. B. Holst, P. J. Hudleston, and C. P. Teyssier, and for careful reviews by T.B. Andersen and A.G. Milnes. T. Torske is thanked for kindly introducing the author to the Caledonian geology on Folgefonnshalvøyen. Support for this work was provided by the Norwegian Research Council for Science and the Humanities (NAVF grant no. 440.89/061). Additional support from Torskeklubben, Statoil and the University of Minnesota is acknowledged.

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Manuscript received May 1992; final revised typescript accepted January 1993.

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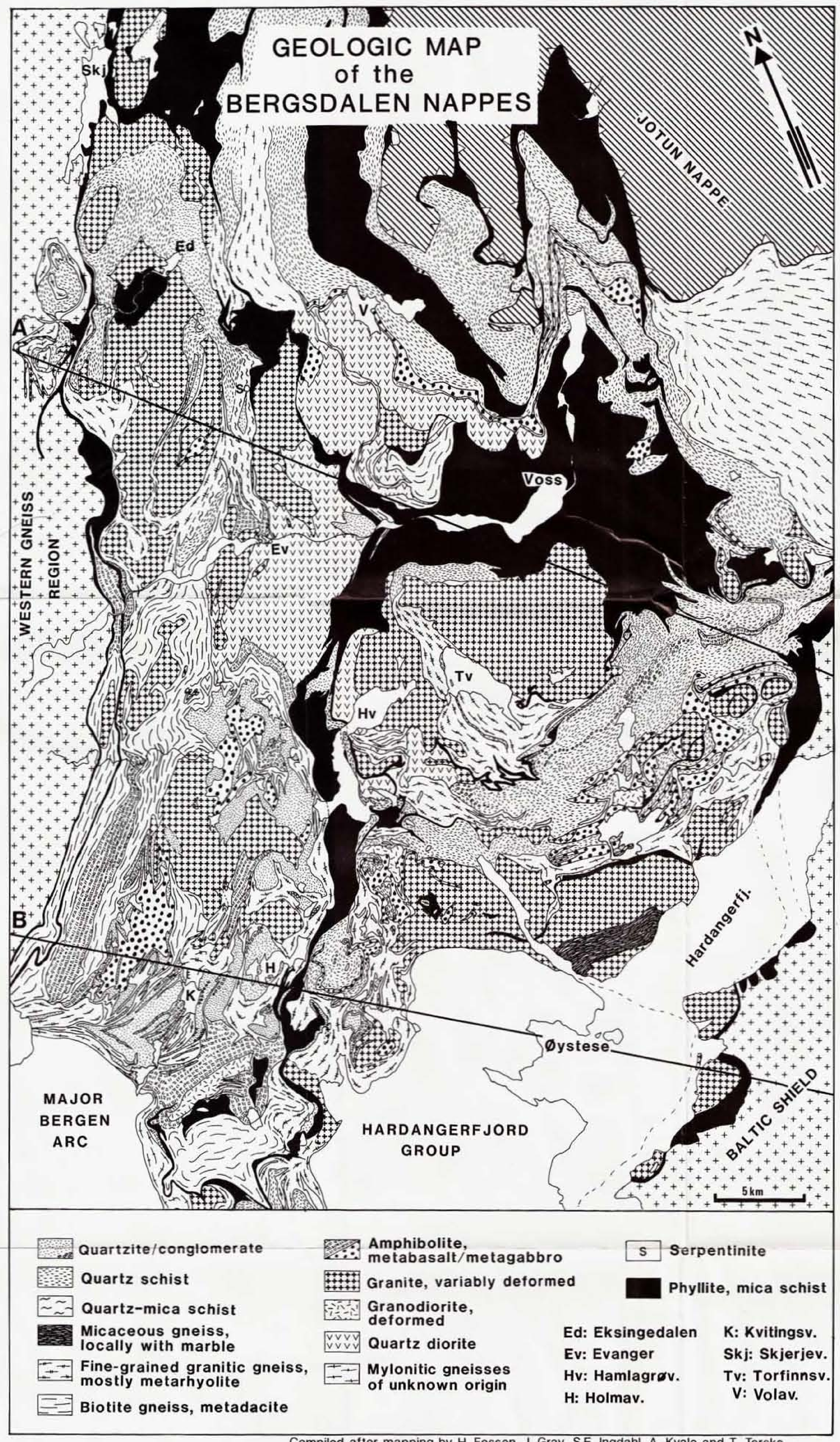

Compiled after mapping by H. Fossen, J. Gray, S.E. Ingdahl, A. Kvale and T. Torske

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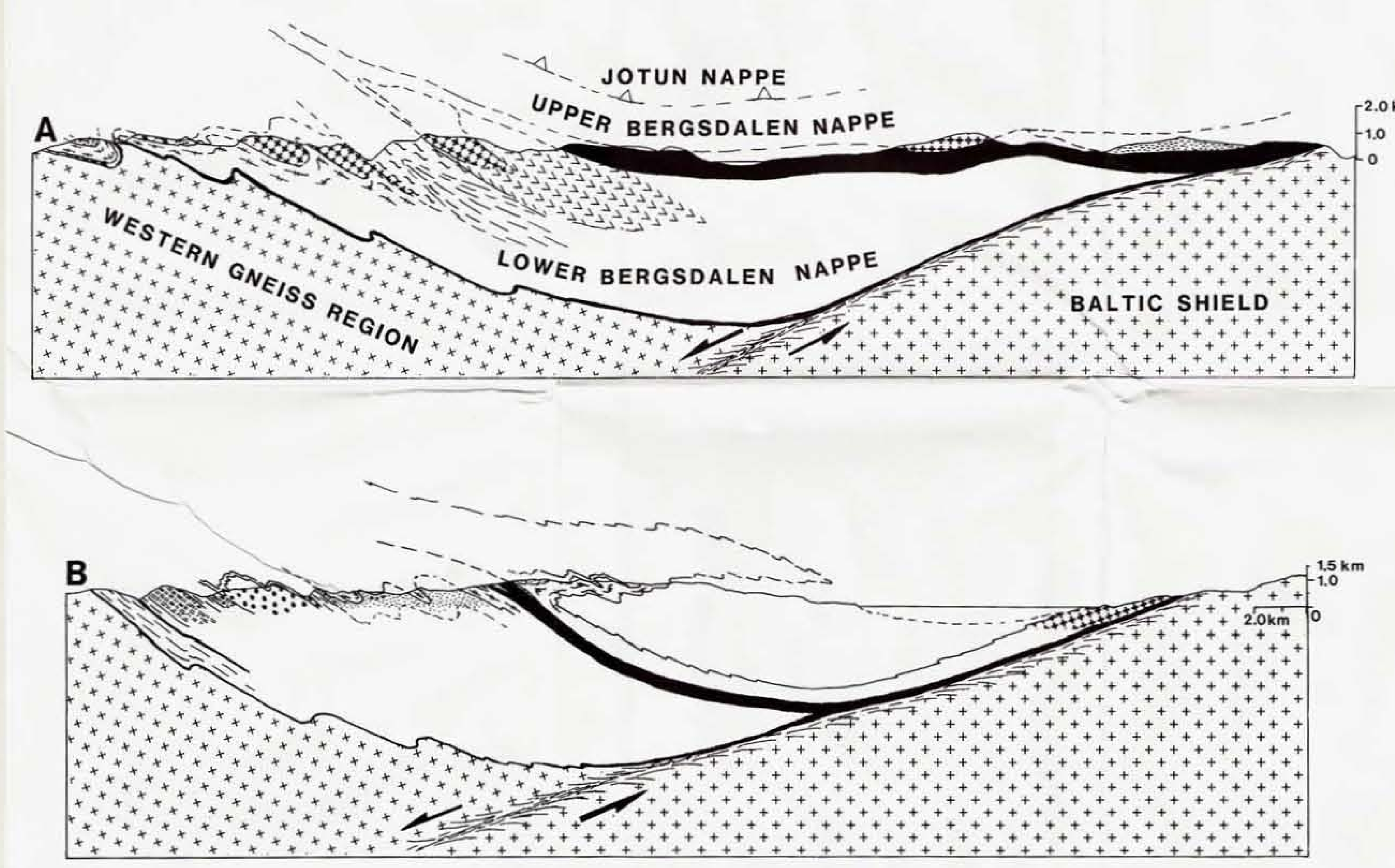
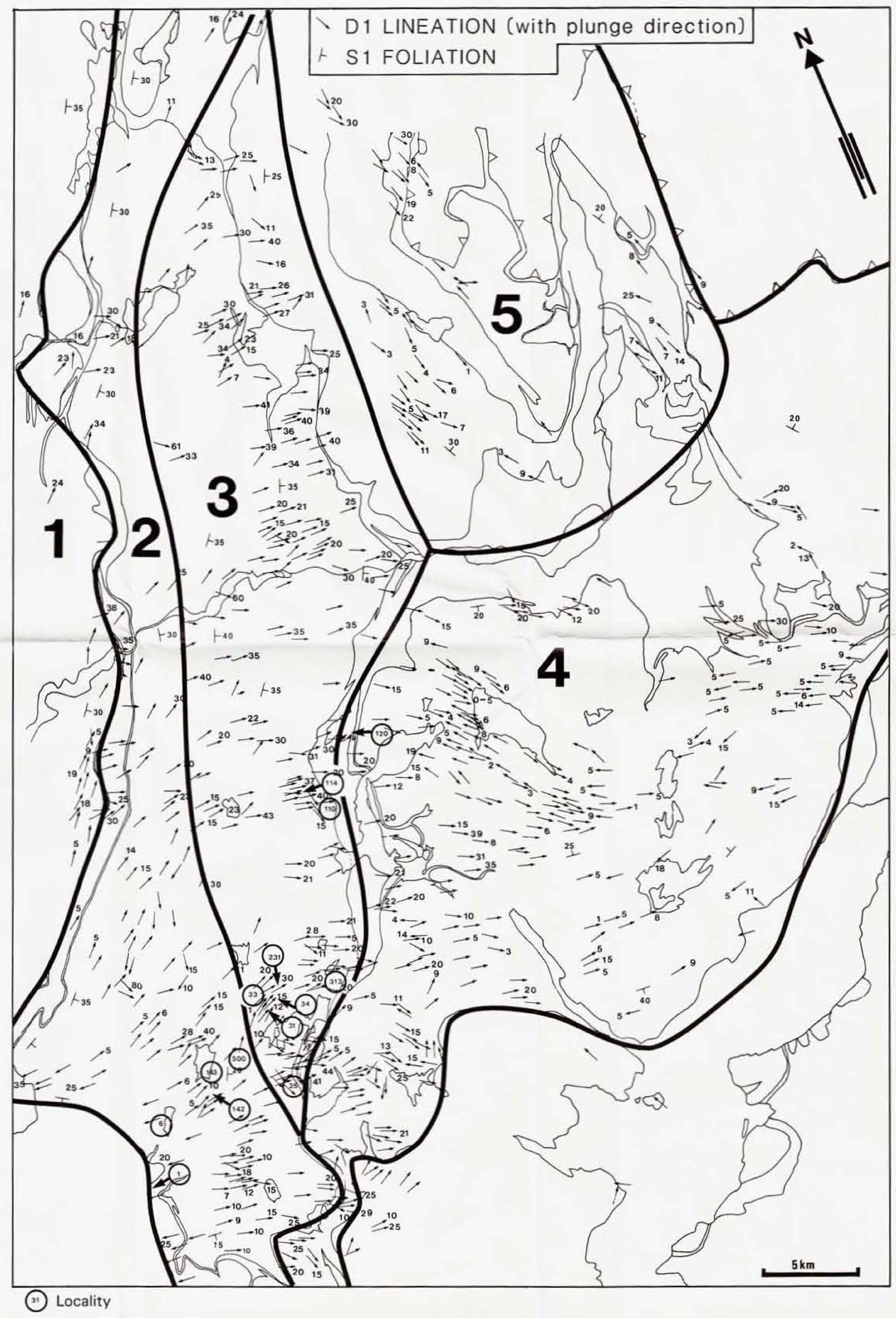


NGU Bulletin 424 - Fossen - Plate II



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