# The style of deformation in the Devonian rocks on Hitra and Smøla, Central Norway

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Bøe, R., Atakan, K. & Sturt, B.A. 1989: The style of deformation in the Devonian rocks on Hitra and Smøla, Central Norway. *Nor. geol. unders. Bull. 414*, 1—19.

The Old Red Sandstone sequences in the Outer Trøndelag Region comprise continental conglomerates, sandstones and mudstones of latest Silurian to Middle Devonian age. During a Late Devonian tectonic phase (Solundian Phase) the deposits were deformed and metamorphosed.
The fold styles on Hitra and Smøla are quite different and probably represent different tectonic levels in a continuous structure. Evidence of mesoscopic thrusting and near bedding-parallel mylonites indicates that the Devonian rocks have been involved in south directed thrusting, which is compatible with deep-seismic reflection data. The metamorphic grade was close to the boundary of the anchizone and lower greenschist facies and pressures in the order of 4.1-4.8 kb are indicated. K-Ar dating of white micas shows that uplift and cooling took place in Late Devonian time.

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## Introduction

The Old Red Sandstone (ORS) (latest Silurian - Middle Devonian) sequence of the Outer Trøndelag region is preserved within a number of coast-parallel ENE-WSW-trending synclines. Outcrops can be traced on a number of small islands to the northwest of Kristiansund via larger islands to the south of Smøla (Edøy region), along the southern coast of Hitra and onto the western part of the Fosen Peninsula (Wolff 1976, Askvik & Rokoengen 1986) (Fig. 1a), thus forming a belt some 145 km in semi-continuous strike-length.

A number of studies have been made of these rocks, from the pioneering works of Schetelig (1913), Reusch (1914) and Vogt (1924, 1929) to more recent studies, essentially of the sedimentary features, by Peacock (1965), Siedlecka & Siedlecki (1972), Siedlecka (1975, 1977), Bøe (1986, 1988, 1989) and Atakan (1988). Less attention, however, has been paid to details of the structural geology, with the exception of generalized descriptions of the major folding on Hitra (Siedlecka & Siedlecki 1972), on Fosen (Siedlecka 1975) and the archipelago south of Smøla (Fediuk 1975, Fediuk & Siedlecki 1977). Details of some minor structures from Hitra have been published by Roberts (1981). During the past four years the authors have made a detailed study of the ORS rocks on Hitra and Smøla (Edøy Region) in relation to their stratigraphic and sedimentological development, deformation and low-grade metamorphism.

The age of the ORS sequences in the Outer Trøndelag Region is fairly well established, though there has been some debate concerning the succession on Hitra. Fossils were found low in the sequence (Member F, Table 1), which were later identified as *Dictyocaris slimoni* Salter and *Hughmilleria* sp., and the succession was assigned a Downtonian age (Reusch 1914, Størmer 1935).

Recent investigations by F. Bockelie (pers. comm. 1987) seem to confirm these results, but it is worth noting that the beds from which the type specimens of *Dictyocaris slimoni* come, in Scotland, are now known to be of late Llandovery or possibly earliest Wenlock age (Bassett 1985). The succession in the Edøy region (Glassøya Formation, Table 2) was established as Early Devonian on the basis of eurypterid fragments (Peacock 1965).

An Early to Middle Devonian age was obtained for the upper part of the succession at Storfosna and on Tristein north of Fosen from fossil plants and spores (Reusch 1914, Vogt 1924, Vogt 1929, Størmer 1935, Høeg 1945, Allen 1976) (Fig. 1a). Although the ORS sequence in the Smøla-Hitra-Fosen region as a whole is thus of Late Silurian to Middle Devonian age, in this account, for convenience, we will refer to the rocks as Devonian.

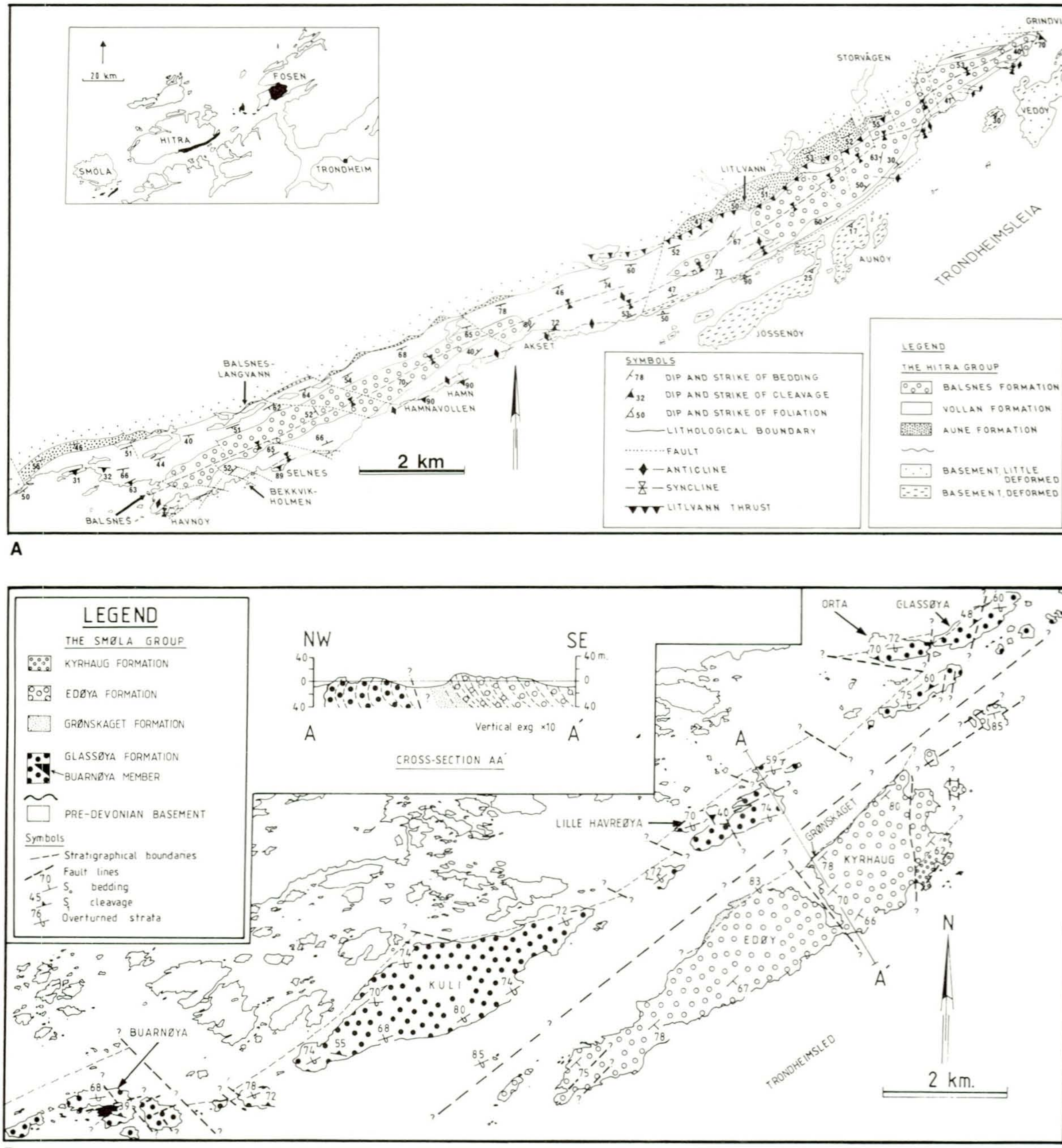


Fig. 1. (a) Geological map of the Hitra Group. Inset map shows the occurrence of ORS deposits in western Central Norway including the Devonian areas on Hitra and southeast of Smøla. (b) Geological map of the Smøla Group.

Table 1. The stratigraphy of the Hitra Group.
The asterix indicates where fossils have been found.

| Formations | Members | Thickness | Lithology |
|---|---|---|---|
| Balsnes Formation | L | ca. 200 m | Conglomerate |
| Vollan Formation | K | ca. 250 m | Alternating sandstones and mudstones |
| | J | 48 m | Mudstone |
| | I | 127 m | Alternating conglomerates and sandstones |
| | H | 12 m | Mudstone |
| | G | 79 m | Alternating sandstones and conglomerates |
| | * F | 157 m | Alternating sandstones and mudstones |
| | E | 40 m | Sandstone |
| | D | 14 m | Alternating sandstones and mudstones |
| | C | ca. 240 m | Conglomeratic sandstones |
| Aune Formation | B | 53–163 m | Alternating red and grey sandstones and siltstones |
| | A | 43–114 m | Basal breccia/ conglomerate |

Table 2. The stratigraphy of the Smøla Group.
The asterix indicates where fossils have been found.

| Formations | Members | Thickness | Lithology |
|---|---|---|---|
| Kyrhaug Formation | | 560 m. | polymict conglomerate and sandstones |
| Edøya Formation | | 1320 m. | polymict conglomerate and sandstones |
| Grønskaget Formation | | 200 m. | alternating oligomict conglomerates and sst. |
| | Buarnøya Member | 130 m. | alternating sandstones siltst. and some cgls. |
| Glassøya * Formation | | 1550 m. | very coarse, polymict conglomerates and occasional sandstones |

## Stratigraphy and sedimentary development

The revised stratigraphies of the Devonian successions on Hitra (Bøe 1986, 1988) and in the Smøla area (Appendix 1) are given in Tables 1 and 2. On Hitra the sediments are referred to as the Hitra Group (Bøe 1988), and on the islands southeast of Smøla as the Smøla Group (Peacock 1965, Fediuk & Siedlecki 1977).

Deposition of the Hitra Group occurred in a continental setting, intermittently in shallow lakes (Siedlecka & Siedlecki 1972, Bøe 1986). The lowermost level preserves a regolith developed on a hilly palaeosurface of diorite (Reusch 1914, Bøe 1989) and locally along the southern margin (north of Bekkvikholmen) on deformed, pre-Late Ordovician metasediments (R. Tucker, pers. comm. 1988). The regolith is overlain by a basal conglomerate (Member A), which along the northern margin grades into a red sandstone/siltstone member (Member B). Together these comprise the Aune Formation, which in turn is overlain by the Vollan Formation. This is an alternation of dark mudstones, sandstones and conglomerates. The top of the preserved section comprises coarse conglomerates of the Balsnes Formation. The entire sequence is approximately 1350 m thick.

The approximately 3600 m-thick sequence in the Smøla area starts with the basal conglomerate of the Glassøya Formation; a very coarse, polymict conglomerate which lies unconformably upon the pre-Devonian rocks filling a somewhat irregular palaeotopography (Atakan 1988). Towards the southeast the coarse conglomerates pass laterally into finer grained varieties, and towards the southwest sandstones and siltstones appear (Table 2). The Glassøya Formation passes up into monomict conglomerates and sandstones of the Grønskaget Formation. This is followed by the Edøya Formation, comprising polymict conglomerates and occasional sandstone lenses, and finally polymict conglomerates of the overlying Kyrhaug Formation. The boundaries between the three uppermost formations are transitional, whilst that between the Glassøya Formation and the Grønskaget Formation is nowhere exposed. The coarse, clastic rocks of the Edøy Region were probably deposited on a series of alluvial fans which coalesced lateral-

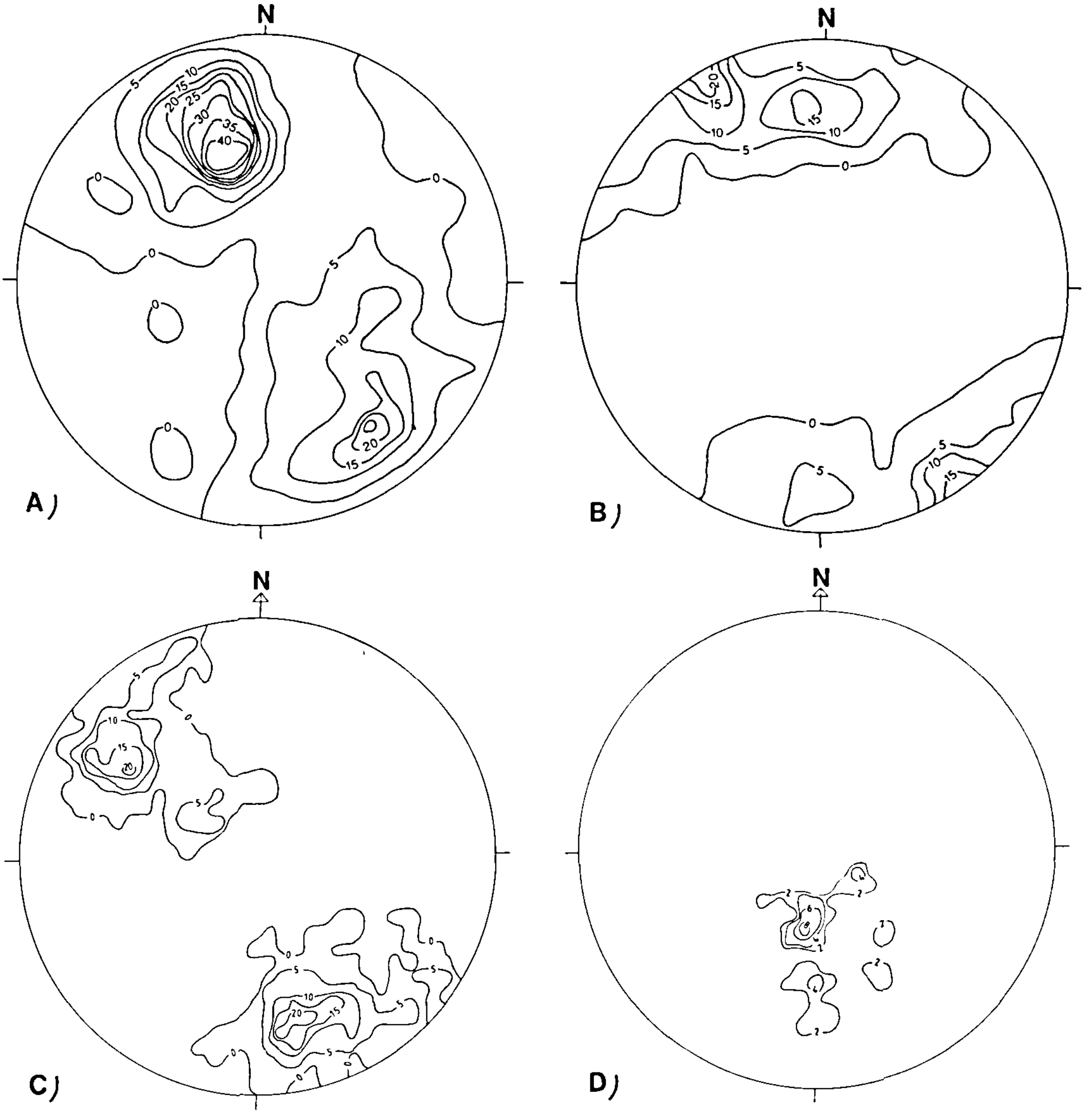


Fig. 2. (a) Stereographic plot of poles to bedding planes in the Hitra Group (354 measurements). (b) Stereographic plot of poles to cleavage planes in the Hitra Group (168 measurements). (c) Stereographic plot of poles to bedding planes in the Smøla Group (360 measurements). (d) Stereographic plot of poles to cleavage planes in the Smøla Group (39 measurements).

ly to build a continuous apron of sediments along a mountain front (Peacock 1965).

Although the general environments of sedimentation during deposition of the Hitra and Smøla Groups appear to have been compatible, it is not known for certain if they are time equivalents. It has been assumed that the Vollan Formation and the Glassøya/Grønskaget Formations are of similar age (Siedlecka 1975), on the basis of eurypterid fragments. It is not clear, however, if the Glassøya/Grønskaget Formations represent a more proximal facies of the Vollan Formation, or if they were deposited in adjacent areas. As on Hitra, no fauna or flora has yet been found in the higher formations of the Smøla Group.

## Structural deformation

Owing to the differences in structural style between the two separated areas of the Hitra and Smøla Groups we have chosen to deal with the deformation of these separately.

### *The deformation of the Hitra Group*

The Hitra Group has been subjected to several phases of deformation. Mylonitic shear zones, reverse faults and a variety of fold structures, as well as a locally strong cleavage, are present. Other structures include a crenulation cleavage and complex fault patterns.

#### *Folding*

The Devonian rocks of coastal Trøndelag are folded in large synclines and anticlines trending approximately parallel to the coast. The structure in which the Hitra Group is preserved consists of two megascopic D1 synclines arranged *en echelon* (Fig. 1a). These folds are generally upright and open to tight, although, the southern flanks may in places be overturned (Siedlecka & Siedlecki 1972). The average trend of the synclinal axes is 060°, and the axial planes are subvertical. The synclines are boat-shaped, but the overall plunge is towards the WSW (Fig. 2a).

On the limbs of the megascopic synclines, especially the southeastern limb of the westernmost syncline, climbing parasitic folds are common (Figs. 3a and 3b). The style of D1 folding is well illustrated at Hamnavollen (613000 28200) where two anticlines climb towards the south (Fig. 4). Their amplitudes are in the order of 60 m, and they plunge approximately 20° towards ENE.

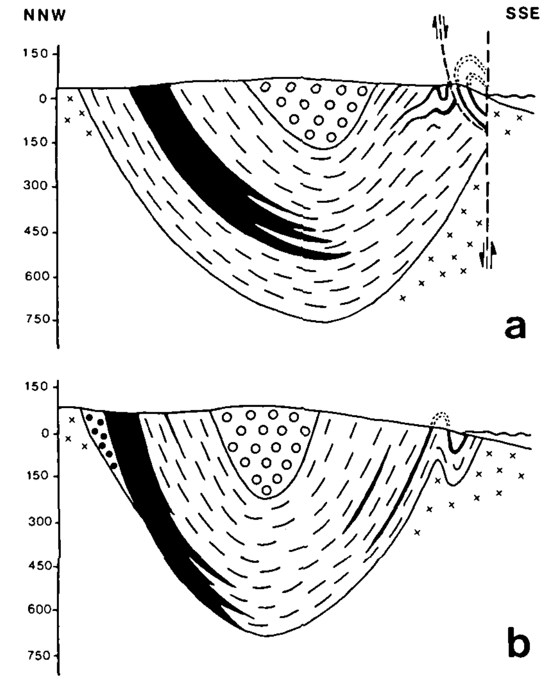


Fig. 3. Cross sections through the Hitra Group. (a) East of Balsneshusvann. (b) At Hamn. Dots: Aune Formation; dashes: Vollan Formation (black: Member F); circles: Balsnes Formation; crosses: basement rocks.

These associated second-order folds are generally open to tight with axes congruous with the megascopic folds, although fold axes are noncylindrical and axial planes curviplanar on a larger scale. Individual folds can be traced along their axes from a few tens of metres up to several kilometres, e.g. at Hamnavollen, west of Sandstad (615300 35000), at

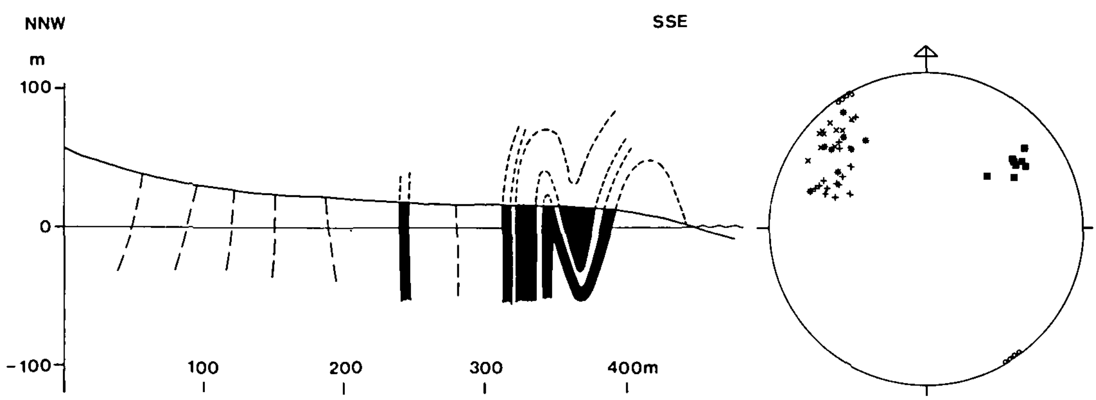


Fig. 4. Structural cross-section at Hamnavollen. Squares: fold axes; circles: poles to slaty cleavage; crosses: poles to kink bands; diagonal crosses: poles to crenulation cleavage; asterix: poles to axial planes of chevron folds.

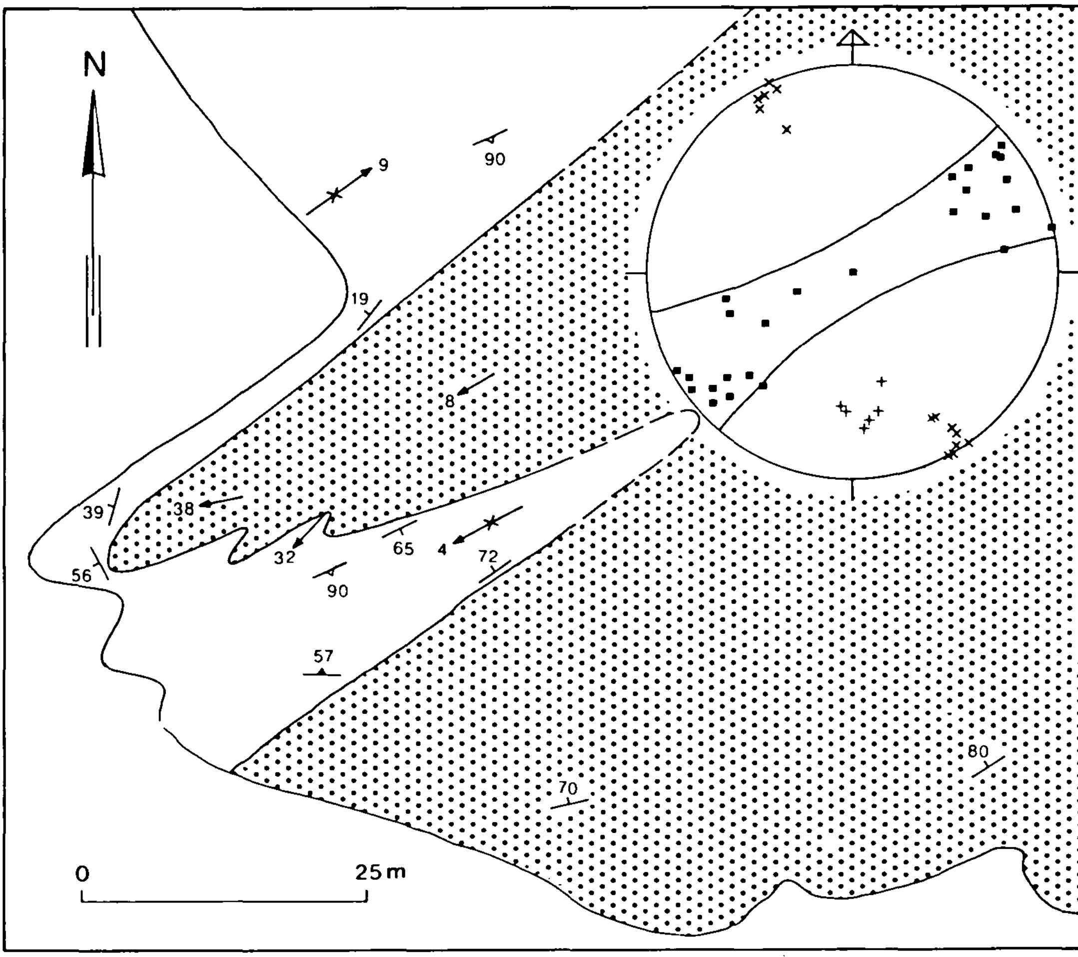


Fig. 5. Structural features at Selnes. All fold axes (squares) fall within a girdle outlined by two small circles. Spaced slaty cleavage (symbol with an open triangle on the map) has approximately the same orientation as the axial planes (diagonal crosses). Horizontal crosses are poles to the crenulation cleavage (symbol with a black triangel on the map).

Havnøy (610000 23000) and northeast of Kalvhaugterna (610350 24000). At Havnøy second-order folds plunge up to 70° towards ESE. 600 m to the west of Litlvannet another set of gentle, open D1 folds is present. The ESE plunge contrasts with the WNW plunge of a similar system observed southwest of Bekkviktjern (611400 25250).

Third-order D1 folds are preferentially developed within finer grained sediments. Their wavelength, amplitude and axial curvature are functions of lithology, being smaller for thin-bedded strata. Both plane and nonplane folds exist, and these may tighten or widen along strike. Most folds have convergent axial plane cleavage fans and belong to Class 1C (Ramsay 1967), but there are also microscopic folds of similar style (Class 2).

The style of folding is well seen at Selnes (610800 25200) (Fig. 5). The maximum recorded amplitude of third-order, climbing folds there is 1 m, but does not normally exceed 15 cm. Folds are markedly noncylindrical, and fold plunges vary through 180°. All fold axes fall within a great-circle striking ENE-WSW (Fig. 5). Individual third-order fold axes display plunge variation up to 130°, and the wavelength of axial curvature usually appears to be more than four times the profile wavelength. In plunge culminations axial curvature is smooth, while in plunge depressions it may be much sharper.

Deformation was apparently more intense in the troughs of second-order synclines. Third-order parasitic folds on the relatively gentle anticlines are congruous with the second-order folds, and the axial surfaces make an angle of about 90° to the mean layering (enveloping surface of second-order folds). The number of third-order folds and the axial curvature increase as profiles tighten and limbs become stretched, and the variation in plunge and azimuth of different fold axes increases.

Marked variation in fold plunge, such as that observed at Selnes, was observed at only one other locality, i.e. at Havnøy. Within other small sub-areas the majority of third-order folds appear cylindrical in nature. This is especially marked at Hamnavollen, Hamn (612500 28900) and Balsneslangvann (611750 24400) Fig. 1a), but fold axes always fall within the girdle outlined at Selnes (Fig. 5). Some of the best exposed folds in the Hitra Group can be observed at Hamnavollen (Fig. 6a).

In the western part of the Devonian outcrop on Hitra there is a system of D2 kink bands and monoclines with NNW-SSE trend and consistent westerly downstep (Roberts 1981, Bøe 1986). Amplitudes of the kink bands are of the order of 5-10 cm, whilst the monoclines reach 40 cm. On Selnes one example of a refolded, tight fold with an axial plane trending 328°/31NE was observed (Fig. 6b), but we are uncertain where to place this in the sequence of deformation.

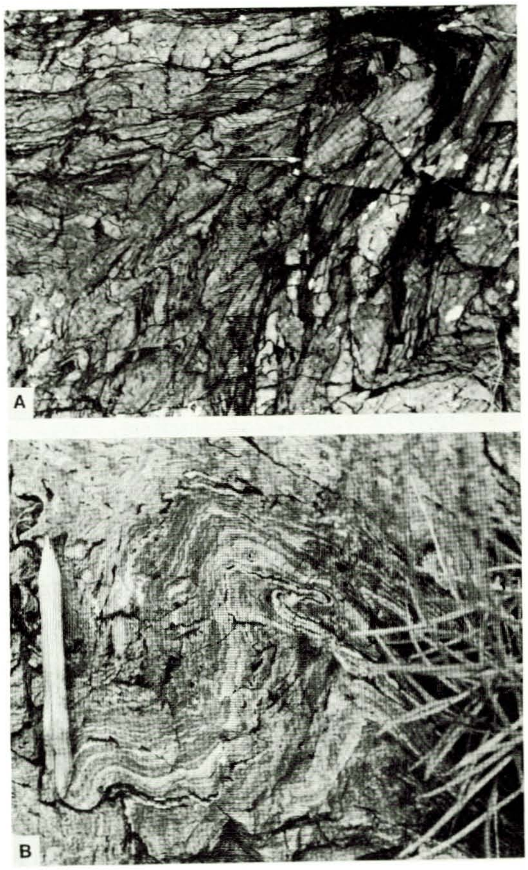


Fig. 6. (a) Fold at Hamnavollen showing Z-, M- and S-folds within the larger, third-order fold. Pencil is 14 cm long. (b) Refolded fold at Selnes. Pencil is 10 cm long.

*Foliation*

Roberts (1981) reported a crude, bedding-parallel solution cleavage in parts of the Hitra Group, i.e. in the vicinity of the Litlevann Thrust and in mudstones of the Vollan Formation (Table 1). Associated with its development there has been a concentration of residual, insoluble minerals, mainly clays and oxides, as solution seams on bedding-parallel surfaces (Fig. 7a). The seams are usually less than 0.1 mm thick but the length, however, may vary from 0.1 mm to several centimetres, and they are concentrated in the fine-grained beds. The solution seams exhibit an anastomosing pattern, thereby wrapping around lenses of coarser siltstone which are enriched in quartz cement (Fig. 7a). The bedding-cleavage either originated during primary compaction (Roberts 1981) or was a result of bedding-plane slip with associated pressure solution during folding and thrusting.

A spaced axial plane cleavage is preferentially developed in mudstones, and is best seen in Member F on the coast southeast of Furuholmen (610950 21500) (Fig. 2b). The cleavage is marked by insoluble residues of clay and oxides concentrated on closely spaced, slightly sinuous, sub-parallel surfaces. It varies from non-planar through planar spaced to a planar, penetrative cleavage (Pique 1982), of which the last-mentioned represents the strongest deformation. The penetrative cleavage is best developed at Selnes, Hamnavollen, Hamn and Akset (613000 29500). In sandstone beds cleavage, where present, is a coarse fracture cleavage.

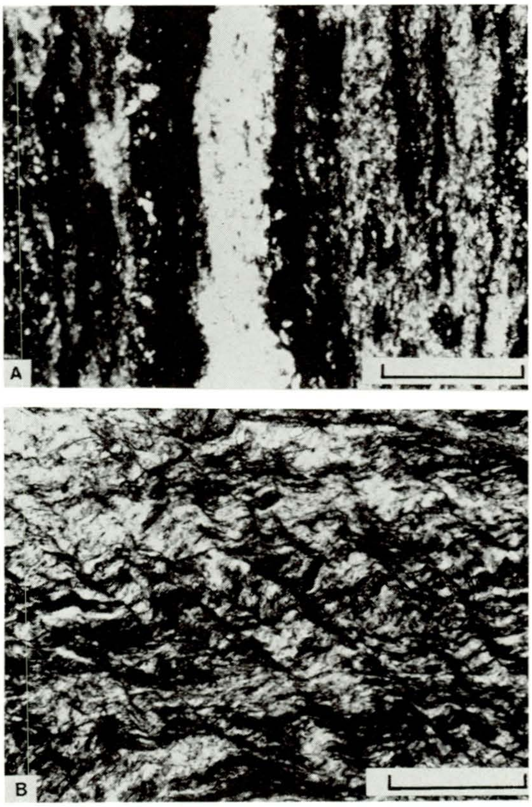


Fig. 7. (a) Bedding-parallel cleavage in Member J on the coast northwest of Balsnes. Note the anastomosing nature of the stylolites. Crossed nicols. Bar scale = 0.5 mm. (b) Crenulation cleavage (trending NW-SE on photograph) at Selnes. The older, axial planar cleavage trends E-W. Plane-polarized light. Bar scale = 0.5 mm.

Sutured contacts and elongation of grains parallel to the cleavage are common, as is the growth of new, fine-grained white mica, especially in siltstones and sandstones.

A bedding-cleavage intersection lineation trends approximately ENE-WSW. Slight variations in plunge between the different localities mirror the weak noncylindrism of the main syncline. During deformation, calcite concretions in mudstones were orientated sub-parallel to the plane of the cleavage and with their long axes striking ENE-WSW (Siedlecka 1977, Roberts 1981).

A well developed crenulation cleavage is observed in mudstones at Hamnavollen and Selnes. This cleavage strikes from E-W to NE-SW (Figs. 4 and 5), making an angle of 20-30° to the older cleavage (Fig. 7b).

Widespread pressure solution features are associated with the crenulation cleavage, and are outlined by dark seams of residual, insoluble minerals. The seams are continuous only over short distances, e.g. 1-2 mm, and at their ends there are transitions into microfolds. Such crenulation-folds reach wavelengths up to 1 mm. Gray & Durney (1979) and Gray (1979) state that both microfolding and solution-deposition processes, in that order, are necessary for crenulation cleavage development, and the crenulation-folds therefore predate the crenulation cleavage. New growth of micas parallel to the crenulation cleavages was not observed. The cleavage is a result only of transfer of dissolved species from the limbs to the hinges of microfolds. This, however, has led to elongation of mineral grains, e.g. quartz and calcite, parallel to the crenulation cleavage.

*Thrusting*

At several localities there is evidence of thrusting, and both brittle and ductile shear zones are present. Indications of movement can be noted on bedding planes, and structures suggesting small-scale ramp-flat geometries have been observed at a number of sites indicating the possibility of, at least locally, stratigraphic repetitions due to thrusting. The only example which can be well documented of such internal stratigraphic repetition is seen southwest of Litlvann (615550 33500 - 616000 34250). Here, in an area 900 m by 175 m at its widest, basement rocks pierce through the sedimentary pile (Figs. 1a and 8). The basement lithologies are mainly granite and granodiorite,

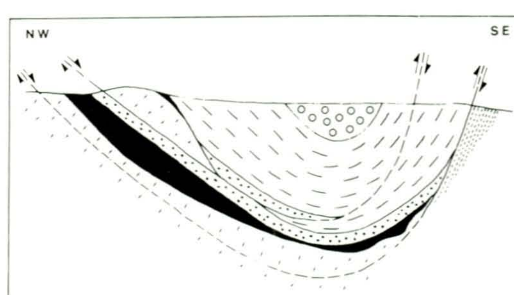


Fig. 8. Cross-section through the Hitra Group southwest of Litlvann. Note the upthrusted basement slice. Crosses: little deformed basement rocks; black: Member A; dots: member B; dashed ornament (long lines): Vollan Formation; circles: Balsnes Formation; dashed ornament (short lines): mylonitized basement rocks.

i.e. similar to the basement substrate north of the unconformity. The basement-wedge overlies Members A and B, and is in turn overlain by Member A which passes up into the Vollan Formation. The presence of Member A above the basement-wedge shows that the stratigraphic repetition and absence of Member B is probably a consequence of both Member B's local development (Bøe 1989) and an unknown amount of transport on the Litlvann thrust.

The northwestern contact of the basement-wedge has been observed over a length of only 0.5 m, exposed by trenching, and is interpreted as a bedding-parallel thrust-fault orientated 060°/65°S. The mudstones of the underlying Member B show a high degree of cataclasis and pressure solution, and are cut by pseudotachylitic veins in various orientations. The bedding laminae are strongly micro-folded, though half a metre beneath the thrust/fault plane significant deformation is no longer apparent. Immediately above the contact the rocks are strongly cataclastic for at least 1 m. All mineral grains are crushed, and variably intruded by veins of pseudotachylite. Some recrystallization occurred, with the growth of large amounts of secondary epidote and chlorite. No prominent lineation has been observed at the contact to give an unequivocal movement direction. Associated microfolds, however, have axes essentially strike-parallel and verge down-dip, implying an approximately N to S movement vector.

The map pattern and structure indicate a minimum southerly displacement of 300 m, though it may be considerably more (Fig. 8).

Because of the degree of overburden it is difficult to trace the Litlvann Thrust eastwards beyond Litlvann; however, to the northeast of Litlvann mapping shows a considerable thinning of the Vollan Formation which cannot be ascribed to sedimentary variation. We have thus extended the Litlvann Thrust along the upper boundary of Member B. At the gravel quarry to the southeast of Strandavann (617000 36000) the contact of Member B with the Vollan Formation is strongly deformed, and small folds show a movement sense from north to south, i.e. a down-dip vector. Close to the contact the siltstones and mudstones of Member B have a tight-packed solution cleavage sub-parallel to the bedding. Thin-section study shows that the matrix quartz has in part recrystallized into polygonal moza-

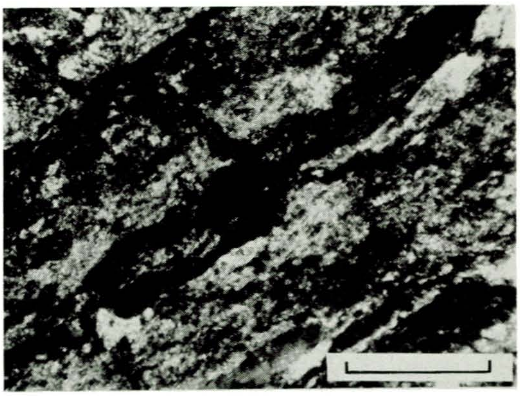

Fig. 9. Mylonitized diorite from the northern coast of Aunøy. Crossed nicols. Bar scale = 0.5 mm.

ics, and fine-grained white mica (<0.01 mm) is abundantly developed. Small porphyroblasts of chlorite and epidote can also be observed. These features imply that the shear-zone fabric is compatible with the metamorphic temperature experienced by the Devonian rocks.

Beyond the observed northeastward extension of Member B, owing to almost complete cover it is not possible to show whether the thrust ramps up into the Vollan Formation or if it continues into basement.

The bedding-parallel nature of the Litlvann Thrust and its steep inclination (65°) suggests that the thrust was folded during the main D1 folding.

Small-scale, ramp structures were observed at several other localities within the Hitra Group. Detailed logging (Bøe 1986), however, has revealed no significant repetition of the sequence.

Within sandstone beds sporadic green lenses and strings of pseudotachylite are locally observed parallel or subparallel to bedding. Such lenses are usually straight, reaching a maximum thickness of 1-2 cm and lengths of 15-20 cm. Strings are much thinner, undulating and thread-like, and appear to be injected from the pseudotachylitic lenses into the surrounding rock. The pseudotachylite is devitrified, and floating within the fine-grained matrix are clasts of strongly deformed quartz and minor amounts of epidote, chlorite, feldspar, sphene and rock fragments. Cataclasite zones up to 0.5 m thick have also been observed. Due to the bedding-parallel nature of the pseudotachylites and the cataclasites it is concluded that they originated either during thrusting or as a result of flexural slip during folding.

The best preserved mylonite zone occurs at the contact between basement and the Devonian on the coast northwest of Jøssenøy (Figs. 1a and 9). The mylonite zone is sharply defined and is approximately 15 m thick, although the thickness varies along strike. The foliation is sub-parallel to bedding, and an intersection lineation plunges 16° towards 286°.

The basement rocks in the area comprise strongly foliated diorites intruded by granites and basic sheets.

There are large variations in the strains over short distances, e.g. transitions from relatively undeformed diorite to blastomylonite over distances of 0.5 m.

The mylonitic rocks are developed from both basement and cover litholigies, and are essentially ultramylonites and blastomylonites which have suffered later brittle reworking, producing brecciation and cataclasis.

In the blastomylonites, quartz is partly recrystallized to polygonal mozaics and ribbon quartz textures are to be observed in some thin-sections. Minute white micas have formed in the matrix, from relict white micas and fractured feldspars. Porphyroblasts of chlorite and epidote are also recorded. Again, the paragenesis of the blastomylonites is compatible with the metamorphic grade of the Devonian.

The westernmost exposure of strongly deformed diorite is on an island 2 km to the west of Jøssenøy (612350 32200). Where basement rocks are observed in contact with the Devonian at Bekkvikholmen and Kalvhaugterna they are relatively undeformed, and the mylonite zone probably continues further to the south, i.e. now rooted in the basement.

Shear zones were also observed in the basement to the north of the Devonian outcrop, e.g. northeast of Storvågen, where approximately 1 m of mylonites dip towards the southeast.

The question may well be asked whether the mylonite zones represent a feature of the strike-slip regime during the evolution of the Møre-Trøndelag Fault Zone (MTFZ). However, owing to the bedding-parallel nature of the zones and their compatability with the metamorphic temperatures affecting the Devonian, implying that they are part of the integrated structural metamorphic development of the rocks, we prefer to interpret this as being related to the movements producing the Livann Thrust. It has to be admitted, however, that no unequivocal evidence for the movement vector is available.

## *The deformation of the Smøla Group*

The structures in the Devonian rocks of the Edøy Region are relatively simple compared with those observed in the Hitra Group. Following deposition the rocks were folded into an asymmetrical syncline trending NE-SW, with an axial surface dipping moderately towards the northwest, and an overall plunge towards the northeast.

This structure can be traced across the area, and is accompanied by an axial planar cleavage in the finer grained lithologies. Later structures include several generations of faults and joints trending NW-SE, NE-SW and N-S.

The Smøla Group lies unconformably on plutonic (predominantly dioritic), volcanic and volcanoclastic rocks in the Smøla Region. The primary contact, which can be traced along the northwestern margin of the area, is tectonically inverted, and dips steeply towards the northwest (Fig. 1b).

The relations between the Glassøya Formation and the underlying basement rocks are best observed along the northwest coast of Glassøya (Fig. 10). The pre-Devonian rocks are essentially volcanoclastics, deformed conglomerates and younger, basic dykes. The volcanoclastic rocks vary from fine- to coarse-grained, and the finer grained fractions are usually well bedded, containing sedimentary structures, e.g. normal grading, cross-bedding and flame structures.

These rocks were folded and cleaved prior to deposition of the Devonian. A secondary crenulation cleavage is only developed in the conglomeratic basement rocks, e.g. at the northeastern end of the pre-Devonian outcrops at Glassøya (Fig. 10). The basic dykes were intruded prior to deposition of the Glassøya Formation.

At most other localities, except at Orta and Glassøya, the pre-Devonian rocks are plutonic. The deformation history of these rocks is not easily deciphered because of their massive character. However, along the northern margin all basement rocks were folded and inverted during deformation of the Devonian.

A simplified block diagram showing the unconformable relationship between basement

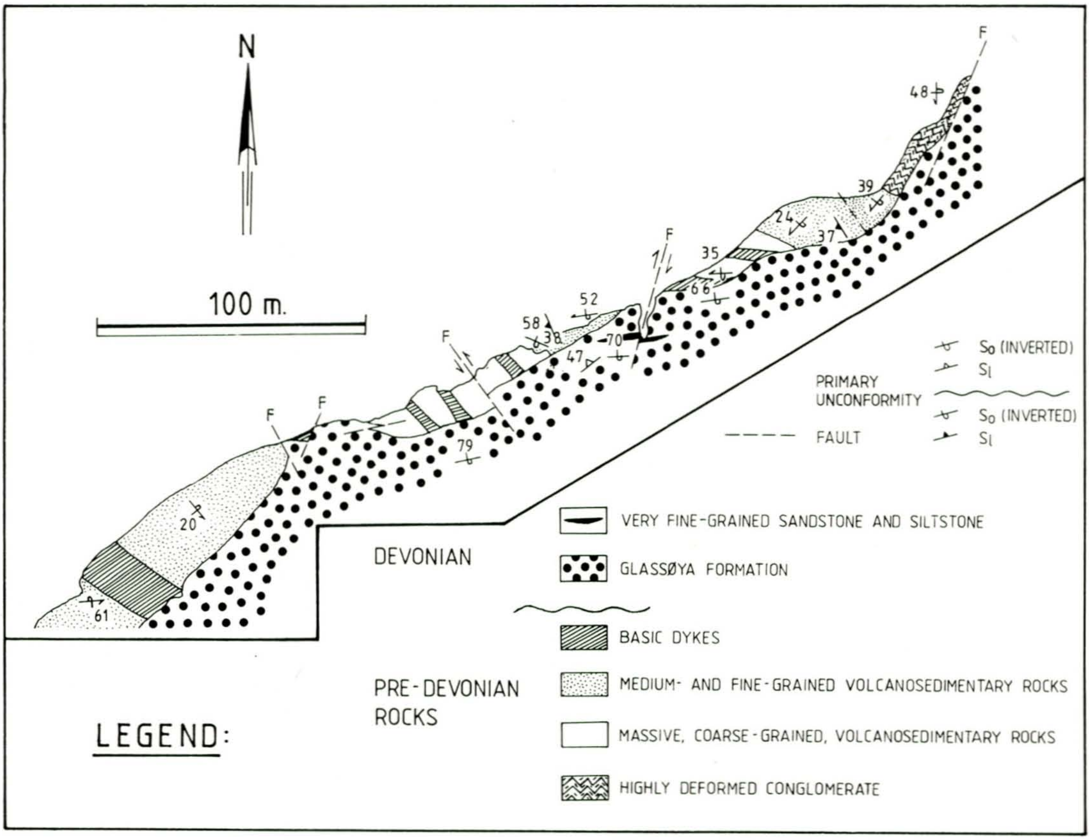


Fig. 10. Map showing the contact relationships between the Glassøya Formation and the basement along the northern coast of Glassøya.

rocks and the Glassøya Formation on the northwest coast of Glassøya is shown in Fig. 11.

The original boundaries of the Devonian sedimentary basin can nowhere be observed, as the strata are now folded and eroded. A previous model inferred a marginal, syndepositional fault along the northwestern boundary of the preserved Devonian rocks (Steel et al. 1985), but this cannot be sustained.

Bedding in the Glassøya Formation is usually inverted, dipping towards the northwest, while in the Grønskaget Formation bedding is more or less vertical. The Edøya and Kyrhaug Formations, on the other hand, show a gradual change from vertical to steep southeasterly dips (Fig. 12). This marks the gradual change from the overturned northwestern limb to the right-way-up southeastern limb of the syncline. A stereographic plot of bedding shows the

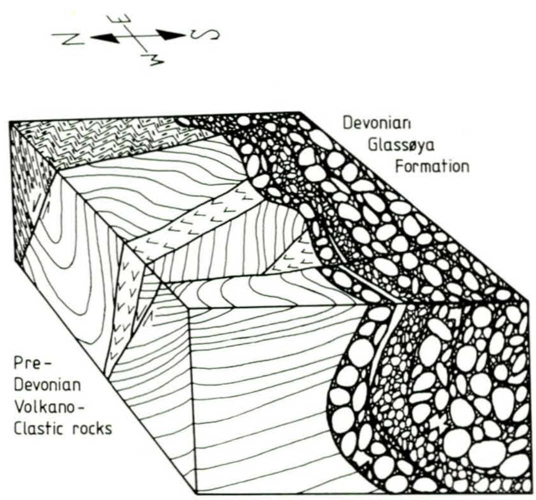


Fig. 11. Simplified block diagram showing the unconformable field relationship between the Glassøya Formation and pre-Devonian, basement rocks along the northern coast of Glassøya.

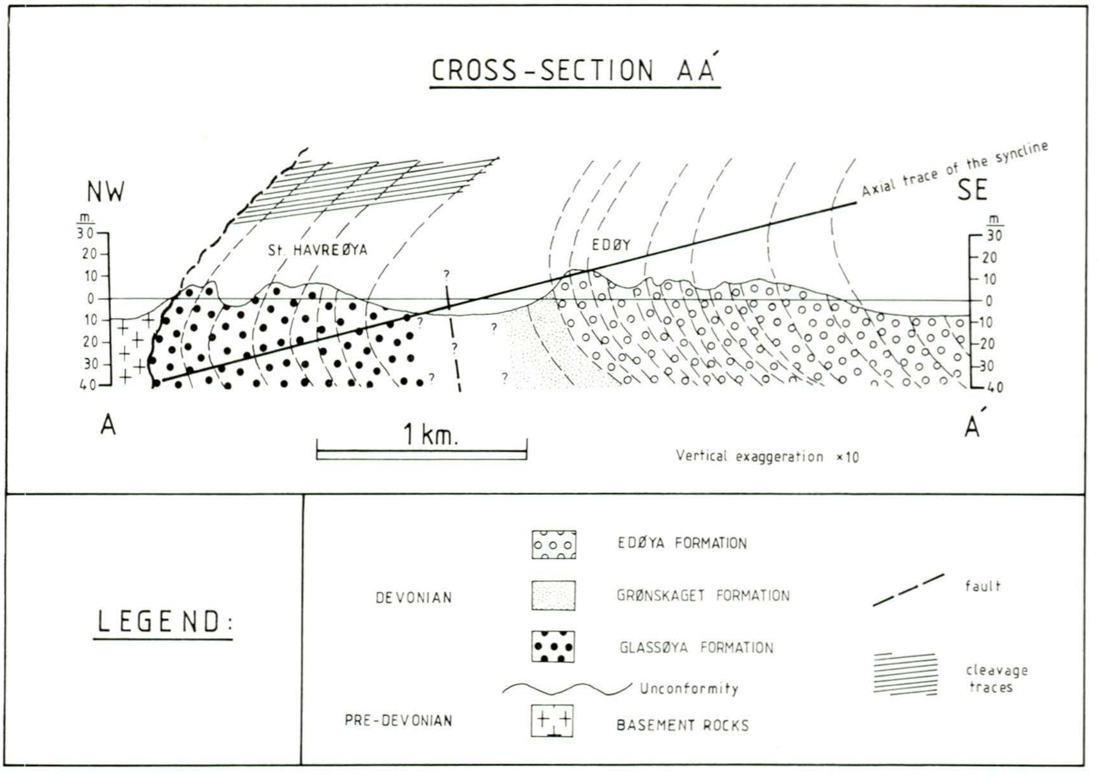


Fig. 12. NW-SE oriented cross-section through the Smøla Group. Note that the southeastern limb of the syncline is exposed at Edøya.

approximate positions of the two limbs (Fig. 2c), and the asymmetrical character of the syncline is obvious from the cross-section in Fig. 12. The syncline is nonplane, noncylindrical, and can be classified as Class 1C (Ramsay 1967) with a covergent axial plane cleavage fan. The axial surface dips moderately towards the northwest, and the axis plunges essentially northeastwards.

In the overturned northwestern limb, cleavage is only sporadically developed in finer grained lithologies such as siltstones. The cleavage in the Glassøya Formation (Fig. 2d) dips more gently than the associated bedding plane, confirming the inversion of the sequence. Magnetic fabric data, from the coarser lithologies in the Edøy region, show a cleavage-parallel orientation of the principal magnetic plane (Torsvik et al. 1989).

## Faulting

The region described lies within the MTFZ, which has been active probably since Late Devonian times (Torsvik et al. 1989). We do not regard a discussion of this faulting as within the purview of this paper, and merely present rose diagrams for the macroscopic, sub-vertical faults that cut the folded Devonian strata (Fig. 13).

## Metamorphism

During deformation of the Hitra and Smøla Groups there was extensive ductile deformation and recrystallization of quartz. This is best seen in finer grained mudstones and siltstones, where grains are elongated parallel to the S1 cleavage.

The quartz has a stronger cleavage-parallel alignment when large quantities of clastic mica are present. This is presumably the result of extensive pressure solution at quartz-mica interfaces. Dimensional orientation of quartz also resulted from growth parallel to the foliation.

In the hinge region of folds the tendency for a preferred orientation of quartz grains is much weaker, and new, relatively strain-free grains have developed. Weakly developed pressure shadows around quartz and feldspar are common, and in these very fine-grained granular quartz and fine-grained white micas are developed.

Feldspar is variably altered and the most common products are fine-grained white mica (possibly phengite), epidote, zoisite and calcite. All these, except zoisite, also crystallized in the matrix, and in cases of very strong alteration it is difficult to differentiate between matrix and original feldspar grains. Pressure solution of feldspar has occurred, and many grains are elongated parallel to the rock cleavage, *especially in contact with muscovite.* Growth of new feldspar directly on the parent crystal is outlined by dirty rims. Fractured feldspar grains are in places healed by unstrained feldspar that grew after the main deformation.

Epidote is among the most common low-grade metamorphic minerals, usually as a very fine-grained alteration product after feldspar, and is abundant in the matrix. However, there are sporadic clusters of equidimensional or prismatic well formed porphyroblasts, up to 3 mm long, with no preferred orientation.

Chlorite is also a commonly developed metamorphic mineral. It is the common retrograde product after biotite, while most of the clastic amphiboles are replaced by chlorite. At Selnes chlorite-white mica aggregates were observed. White micas are strongly deformed, whilst chlorites exhibit a varying degree of deformation, suggesting continuous growth. Growth of chlorite has also occurred parallel to the (001)-planes of white mica.

Within the finer grained lithologies of the Hitra Group minute flakes of white mica grew during deformation. They are especially abundant in the micro-folded mudstones along the southern margin of the area, and are also common in pressure shadows. Commonly, the (001)-planes are sub-parallel to cleavage. Minute, white micas have also grown parallel to grain boundaries, e.g. between clastic mica/quartz and feldspar/quartz. Electron-microprobe analyses of micas in sandstone samples from the Edøy Region and XRD analyses of clay minerals (illite crystallinity: Kübler 1967, Hoffmann & Hower 1979, Kisch 1983) in siltstone samples have been made. Electron-microprobe studies were carried out on micas from four thin-sections of sandstones of the Buarnøya Member (Table 2), and three different categories of micas were identified:

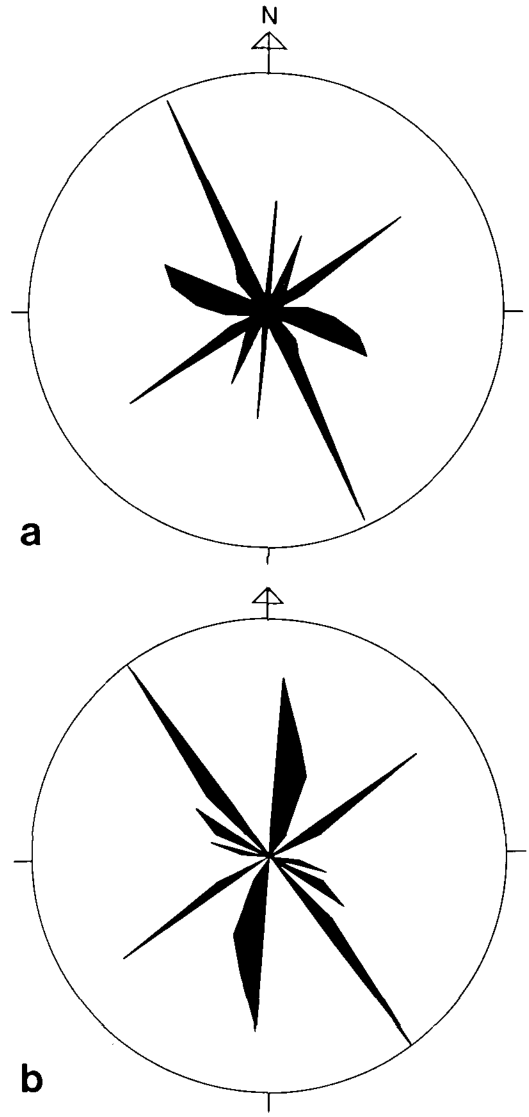


Fig. 13. Azimuth directions of major, subvertical faults. (a) Hitra Group (86 measurements). (b) Smøla Group (55 measurements).

i) Small phengite grains growing in intergranular areas (New-Phengites I) (6 analyses).

ii) Phengites growing on clastic White Micas (New-Phengites II) (8 analyses).
iii) Detrital White-Micas (12 analyses).

The analytical results are presented in Table 3. Major element analyses of the New-Phengites (I and II) are seen to ble clearly distinct from those of detrital White-Mica compositions. This is also observed on the AFK diagram, where the groups display distinct clusters of data points (Fig. 14a) (see also White et al. 1985). The difficulty of placing the electron beam in the centres of these minute grains should be emphasized, and in some cases the beam may have inadvertently been partly, or at worst wholly, in the host clastic grain. The schnitt-effect also probably contributes to the scatter of the New-Phengite II analyses. Group 1 (New-Phengites I) shows a better concentration of data points than Group 2 (New-Phengites II). This is probably because Group 2 phengites developed at the expense of detrital White-Micas, and their chemistry reflects in part the micas from which they are derived (possibly due to the schnitt-effect).

The Group 1 phengites, on the other hand, have a composition which has not been influenced by the clastic micas, since they grew in intergranular areas and are therefore the more reliable as metamorphic grade indicators. The detrital White-Micas and the New-Phengites are plotted in a P/T diagram (Fig. 14b) (Velde 1967). Taking the average of their $Si^{4+}$ contents based on the assumption of $O_{10}(OH)_2$ per formula unit, New-Phengites I and II and the detrital White-Micas give two separate stability curves; though, in itself, this diagram does not give precise pressure or temperature estimates (see below).

Table 3. Comparison of the element compositions of the New-Phengites I&II, and detrital White-Micas (Oxide wt.%).

| | *New-Phengites I (intergranular areas) (6 Analyses)* | | *New-Phengites II (growing on clastic white-micas) (8 Analyses)* | | *Detrital White-Micas (12 Analyses)* | |
|---|---|---|---|---|---|---|
| | Mean | Std.dev. | Mean | Std.dev. | Mean | Std.dev. |
| $SiO_2$ | 49.84 | 0.85 | 49.51 | 2.47 | 46.61 | 0.64 |
| $Al_2O_3$ | 26.65 | 1.43 | 27.44 | 1.72 | 32.88 | 1.44 |
| $TiO_2$ | 0.03 | 0.03 | 0.17 | 0.14 | 0.44 | 0.30 |
| FeO* | 3.16 | 0.40 | 2.87 | 0.61 | 1.97 | 0.24 |
| MnO | 0.04 | 0.03 | 0.05 | 0.03 | 0.04 | 0.02 |
| MgO | 2.60 | 0.28 | 2.16 | 0.96 | 1.05 | 0.40 |
| CaO | 0.04 | 0.03 | 0.06 | 0.05 | 0.03 | 0.04 |
| $Na_2O$ | 0.10 | 0.08 | 0.18 | 0.18 | 0.53 | 0.15 |
| $K_2O$ | 9.44 | 0.34 | 9.32 | 0.98 | 9.41 | 0.56 |
| Total | 91.90 | | 91.76 | | 92.95 | |
| Formula proportions assuming $O_{10}$ $(OH)_2$ per formula unit. | | | | | | |
| Si | 3.44 | 0.06 | 3.41 | 0.12 | 3.17 | 0.05 |
| Al | 2.17 | 0.10 | 2.23 | 0.17 | 2.64 | 0.10 |
| Ti | – | – | – | – | 0.04 | 0.02 |
| Fe | 0.19 | 0.02 | 0.17 | 0.03 | 0.11 | 0.01 |
| Mn | – | – | – | – | – | – |
| Mg | 0.27 | 0.03 | 0.22 | 0.10 | 0.11 | 0.04 |
| Ca | – | – | – | – | – | – |
| Na | 0.02 | 0.01 | – | – | 0.14 | 0.04 |
| K | 0.83 | 0.02 | 0.82 | 0.09 | 0.82 | 0.05 |

* FeO as Fe total.
(Original data in Atakan, 1988)

XRD analyses based on the illite crystallinity index of Kübler (1967) were carried out, and crystallinity values ($\Delta 2\Theta°CuK\alpha$) obtained (Kübler 1968, 1984 p. 578). The experimental conditions are as suggested by Kisch (1983 p. 348). After ethylene glycol (EG) treatment illite peaks were approximately 26% narrower due to removal of interstitial $H_2O$ in illite/smectite mixed layers. The results of the EG treated samples vary from 0.58° to 0.20° with an avarage value of 0.29° (st.dev. = 0.11), and this indicates high anchizone/low epizone boundary conditions (approx. 300-350°C) (Kisch 1983 Table 5-IV, Kisch 1987 p. 293, Niedermayr et al. 1984).

The areal distribution of the results is displayed in Fig. 14c. It should be noted that the distribution of data points with high illite crystallinity (i.e. low $\Delta 2\Theta°CuK\alpha$ values) appears to be independent of stratigraphic level, as the highest values (0.20°) are generally found in the uppermost stratigraphic units. This implies that the crystallinity of illites was not directly related to burial.

The temperature estimates deduced from the XRD analyses of phengites may be used on the phengite-stability curves obtained from the electron-microprobe analyses of micas. Assuming 300-350° for the high anchizone/low epizone boundary, the stability curve for the New-Phengites indicates approximate pressures in the range 4.1-4.8 kb (Fig. 14b).

Thus, the Devonian rocks of the Hitra and Smøla Groups have undergone syn-tectonic metamorphism close to the anchizone/greenschist-facies boundary. An upper time limit for this metamorphism is given by K-Ar ages of 365 ± 17 and 382 ± 18 Ma from white micas in the Smøla Group (J.G. Mitchell, pers.comm. 1987).

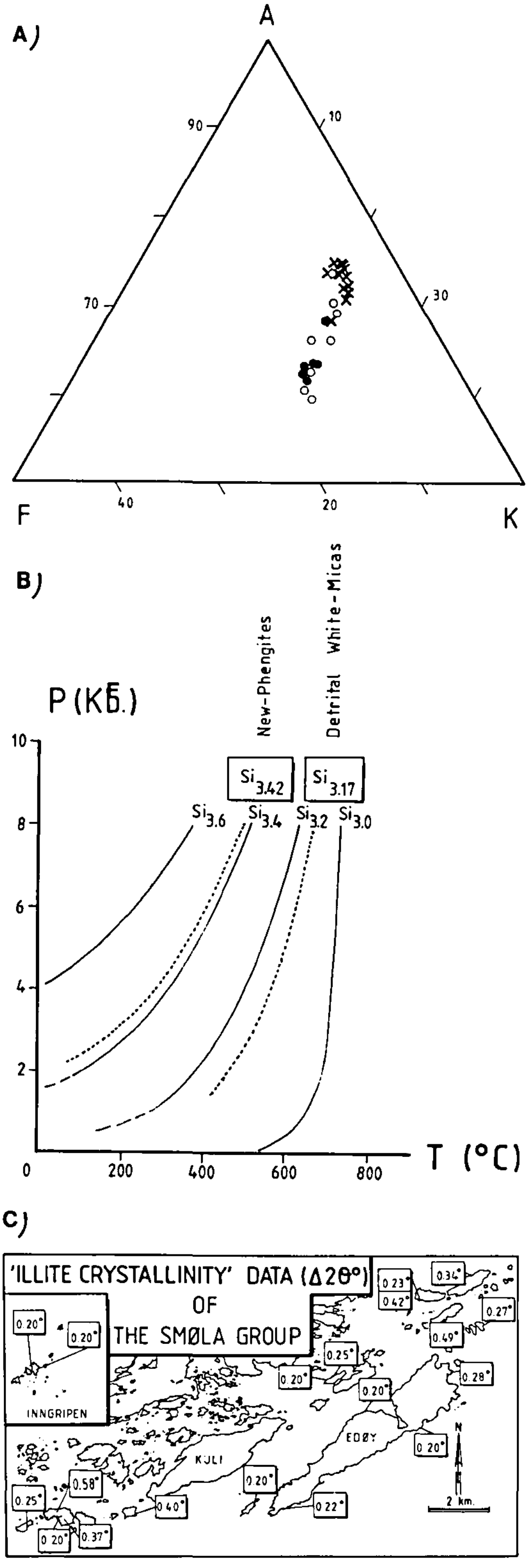


Fig. 14. (a) AFK-diagram showing the compositional trends of the detrital White-Micas and the New-Phengites I and II. Note that detrital White-Micas and New-Phengites I display distinct clusters of data points. (A = $Al_2O_3/SiO_2$; F = $(FeO+MgO)/SiO_2$; K = $K_2O/SiO_2$). Crosses: detrital White-Micas; Dots: New-Phengites I; Circles: New Phengites II. (b) Stability curves of the various phengite micas (after Velde 1967). The compositions of each mica are represented by the number of $Si^{4+}$ ions present, thus $Si_{3.42} = K((Mg_{0.58}\ Al_{1.42})(Si_{3.42}\ Al_{0.58})\ O_{10}\ (OH)_2)$ and $Si_{3.17} = K((Mg_{0.83}\ Al_{1.17})(Si_{3.17}\ Al_{0.83})\ O_{10}\ (OH)_2)$. The two dottet lines represent the analysed detrital White-Micas and New-Phengites. For the New-Phengite curve, the average value of the New-Phengites I and II is used. (c) Map showing the distribution of the illite crystallinity data in the Smøla Group.

# Discussion

The present study has revealed important differences in style between the major fold structures developed on Hitra and the island complex of the Edøy Region southeast of Smøla. The pattern on Hitra is of two major *en échelon* synclines which have subvertical axial planes. The island complex of the Edøy region, on the other hand, lies entirely in the partly inverted northwestern limb of a major syncline with a fairly gentle northwestly dipping axial plane. The exposed section in the northwestern limb of this structure measures almost 4 km in the cross-strike direction implying that the syncline has a considerable, though as yet unknown amplitude.

The reason for this major change in style along what is apparently a continuous trend is not immediately obvious. There are two possibilities that we can envisage in a continuous belt situation:

i) That the structural style changes radically along strike but at the same level.
ii) That the structural style changes in depth and that we are observing different depth profiles in the two areas.

The latter feature is fairly common in orogenic belts, as remarked by de Sitter (1956) where he showed how many fold structures have a disharmonic form in the depth profile. Such style changes with depth are well documented from Alpine ranges (e.g. Lemoine 1978) and indeed from many other orogenic belts. If this was the case, the situation can be envisaged where the Edøy Syncline represents the deeper and Hitra the higher tectonic level with the axial surface having an upward concave form. Recent shallow-seismic profiling carried out by two of the authors (R.B. and B.A.S.) indicates the presence of an approximately N-S trending fault between Hitra and Smøla, but the exact nature of the fault has not been established.

The establishment of the Litlvann Thrust with its indicated south-directed movement is relevant in the above context as the sense of overturning of the Edøy Syncline is also towards the south, which is compatible. It is of considerable interest that the Mobil Search deep-seismic profile line ILP-7 shows ascending reflectors at 18 and 10 km depth at a distance of 20 km to the northwest of Hitra. The upper of these zones, when projected southeastwards to the southern coast of Hitra would lie at a depth of 3.5-4.0 km (C. Hurich, pers.comm. 1988). This, combined with the terrane characteristics of Hitra-Smøla (Sturt & Roberts 1987), may imply a major thrust boundary to a nappe at that level, of which Hitra and Smøla are surface expressions.

The Litlvann Thrust and associated essentially bedding-parallel zones of mylonitic/blastomylonitic rocks may represent the early product of the compressional stress system leading to the formation of the fold structures, and are in turn perhaps an expression of the southerly movement of a major nappe which includes the Devonian strata, although we would stress the speculative nature of this model. It is of interest that in the Fosen area there is evidence of Devonian, southwest-directed, subhorizontal thrusting (Piasecki & Cliff 1988).

The present study also demonstrates how the Devonian rocks of the Hitra-Smøla region have been affected by syn-folding, low-grade metamorphism close to the boundary of the anchizone and the lower part of the greenschist facies. This is a feature observed in other areas of Devonian rocks in western Norway and western Shetland. The uplift/cooling ages (K-Ar) on white micas from the Edøy region and the age of the sediments in the belt along Trondheimsleia indicate that this metamorphism must have occurred during the interval between the Middle Devonian and the base of the Carboniferous. The uplift-cooling ages fit well with the preliminary palaeomagnetic results from rocks of the Hitra Group (Appendix 2) which show that the Late Devonian/Early Carboniferous pole is post-tectonic, i.e. post-folding. The cause of the widespread low-grade metamorphism of the West Norwegian and West Shetland Devonian is a major problem which has yet to be resolved.

In our model the major strike-slip movements along the MTFZ proposed by many authors would effectively post-date the folding and thrusting of the Hitra-Smøla Devonian. We are aware that subsequent modifications by transpressional deformation may have taken place.

*Acknowledgements*

We acknowledge financial assistance from Elf Aquitaine Norge A/S and Norges geologiske undersøkelse. We wish to thank Dr. J.G. Mitchell for making available the K-Ar ages on white micas. We are grateful to Drs. D.M. Ramsay, D. Roberts and T.H. Torsvik for fruitful discussions and to I. Lundquist and F. Haugan for help with the illustrations.

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# Appendix 1

The stratigraphy of the Devonian rocks on the Island complex of Smøla has previously been described by Fediuk & Siedlecki (1977). The rocks were defined as the Smøla Group, comprising two formations; the Basal Kuliøy Formation and the younger Edøy Formation, which was again subdivided into three members (lower, middle and upper). A further internal subdivision of the Smøla Group (after Peacock (1965)) was published by Siedlecka (1975), and later used by Steel et al. (1985).

The stratigraphy of the Smøla Group is here revised. The new nomenclature and the reason for abandoning certain names and introducing some new ones are given below. There are four mappable lithostratigraphic units (formations) in the area, and one member embedded within the basal formation. Each unit is well exposed in its type section.

The revised nomenclature is defined according to the code of the Norwegian Commitee on Stratigraphy (Nystuen (ed.) 1986). Grid references refer to map-sheets: BKL 123124-20, Kuli; BMN 123124-20, Edøya; BMN 125126-20, Lyngværet, Økonomisk Kartverk.

## *The Glassøya Formation*

*Name:* Glassøya is an island situated in the northeastern part of the Devonian south of Smøla. The name was introduced by Peacock (1965).

*Type section:* On the northern coast of Glassøya, where there are good exposures (MR-6160 2475).

*Thickness:* The average thickness of the Glassøya Formation is estimated to be approximately 1550 m. The minimum preserved thickness is measured on Henningsholmen, and is approximately 860 m. The maximum preserved thickness is measured from the small skerry between the islands Svelunn and Kuli (MR-5265 1954) to the small islands Håskjæra exposed in the Kulisvaet between the islands Kuli and Edøy (MR-5405 1770) and is approximately 2320 m.

*Lithology:* The Glassøya Formation is a very coarse-grained, clast-supported, polymict conglomerate with boulders reaching several metres in diameter, and dominated by clasts of dioritic, granitic and basic volcanic rocks. There are also minor amounts of felsic volcanics, gabbroic, porphyritic and metamorphic rock-types. The matrix consists of coarse- to medium-grained sandstone.

A basal breccia is present at several localities (MR-5688 2131) (MR-5215 1980). It is predominantly composed of angular blocks of diorite, basic volcanics and porphyritic rocks mixed in a red, silty matrix containing subangular to angular clasts of varied origin, mainly diorites, gabbros, volcanic and metamorphic rock fragments. In areas close to the unconformity the clasts typically show a bimodal roundness distribution, with large, angular, volcanic fragments and subrounded to rounded boulders of diorite, gabbro, felsic and basic volcanics and red igneous and metamorphic rocks. Upwards all clasts become rounded to well-rounded.

The matrix content ranges from 3-14% with an average of 7.8%. Average clast size is 41 cm (St.dev. = 36.2). Sorting is generally very poor, with particle sizes ranging from a few centimetres to several metres, and there is a general decrease in grain size upwards. Some finer grained conglomerate beds show clast imbrication. More than 60% of the beds are ungraded, and the average bed thickness is 185 cm (St.dev. = 145). Several sandstone and sandstone-siltstone lenses occur throughout the Glassøya Formation. These vary in thickness from a few centimetres to several metres.

*Fossils:* Peacock (1965) identified fragments of eurypterids of possible Downtonian age from a sandstone-siltstone lens on the southern coast of Lauvøya (MR-6046 2321).

*Boundaries:* The base of the unit is an angular unconformity separating the Glassøya Formation from the pre-Devonian rocks. The top of the unit is nowhere exposed, and the contact relationships with the overlying Grønskaget Formation are not known.

*Distribution:* Outcrops of the Glassøya Formation are scattered along a belt of islands stretching from Remningene in the northeast (MR-6287 2541) to Blåsværet in the southwest (MR-4681 1667). The major islands in this belt are Glassøya, Orta, Lauvøya, Lille Havreøya, Store Havreøya, Kuli, Arnøya, Buarnøya and Blåsværet.

*Subdivision:* On Buarnøya, coarse conglomerates of the Glassøya Formation pass laterally into finer grained conglomerates and sandstone-siltstone facies. This lateral facies change represent a distinct unit, the Buarnøya Member.

## *The Buarnøya Member*

*Name:* Buarnøya is an island situated in the southwestern part of the Devonian south of Smøla. New name.

*Type section:* Along the western coast of Buarnøya (MR-4810 1695 to MR-4813 1678).

*Thickness:* The thickness of the Buarnøya Member is approximately 130 m at the type locality. The unit probably wedges out towards the east.

*Lithology:* The Buarnøya Member consists of alternating layers of sandstones, with minor conglomerate and siltstone layers in between. Major conglomerate beds are restricted to the basal parts of the sequence, where clasts reach 30-40 cm in diameter. Most of the sandstone layers show fining-upward facies motifs, and have erosional, lower contacts. Sedimentary structures, e.g. horizontal bedding and cross-stratification, are common. The grain size varies from layer to layer, ranging from very fine-grained to coarse-grained sandstone and sometimes to pebbly sandstone. The petrology of these grey to brown sandstones indicates a lithic greywacke composition. The bed thickness varies from a few centimetres to a few metres.

*Boundaries:* The Buarnøya Member is embedded within the Glassøya formation, and wedges out towards the east. The upper boundary is transitional within a few metres, whereas the lower boundary is sharp.

*Distribution:* The Buarnøya Member is restricted to the island of Buarnøya and some small skerries to the southwest of this. A possible west-southwestwards extension is likely.

## *The Grønskaget Formation*

*Name:* Grønskaget is situated on the northwestern side of Edøya. New name.

*Type section:* The Grønskaget Formation is best exposed along the northwestern coast of Edøya (MR-5895 2073 to MR-5924 2104).

*Thickness:* The thickness of the Grønskaget Formation is approximately 200 m.

*Lithology:* The Grønskaget Formation comprises alternating layers of sandstones and conglomerates. Weathered diorite constitutes up to 80% of the clast population in the conglomerates, whilst other igneous rocks and minor quartzite clasts occur in smaller amounts. The maximum particle size ranges from 4 cm to 30 cm, whilst bed thickness varies from 10 cm to 2 m. Individual beds show well defined, more or less sharp boundaries, and are usually ungraded. Some normal graded beds and a few beds where only the base is inversely graded also occur. Clasts are rounded to subrounded, in places subangular, and commonly imbricated. Sorting is generally poor.

The matrix is generally a medium- to coarse-grained, grey to greyish-brown sandstone, and constitutes up to 40% of the conglomerates with an average at 15%. Sandstones and thin siltstones occur sporadically and at some localities, they constitute the dominant facies. The sandstone layers are usually horizontally bedded or planar cross-bedded.

*Boundaries:* The lower boundary of the Grønskaget Formation against the Glassøya Formation lies somewhere along the Kulisvaet, but is nowhere exposed. The upper boundary is transitional towards the overlying Edøya Formation, and the two formations show an interfingering relationship. The previous suggestion of an intra-Devonian unconformity at this boundary (Steel et al. 1985) is therefore not sustained.

*Distribution:* The Grønskaget Formation is restricted to a narrow strip of land along the northwestern coast of Edøya.

## *The Edøya Formation*

*Name:* Edøya is the largest island with Devonian rocks south of Smøla. The name Edøya was introduced by Peacock (1965), and has later been used by other authors.

*Type section:* The type section of the Edøya Formation is along Edøyvalen at Edøy (MR-5900 2070 to MR-5935 1998).

*Thickness:* The thickness of the Edøya Formation varies from 390 m to 1320 m. At the type locality it measures 1320 m.

*Lithology:* The Edøya Formation is composed mainly of polymict conglomerates with scattered sandstone lenses. The dominant clast types in the conglomerates are green-

coloured sandstone and reddish granite. Other clast types include grey sandstones, diorites, quartz porphyries, tuffs and some metamorphic rock fragments. Maximum particle size ranges from 4 to 40 cm, with an average of approximately 16 cm (St.dev. = 7.2). Clasts are usually rounded to well rounded. Most of the beds are poorly sorted, but show well developed clast imbrication. Bed boundaries are well defined by either erosional surfaces or abrubt changes in grain size. The average thickness of conglomerate beds is 108 cm. Approximately 50% of the beds are ungraded, whilst 25% show normal grading. Other grading types also occur. The matrix is usually composed of medium- to coarse-grained, brown-grey sandstone, and constitutes 10-12% of the conglomerates. Sandstone lenses usually contain horizontally bedded, brownish-grey to grey, medium- to coarse-grained sandstone. The thickness of the sandstone beds varies from a few centimetres to several metres. Planar cross-bedding, trough cross-bedding, channels and scour-and-fill structures are observed at several localities. There are also intraformational clasts of mudstone, indicating small breaks during deposition.
*Boundaries:* The lower boundarie of the Edøya Formation is transitional towards the Grønskaget Formation. The upper boundary is transitional towards the Kyrhaug Formation.
*Distribution:* Outcrops of the Edøya Formation are mainly present on the island of Edøya. Three small islands, Nørholmen, Otterholmen and Isaksholmen, mark the northeasternmost extension of the Edøya Formation. To the southwest, the outermost outcrops occur at Bjørnholmen.

### *The Kyrhaug Formation*

*Name:* Kyrhaug is situated in the eastern part of Edøya. New Name.
*Type section:* The type section of the Kyrhaug Formation is along the eastern coast of a small bay at Kyrhaug Fyr (MR-6070 2041).
*Thickness:* The observed thickness of the Kyrhaug Formation is 560 m. This is probably a minimum thickness, as the upper boundary of the unit is nowhere exposed.
*Lithology:* The Kyrhaug Formation is composed of polymict, poorly sorted conglomerates with sporadic sandstone lenses and thin siltstone horizons. The conglomerates are similar to the conglomerates of the Edøya Formation; however, the reddish granitic clasts only occur in minor amounts in the Kyrhaug Formation. The clast population is dominated by green and grey sandstones, which may constitute up to 70% of the conglomerates. Other clast types include chert, quartzite, and volcanic, metamorphic and granitic rocks. The maximum particle size of the conglomerates varies from 10 to 35 cm, with an average of 18 cm (St.dev = 6.2). The clasts are usually subrounded to rounded. Normal grading and clast imbrication are common features. The conglomerates are usually clast-supported, with less than 10% matrix, composed of medium- to coarse-grained brownish-grey sandstone.

The thickness of individual conglomerate beds and sandstone lenses ranges from 20 cm to a few metres. The sandstone are fine- to coarse-grained, and vary in colour from brown to grey. Horizontal bedding, planar cross-bedding, trough cross-bedding and flame structures are occasionally observed within these.
*Boundaries:* The lower boundary towards the Edøya Formation is transitional. The upper boundary is nowhere exposed; however, on a small island at Kyrhaug Fyr there are abundant pockets of conglomerate with reddish granitic clasts similar to those of the Edøya Formation. This may indicate an interfingering between the Edøya Formation and the Kyrhaug Formation.
*Distribution:* The Kyrhaug Formation is exposed only in the southeastern part of Edøya at Kyrhaug Fyr and adjacent small skerries, including the Kyrhaug Fyr.

# Appendix 2 (by T.H. Torsvik)

Preliminary paleomagnetic data from 5 sites within the Hitra Group, carefully selected to optimize a fold-test, are reported in Table 1. Owing to low natural remanent intensity (mean = 0.3 mA/M), only 16 of 43 samples (measured on a two-axis cryogenic magnetometer, University of Oxford) gave sensible results during progressive thermal demagnetization. These, however, provide a well-defined dual polarity magnetization aligned NNE-SSW. Stepwise tectonic unfolding resulted in a negative fold-test at high statistical confidence: thus the Hitra Devonian is clearly post-tectonically remagnetized, probably at temperatures exceeding 350-400°C. The *in-situ* virtual geomagnetic pole position indicates a Devonian age, probably compatible with Late Devonian (Solundian) remagnetization obtained from Devonian rocks in western Norway (Torsvik et al. 1988).

Table 1. Palaeomagnetic data from the Hitra Devonian.

| In-situ | | | | Tectonic Correction | | | | Polarity | |
|---|---|---|---|---|---|---|---|---|---|
| DEC | INC | a95 | k1 | DEC | INC | a95 | k2 | N | R |
| 202 | +34 | 8.8 | 15.9 | 218 | +46 | 20.7 | 2.8 | 9 | 7 |
| VPG: 6.1°S, 347.9°E, dp/dm: 6/10 | | | | | | | | | |

DEC = mean declination; INC = mean inclination; a95 = 95 percent confidence circle; k1/k2 = precision parameter; Polarity (number of samples): N, Normal, R, Reverse; VPG = virutal geomagnetic pole. Fold-test negative at 99 percent confidence level.

# Reference

Torsvik, T.H., Sturt, B.A., Ramsay, D.M., Bering, D. & Fluge, p. 1988: Palaeomagnetism, magnetic fabrics and the structural style of the Hornelen Old Red Sandstone, Western Norway. Geol. Soc. 145, 413—430.