Tectonophysics, 207 (1992) 331–343
Elsevier Science Publishers B.V., Amsterdam

The interaction between oblique and layer-parallel shear in high-strain zones: observations and experiments

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(Received June 5, 1991; revised version accepted November 19, 1991)

ABSTRACT

Fossen, H. and Rykkelid, E., 1992. The interaction between oblique and layer-parallel shear in high-strain zones: observations and experiments. *Tectonophysics*, 207: 331–343.

The Scandinavian Caledonides consist of a rigid basement, a weak décollement zone, and a series of crystalline nappes. The basement is cut by oblique, west-dipping extensional shear zones, associated with west-verging folds in the overlying décollement zone. The oblique shear zones may, theoretically, have formed after, prior to, or during post-contractional, top-to-the-west shear movements. Physical modelling of this geometrical/mechanical situation shows that each of the three deformation histories results in a characteristic structure. If the oblique, extensional shear zones postdate the layer-parallel (décollement) shear, a normal-type, oblique shear zone structure occurs. If the layer-parallel shear occurs after the development of oblique shear zones, the resulting structure is characterized by a low-strain region above the oblique shear zone. Some folds typically develop above this "protected" region. However, simultaneous oblique and layer-parallel shear provides abundant asymmetric folds with gently dipping axial planes and sub-horizontal, shear zone-parallel fold axes in the region above the oblique shear zone. A neutral surface may be encountered above, and parallel to the oblique shear zone, separating a lower, thin zone of layer-parallel extension from the main, contractional region. The differences between the three types can be used to interpret field observations of this type of structures. The large-scale, post-contractional structures in the Scandinavian Caledonides are consistent with layer-parallel shear related to back-movement of nappes, synchronous with oblique shear zones in the basement. Similar geometries are found as meso- and micro-scale structures in ductile high-strain zones, where oblique shear bands or shear zones affect more competent layers within the tectonites. Also, these structures compare well with those produced experimentally by simultaneous layer-parallel and oblique shear, and are consistent with the general assumption that oblique shear bands form contemporaneous with the general shear in shear zones.

Introduction

The understanding of the formation of folds and boudins in weakly to moderately deformed rocks has increased considerably during the last few decades from observations and theoretical and experimental modelling (e.g., Biot, 1961; Ramberg, 1961, 1963; Hudleston, 1973; Treagus, 1973; Smith, 1975; Ramsay and Huber, 1987). Folds and boudins are also very common in mylonite zones, where strain is high and the interacting processes are more complicated. Although structures formed in high-strain zones may be composite and geometrically complex, they may in many cases be explained adequately by relatively simple models (e.g., Ghosh and Ramberg, 1976; Hudleston, 1977, 1989; Cobbold and Quinquis, 1980; Vollmer, 1988; Bjørnerud, 1989). In this paper, we investigate the special case where a layered, anisotropic sequence of weak rocks overlies a more rigid and mechanically stronger layer (Fig. 1). The case involves evidence for two sets of shear systems: a general shear (sub)parallel to a well developed layering in the weak rocks, and another set of subordinate, normal-sense shear zones (oblique shear) developed in an un-

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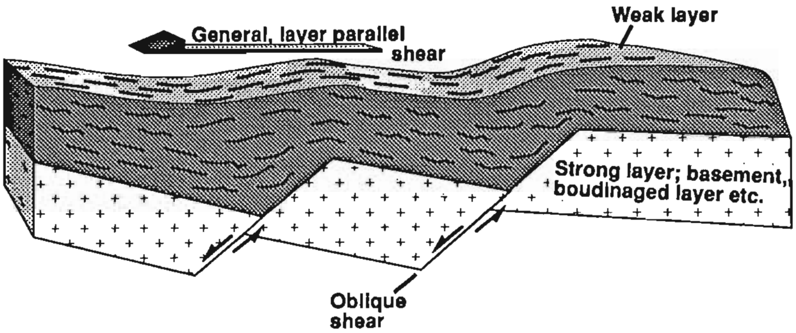


Fig. 1. Principal outline of the geometric situation considered in this article.

derlying, more rigid material. The oblique shear zones under consideration dip in the shear direction.

This situation commonly occurs on a variety of scales in sheared portions of the crust. A common small-scale example is where a rigid layer in a high-strain zone deforms by boudinage to form asymmetric boudins (e.g., Hanmer, 1986). Each boudin is separated by a small extensional shear zone, oblique to the general layering in the shear zone. A large-scale example is where the layer-parallel shear is along a weak horizon (décollement), and normal-sense, oblique shear zones affect the more rigid rocks beneath the décollement. In both situations folds and boudins may evolve in the vicinity of the oblique shear zone, depending primarily on the relative timing of the layer-parallel and oblique shear systems. In fact, we will show that oblique shear followed by layer-parallel shear, layer-parallel shear followed by oblique shear, and simultaneous oblique and layer-parallel shear result in three similar, but distinct types of structures (Fig. 2). Field observations from the Scandinavian Caledonides and simple clay model experiments are used to investigate these relationships.

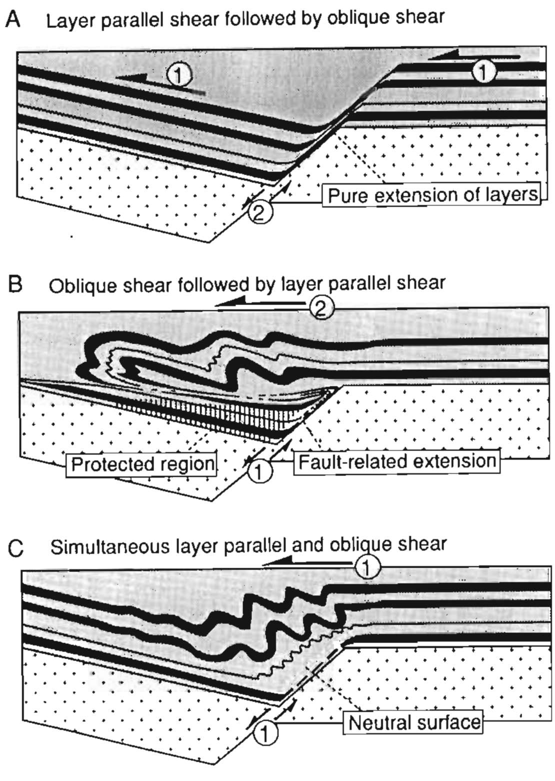


Fig. 2. The three types of structures formed by various interactions of oblique and layer-parallel shear.

Large-scale field examples

The Scandinavian Caledonides

Well developed structures of the kind illustrated in Figure 1 are well developed in the Scandinavian Caledonides as a result of post-collisional orogenic extension. During the previous collisional stage, slices of crustal rocks were thrust up to hundreds of kilometers above the mostly Lower Paleozoic continental cover sediments (cf. Hossack and Cooper, 1986). The relatively weak, phyllite-dominated cover acted as a décollement during thrusting. A strong planar anisotropy was established in the sheared sedimentary cover subparallel to the basement–nappe interface. Following the collisional stage, the Caledonian nappe wedge, built up by southeast translations of detached continental and oceanic rocks, experienced a reversal of translation direction. Kinematic analyses within the Scandinavian, as well as in the British Caledonides, show generally Ordovician–Silurian thrust-related structures (shear bands, asymmetric folds, S–C relationships, microfabrics) overprinted by Lower to Middle Devonian back-movement or extensional structures (Fossen and Rykkelid, 1988; Milnes et al., 1988; Rykkelid, 1988; Chauvert, 1989; Séranne et al., 1989; Powell and Glendinning, 1990; Fossen, 1991, and in prep.; Rykkelid in prep.). In addition to back-movement of Caledonian nappes, the underlying basement was stretched by the development of discrete extensional shear zones (Fossen, 1991). The resulting geometry is directly comparable to Figure 1.

The Rombak window

The basement–cover–nappe sequence is exposed in glacial valley walls in the Rombak window, North Norway (Fig. 3). The granitic basement and its sedimentary cover were both affected by early west-dipping reverse faults and thrusts, related to Caledonian nappe translations (Fig. 4). A set of west-dipping normal-sense shear zones, showing displacement on the order of tens to hundreds of meters is restricted to the basement. The average spacing of the west-dipping shear zones is about 500 m, and together with asymmetric, west-verging structures in the cover, they are thought to be related to the subsequent extensional event (Rykkelid, 1988 and in prep.). Folds are developed in the cover above the hanging wall in the vicinity of these shear zones (Fig. 5). The folds are mostly sub-cylindrical, concentric buckle folds or modified buckle folds, and slightly asymmetric with axial surfaces dipping to the east. The fold axes parallel the trend of the associated basement shear zone (Fig. 4). The intensity of folding decreases up-section until the folds die out 10–100 m above the basement. Characteristically, these folds are restricted to a triangular area west of the oblique shear zone.

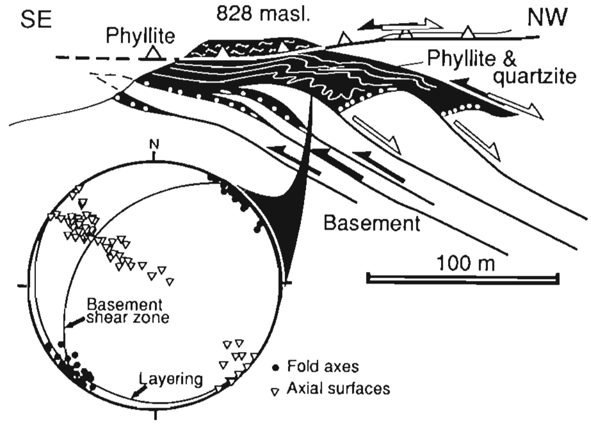


Fig. 4. Profile through the imbricated basement–cover interface in the Rombak window (Fig. 3), as observed along a NW-trending valley side along Mount Storriten, south of Skjomen, North Norway. Note reactivation of imbricate sheets and allochthonous phyllite during the post-thrusting extensional phase, with folding of the cover (phyllite and quartzite). Filled arrows: thrust-related shear. Open arrows: extension-related shear. Lower-hemisphere equal-area projection of late fold axes and axial surfaces from folded layers in the cover above the basement shear zones is shown. Note that the intersection between the shear zone and the general orientation of the layering approximates the trend of the basement shear zones.

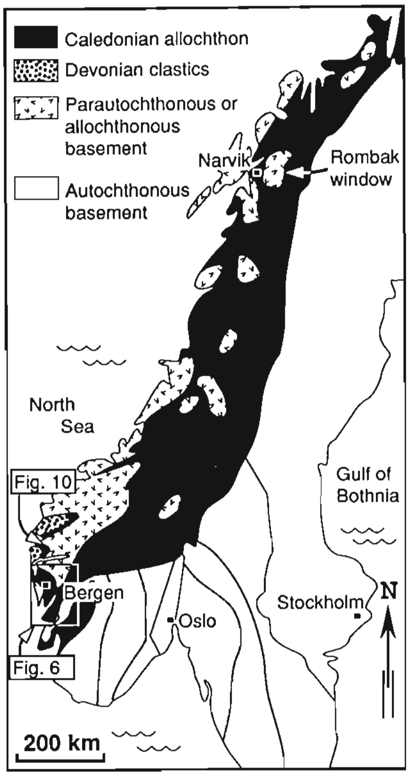


Fig. 3. Simplified geologic map of the Scandinavian Caledonides showing the locations of areas referred to in the text.

The asymmetry of the folds, the relationship between folds and basement shear zones, and the nature of the basement shear zones are all consistent with post-collisional crustal extension, indicating a simultaneous development of folds and basement shear zones by back-movement of nappes and basement extension.

Southwest Norway Caledonides

The general evolution of the Caledonides of southwest Norway (Figs. 2 and 6) is similar to that of the Rombak area, with a rigid basement separated from overlying Caledonian nappes by a weak décollement zone of essentially phyllites and mica schists. Caledonian southeast-verging structures (asymmetric folds, S–C structures, shear bands, rotated porphyroblasts, quartz microfabrics) in the décollement zone were inhomogeneously overprinted by top-to-the-northwest deformation. At some stage during this back-movement of the nappes, large-scale, narrow shear zones developed in the basement, with km-scale displacements. One such extensional shear zone runs along Hardangerfjorden, and another is located to the eastern margin of the Bergen Arcs (Fig. 6). There are folds (F2.2) that only occur in the vicinity of these shear zones, and refold somewhat earlier back-movement-related folds (F2.1). This is most clearly seen along the northeast margin of the Bergen Arcs, where the local trend of the basement shear zone is different from the trend of the F2.1 folds. In this region the F2.1 and F2.2 fold axes are nearly orthogonal (Fig. 7), and their age relationship is clear. The trend of the F2.2 folds is parallel with the local trend of the shear zone. The folds are commonly asymmetric with easterly dipping axial planes, and are restricted to the vicinity of the shear zone. Most folds are observed at outcrop-scale, but kilometer-scale folding of the Caledonian foliation in the eastern portion of the Lindaås Nappe (Fig. 6, profile A) is probably also related to deformation above the basement shear zone. In general, the structures are very similar to the example from the Rombak window described above. However, a complicating factor in the Southwest Scandinavian Caledonides is the back-movement deformation extending into the basement itself (Fig. 6, profile A). Hence, a well

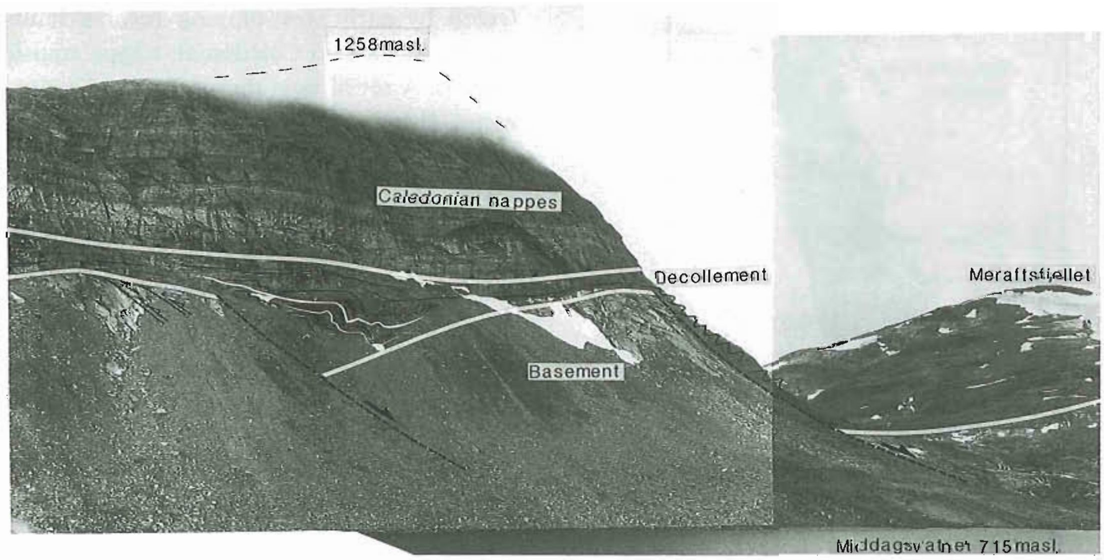
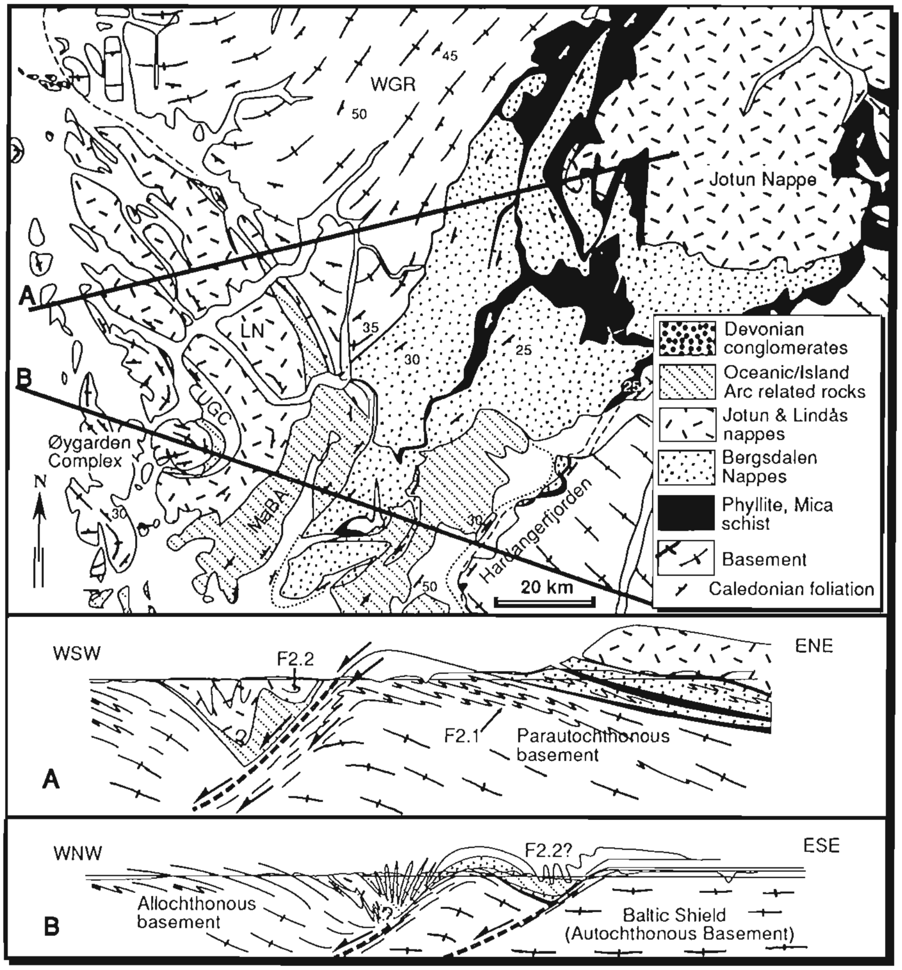


Fig. 5. Photograph of a valley slope in the Rombak window, looking south. The Precambrian basement is deformed by extensional normal shear zones. The overlying weak metasediments (décollement zone) contain abundant folds in the region above and to the west of the basement shear zones. Note that the overlying crystalline, Caledonian nappes are unaffected by this deformation. Minor reverse faults in the basement can be seen in the left part of the photograph, formed during the earlier Caledonian contractional deformation. From M711 map sheet 1331 II, UTM coordinates 950595.



U/LBN = Upper/Lower Bergsdalen Nappes, MaBA = Major Bergen Arc, LN = Lindås Nappe,
UGC = Ulriken Gneiss Complex, WGR = Western Gneiss Region

Fig. 6. Simplified geologic map of part of southwest Norway. Modified from Kvale (1960).

defined allochthon–autochthon interface is absent.

Small-scale field examples

Small-scale examples can be found in (semi-) ductile, gneissic high-strain zones, where competent layers have been asymmetrically boudinaged during shear deformation. We consider here the case where the boudin-separating shear zones are sympathetic with the general shear and truly extensional, i.e. they cut down through the local mylonitic layering. The boudinaged layers correspond to the basement, and the overlying weaker, mica-rich tectonites correspond to the sheared cover sediments in the large-scale examples above. Such rocks and structures occur in basement rocks involved in Caledonian deformation in the Bergen Arc region, southwest Norway (Øygarden Complex, Fossen and Rykkelid, 1990) (Fig. 6). In some cases, folds are associated with the extensional shear zone separating two boudins (Figs. 8 and 9a). The geometric configuration is similar to the above large-scale examples, with folds restricted to an area above the hanging wall near the extensional shear zone. The folds are typically asymmetric, overturned in the direction of transport. Local competent layers with "characteristic" wavelength and constant thickness indicate that these folds formed by buckling, although some

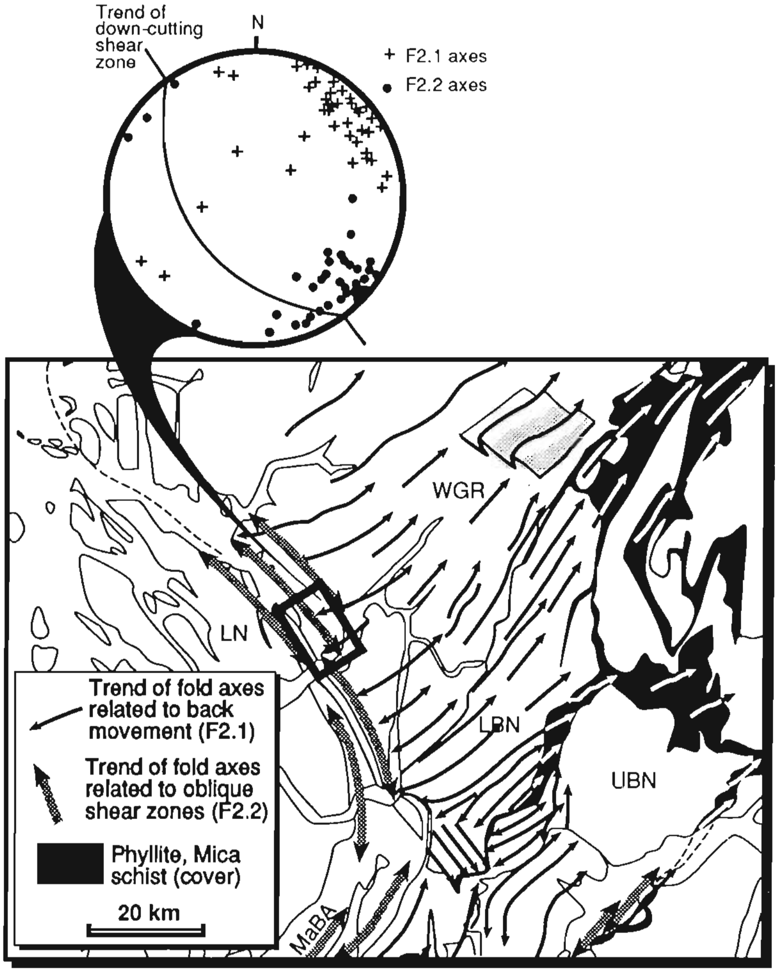


Fig. 7. Trend of post-thrusting fold axes in the area covered by Figure 6. A wide-spread family of folds (F2.1) are folded by a set of folds (F2.2) along the Major Bergen Arc (*MaBA*), i.e. near a basement-cutting shear zone. Lower-hemisphere equal-area projection projection of F2.1 and F2.2 fold axes is shown for the outlined rectangular region. Note that the F2.2 axes are approximately parallel to the trend of the basement shear zone.

additional amplification by a passive fold mechanism may be present. Although folds do not always develop near boudin-separating shear zones (Figs. 8a and 9b), many of these cases may be explained by the absence of competent layers of reasonable thickness necessary to produce buckle folds.

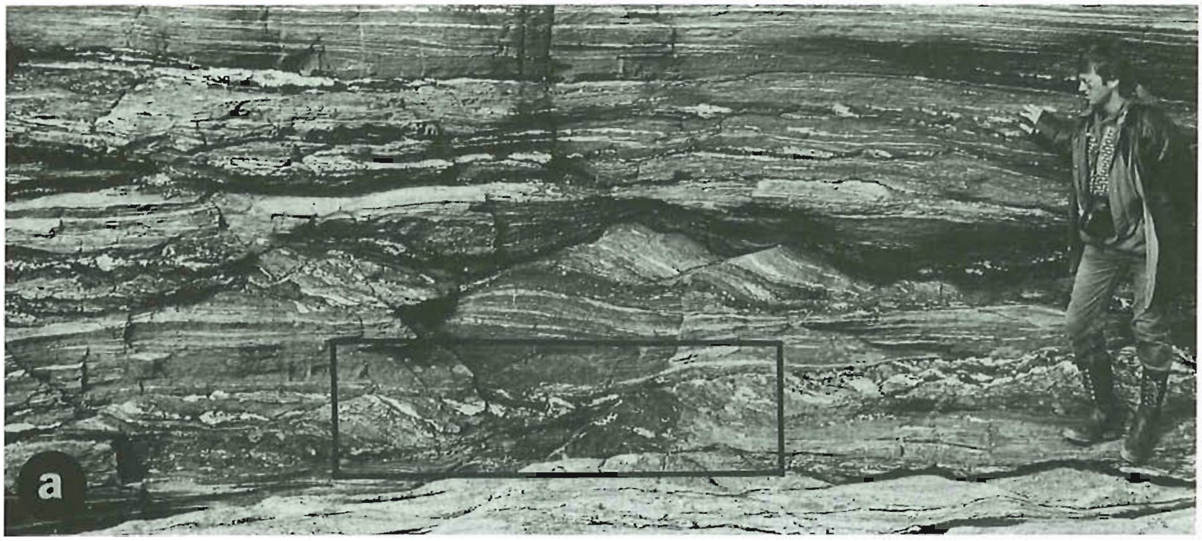
Microscale examples also occur in high-strain zones where downward-cutting shear bands correspond to the oblique shear zones in the large-scale examples. Quartz-feldspar domains or lenses, corresponding to the rigid basement in the large-scale examples above, have been cut by microscopic extensional shear bands. Microfolds may be developed above the oblique shear zone as shown in Figure 10.

Experimental study

Clay experiments were carried out in order to try to reproduce the observed structures, and to test the interpretation of the field observations.

Apparatus and material

A modified shear box was constructed as shown in Figure 11a. The box was made from wood with transparent plexiglass on one side. The vertical



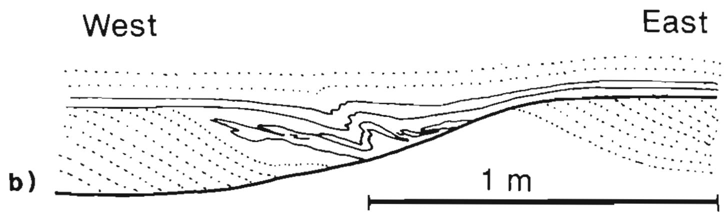


Fig. 8. (a) Asymmetric boudinage of layered tectonites from high-strain zone in the Øygarden Complex near Bergen, Norway (Fig. 6). Note that folds are not developed above all the oblique shear zones. (b) Close-up drawing of area indicated in (a). In this case folds occur above and west of the oblique shear zone.

sides of the shear box were lubricated to minimize friction. The basal and top wood plates had rough surfaces to avoid localized slip along the clay–wood interface. Rotation of the hanging wall part of the basal wood plate imitates the development of a normal shear zone in the basement. In

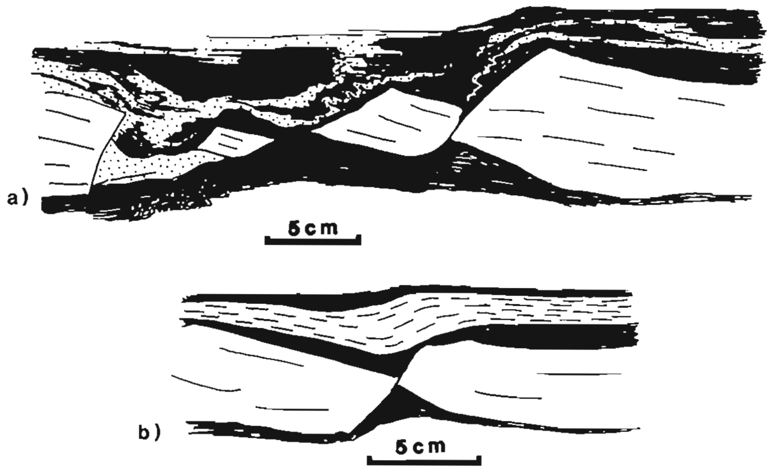


Fig. 9. (a) Boudinaged granitoid layer in biotite gneiss (deformed migmatites) of the Øygarden Complex, southwest Norway. Leucocratic veins (dashed) are folded to the left (west) of narrow shear zones separating the boudins. (b) Asymmetric boudinage ca. 1 m away from (a). The foliated, leucocratic layer above the boudinaged, granitic layer has shortened by thickening rather than folding due to the relatively large thickness of the competent layer.

the back end, an arrangement was made to induce simple shear to the material within the shear box (Fig. 11b).

Two different types of clay were used in the experiments. Dark, medium-soft clay defined relatively competent layers. This clay formed layers ca. 2 mm thick within a light grayish, water-saturated, and less competent powder clay (Fig. 12).

Two preliminary pure-shear type experiments were carried out to study the rheological behavior of the composite material. The clay "stratigraphy" was (a) shortened by a factor of 0.5, and (b) stretched by a factor of 2. For case (a), approxi-

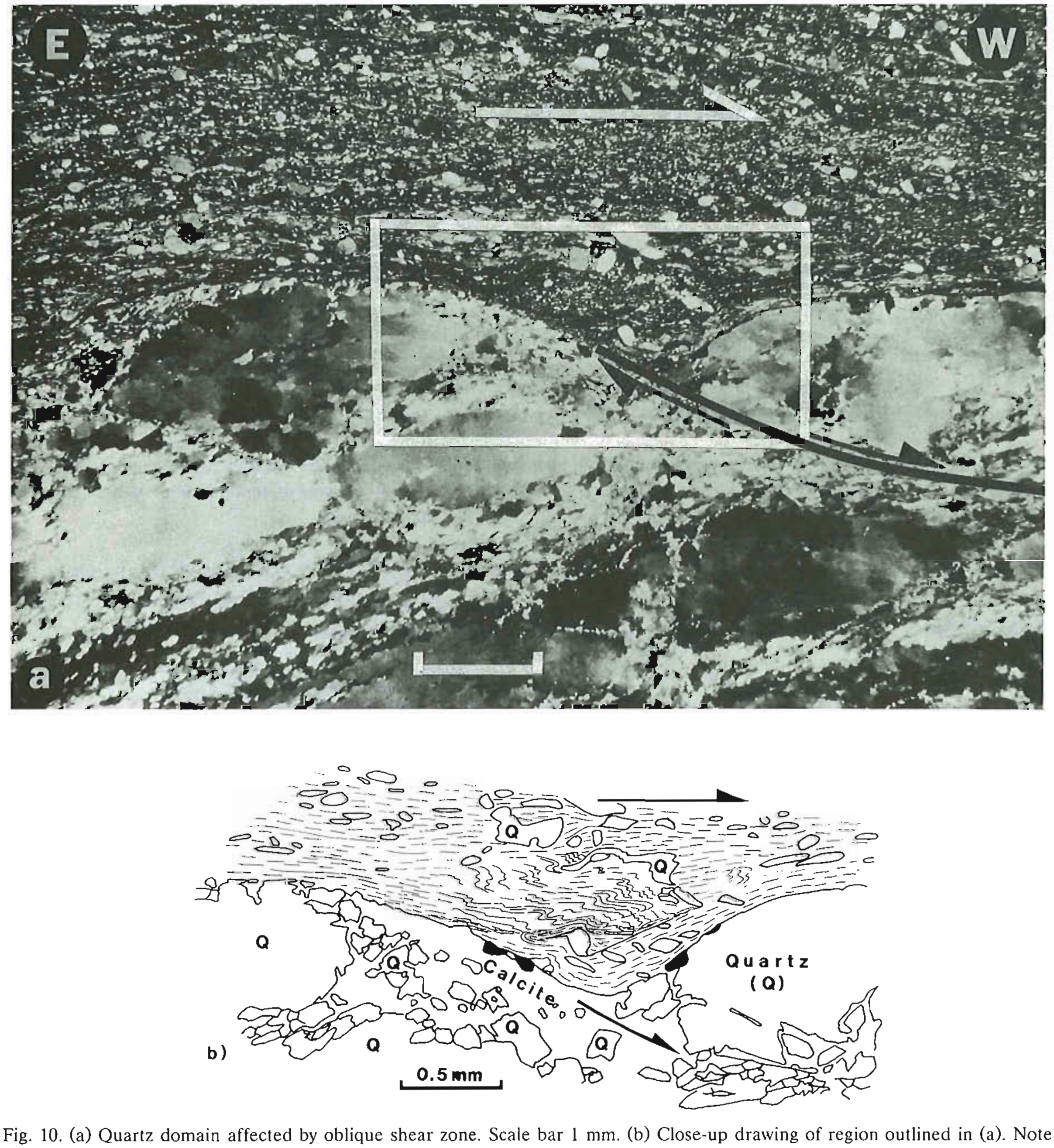


Fig. 10. (a) Quartz domain affected by oblique shear zone. Scale bar 1 mm. (b) Close-up drawing of region outlined in (a). Note folds developed in the region above the down-cutting shear zone. Calcite has precipitated in the shear zone due to pull-apart of the quartz domain. From Nordfjord-Sogn detachment west of Førde, UTM coordinate 057196, M-711 map sheet 1117 I.

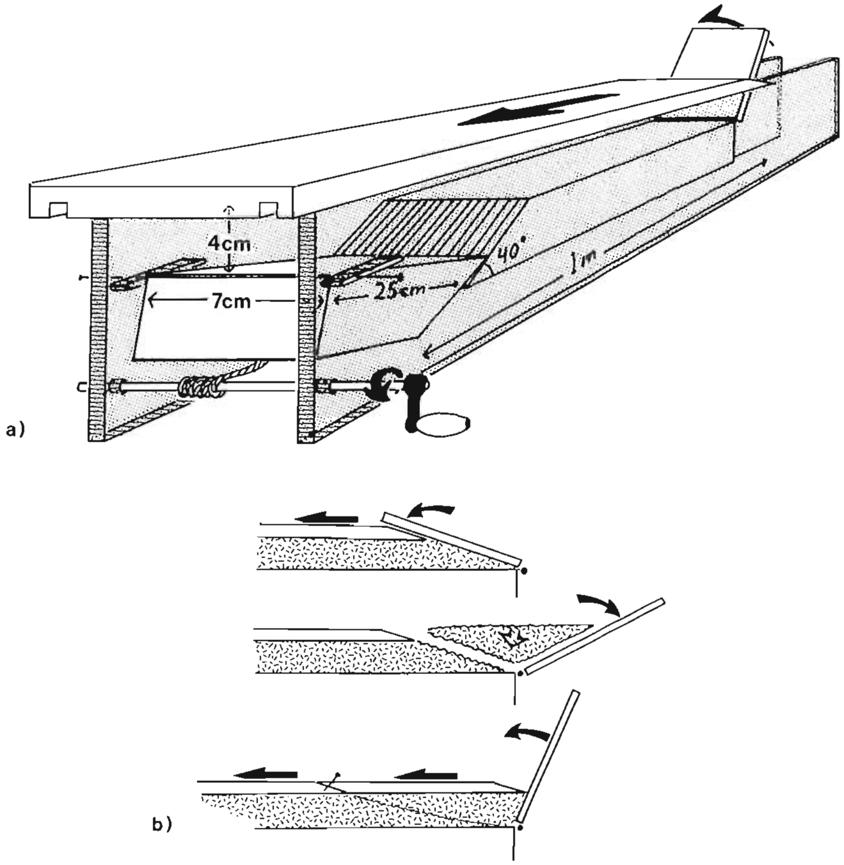


Fig. 11. (a) The modified shear box set-up for the experiments. (b) Shear strain is produced by anticlockwise rotation of a back plate which simultaneously pushes the top plate forward. When the back plate has been rotated a certain amount, it is rotated back, the space filled by clay and the top-plate is extended backwards. Renewed rotation of the back plate can now add more shear strain to the clay sequence.

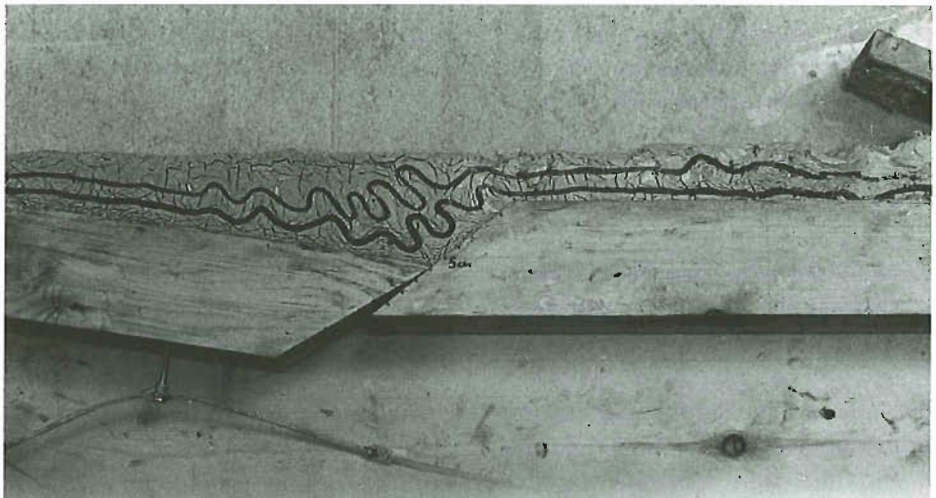


Fig. 12. Photograph of shear box after experiment. Dark layers are more competent clay in light-grey incompetent clay. The experiment corresponds to Figure 13a. See text for discussion.

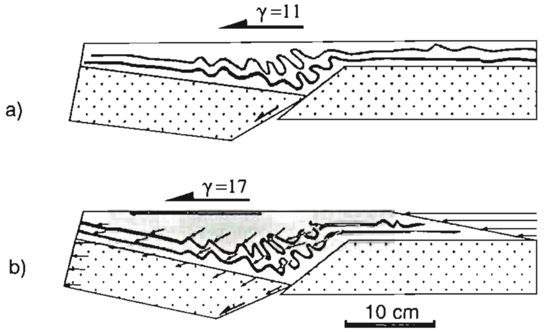


Fig. 13. Result of simultaneous layer-parallel shear and faulting for (a) $\gamma = 11$, and (b) $\gamma = 17$. Displacement vectors for the step from (a) to (b) are indicated.

mately 80% of the shortening of the competent layer occurred by buckling, and 20% by homogeneous shortening. For case (b), boudinage accounted for ca. 40% of the stretching so that 60% of the stretch was homogeneous.

The shear box provides the opportunity to shear the clay stratigraphy parallel with the layering (layer-parallel shear), as well as inducing a complicating fault movement oblique to the layering (oblique shear) in the basement plate. We want to investigate the effect of different combinations of oblique and layer-parallel shear. Three cases were studied: (1) all layer-parallel shear occurs before the oblique shear; (2) the layer-parallel shear follows the oblique shear; (3) simultaneous layer-parallel and oblique shear. The first case will obviously result in a normal-sense fault or shear zone, transecting and off-setting the previously sheared clay stratigraphy (Fig. 2A). The two other cases are less predictable and have been subject to the modelling discussed below.

Layer-parallel shear only

The response of the clay sequence to layer-parallel shear may be studied in the portion of the clay situated to the right of the basement fault (Figs. 12 and 14), i.e. clay that is unaffected by the oblique shear. The deformation in this region approximates simple shear. Some internal variations in slip rate were evidently present, resulting in local compression (folding) and extension (boudinage) of the competent clay layers. These folds and boudins are typically asymmetric. The folds are overturned in the transport direction, and the boudins are rotated. These regions of local layer-parallel contraction (folds and thickened layers) and extension (boudins and thinned layers) coexisted to satisfy the overall simple shear deformation within the shear zone. The amount of local layer-parallel shortening or extension, as indicated by folding and boudinage, is low compared to the finite shear strain. The shortening is largest in the experiment shown in Figure 15c and was estimated to be $0.5 < (1 + e) < 3$ over 10-cm intervals with a shear strain of 13.

Synchronous layer-parallel and oblique shear

In this experiment we let the normal fault be active during the layer-parallel shear. The velocity of the hanging wall of the normal fault was ca. 1/7 of that of the top plate of the shear box. For ratios higher than this, the volume of clay entering the fault zone will be smaller than the space created above the hanging wall, and a void would result, i.e. a generally unrealistic geologic situation.

The resulting structures were examined for a shear strain (γ) of 11 and a fault offset of 6 cm (Fig. 13a). The shear experiment was subsequently continued until γ reached 17 and the fault offset was 9 cm (Fig. 13b). Overturned folds developed above the hanging wall similar to the field examples described above. The folds have horizontal axes parallel to the footwall, and their axial surfaces dip 30–60° towards the footwall.

The folding (and stretching) of the competent clay layers is closely related to the local displacement rate. For steady state deformation there is an imaginary surface above the basal plate (footwall) where the displacement rate or velocity of the clay particles equals the displacement rate

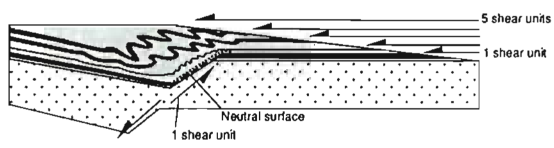


Fig. 14. Generalized figure for simultaneous layer-parallel shear and faulting, showing the position of the neutral surface and its relation to the offset along the fault.

of the normal fault (Fig. 14). Material particles beneath this imaginary surface will experience a velocity *increase* when passing the edge of the footwall. Hence, layer-parallel extension occurs in this region. Similarly, particles situated above this surface will experience a *decrease* in velocity when entering the fault zone, and layer-parallel compression occurs. This surface of constant velocity separates regions of compression and extension, and is in this sense a neutral surface, somewhat similar to neutral surfaces of buckle folds (e.g., Ramsay and Huber, 1987). The location of the neutral surface can be raised or lowered, and hence the volume of rock undergoing extension can be increased or decreased if there is a shear strain gradient across the shear zone. For example, a concentration of shear strain in the lower part of the weak layer will lower the neutral surface, and thus cause a relatively smaller volume of clay to undergo extension. The neutral surface will also be lowered if the ratio between the oblique shear component and the layer-parallel shear is increased. For very small components of oblique slip (movement along the fault), extension only occurs in a narrow band close to the footwall. If the interface between the stiff and weak layers is a slip plane, the syn-deformational velocity within the sequence of weak material may everywhere be higher than the velocity of the oblique shear movement. This would lead to retardation, thus potentially folding the material entering the shear zone, resulting in no neutral surface.

Layer-parallel shear following oblique shear

In this case we let the normal fault or shear zone be active first, and then superimpose layer-parallel shear. During the experiment, the fault did not propagate into the clay, and the clay did not by itself flow into the open space created by fault movement. The fault was moved in steps, and between each step the open space problem was overcome by increasing the vertical load so that the layered clay sequence draped the fault escarpment (Fig. 15a). The "syncline" above the fault zone was filled by new clay to retain a planar upper surface. Later layer-parallel shear ($\gamma = 15$) resulted in folding to the left of the fault escarpment (Fig. 15b), but this was less pronounced than in the previous experiment, and at only a relatively high level in the clay sequence. The fold developed because the draped layering was in the field of compression during shearing. A pronounced feature is the sudden increase in strain across a relatively narrow zone extending to the left from the top edge of the footwall. Folds are developed above this zone, whereas layers below are protected from shear deformation or undergo some extension away from the fault. The high-strain zone takes on the character of a thrust in the experiment. This unpredicted strain localization is probably related to the initial boudinage of the competent layer, with shear strain concentrated between the boudins. A similar high-strain zone above the protected region would occur if there were a localization of slip along the wood plate–clay (basement–cover) con-

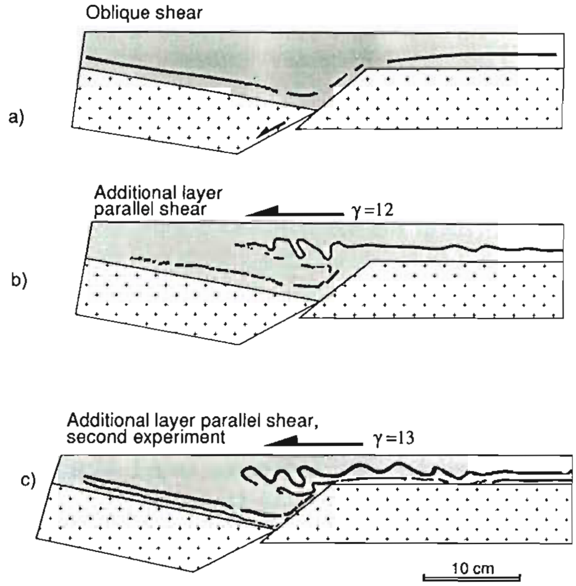


Fig. 15. Layer-parallel shear followed by faulting: (a) situation after faulting only; (b) after faulting and shearing ($\gamma = 12$); (c) final result of second experiment of oblique shear followed by layer-parallel shear ($\gamma = 13$). The formation of folds and boudinage in the right-hand region of the clay indicate local deviations from the overall simple shear deformation.

tact. A second experiment gave a similar result (Fig. 15c).

Discussion and conclusions

The three cases of oblique shear and layer-parallel shear discussed in this paper are summarized in Figure 2, resulting in three fundamentally different types of structures. If oblique shear occurs after the layer-parallel shear, the basement shear zone will simply migrate into, or thin and flex, the overlying rocks (Fig. 2A). On the other hand, if oblique shear occurs prior to the layer-parallel shear, the rocks (clay) within the half graben structure tend to be protected from most of the shear deformation (Fig. 2B). In general, both large- and small-scale structures observed in the field are very similar to those produced experimentally by synchronous layer-parallel shear and movement along the basement fault (Fig. 2C).

Many meso- and micro-scale mylonitic structures (asymmetric boudinage, shear bands) show the characteristics of simultaneous layer-parallel and oblique shear. However, there are many cases where no folds have formed in the vicinity of the oblique shear. The absence of folds could simply be due to absence of layers thin enough and/or with sufficient viscosity contrast for folding to be initiated. In some cases the layers may have deformed by passive thickening rather than by folding. In other cases the displacement rate of the oblique shear zone may have been high relative to the layer-parallel shear strain rate. Another possible reason is that the dip of the extensional shear zone with respect to the layering, is very low. In the experiments, an angle of 40° was chosen to model the large-scale observations from the Rombak window. If the angle is low, the layer-parallel compression will be weaker, and the folding is less likely.

The geometry and distribution of folds predicted by synchronous layer-parallel and oblique shear also match the large-scale features observed in the basement–cover–nappe level in the Scandinavian Caledonides (Fig. 2C) very well. The basement shear zones are therefore interpreted as having formed during the back-movement of the overlying nappes rather than being older, rift-related extension faults related to the Late Precambrian continental rifting prior to the opening of the Iapetus Ocean, or younger than the back-movement. This has important tectonic implications for the interpretation of the structural development in the Caledonides.

The constant movement rate ratio between the layer-parallel and oblique shear in the physical modelling of simultaneous layer-parallel and oblique shear is probably idealized. Temporal variation of strain rate related to the layer-parallel shear and, particularly to the oblique shear zone, will result in fluctuation of the neutral surface and thus successive overprinting of extensional and compressional structures. The abundance of mechanical/lithological heterogeneities may result in large variations in strain profiles and strain partitioning during deformation, which may cause complications. This is particularly true if the base of the weak rocks is a non-continuous slip-plane. Additional complications would occur in cases where the basement (strong layer) is affected by the layer-parallel shear, and/or the oblique shear zone is a broad shear zone. In the last two cases the neutral surface may extend into the basement. This seems to be the case in the Bergen Arcs region, where the oblique shear-related folds (F2.2) can be mapped a few kilometers into the basement hanging wall. In general, however, there are close similarities between the field observations and the experimental results, suggesting that the characteristics of the three types of structures shown in Figure 2 have general application.

Acknowledgements

The authors are grateful for comments by P.J. Hudleston, C.P. Teyssier and B. Tikoff, and thank E. Swendsson for providing thin sections from the Nordfjord-Sogn detachment. Support was provided by the Norwegian Research Council for Science and the Humanities (NAVF grant no. 440.89/061 for H.F. and no. D.40.31.182 for E.R.). This work is part of the authors PhD (H.F.) and Dr. Sci. (E.R.) theses.

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