The role of extensional tectonics in the Caledonides of south Norway

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Abstract—Detailed and regional structural–kinematic analyses of the Caledonides of south Norway show evidence for a crustal extension that is far more widespread and substantial than previously thought. Two different modes of extension are recognized. The first (Mode I) involved kilometer-scale back movement of the Caledonian nappes. The back movement resulted in extensive shear deformation within the décollement zone between the nappe wedge and the basement as seen by abundant asymmetric mylonite structures indicating translation of the Caledonian nappes towards the northwest. A second mode (Mode II) involved extension of the Baltic Shield by the formation of large-scale, normal-sense oblique shear zones. Mode II extension occurred partly during and partly after the back movement of the overlying nappes.

The extensional deformation consistently overprints the contractional Caledonian deformation in the whole region, and is therefore interpreted as a distinct (Devonian) event post-dating the Ordovician–Upper Silurian construction of the Caledonian orogenic wedge. The uniform and regional-scale back movement of the orogenic wedge towards the central parts of the orogen indicates that the extension at the end of the Caledonian orogeny was closely related to post-collisional, Lower to Middle Devonian plate divergence.

INTRODUCTION

OROGENIC belts are characterized by extensive contractional deformation in the form of thrust and nappe tectonics. However, extensional deformation has recently been reported from several mountain ranges, and has been suggested to be an important mode of deformation during, as well as after, the overall contractional history of an orogen (e.g. Platt 1986, Dewey 1987, Behrmann 1988, Ratschbacher *et al.* 1989, Gibson 1991). For example, the Himalayan orogen shows evidence for internal extensional deformation in an overall contractional setting (England 1983), and the Basin and Range region in the western United States shows indication of being an example of a contractional orogen which has experienced large amounts of extension (Coney & Harms 1984, Dewey 1987).

The present study focuses on extensional deformation in the Lower Paleozoic Caledonide orogen in southwest Scandinavia, where zones of strong Caledonian ductile deformation have traditionally been related to thrusting. Extensional deformation has recently been related to Devonian molasse-type basins (Norton 1986, Séranne *et al.* 1989), and some reversed movement sense of the Jotun Nappe has been indicated (Milnes & Koestler 1985). The main purpose of the present paper is to present and discuss the characteristics, distribution, evolution and significance of the extensional deformation in the southern part of the Scandinavian Caledonides.

GEOLOGICAL FRAMEWORK

The Caledonian framework of south Norway (Fig. 1) may be broadly subdivided into three tectonic units: (1) the Baltic Shield (Precambrian basement); (2) a décolle-ment zone; and (3) an overlying orogenic wedge of mostly far-travelled nappes which built up during convergence of Laurentia and Baltica in Ordovician–Silurian times. The regional-scale setting towards the end of the Caledonian orogeny is thought to have been a W-dipping subduction zone under the Laurentian margin (Greenland) which resulted in A-type subduction of the Baltic Shield in Silurian times (Griffin *et al.* 1985) (Fig. 2).

The Baltic Shield

The basement to the Caledonide orogen in south Norway is known as the Baltic Shield (e.g. Gorbatschev 1985), and is composed of numerous Proterozoic plutonic rocks, migmatites and gneisses, and locally also mid-Proterozoic supracrustal rocks (Dons 1960). This basement was overlain by Cambro-Silurian and locally Late Precambrian sediments above which the Caledonian nappes were emplaced. The main portion of the basement is autochthonous, and the Caledonian basement deformation is mostly restricted to the contact zone with the overlying décollement rocks. However, the western part of the Baltic Shield, the Western Gneiss Region (WGR), is internally affected by some Paleozoic deformation, with a general increase in intensity and metamorphic grade to the northwest (Bryhni & Sturt 1985, Griffin *et al.* 1985). Metasediments of assumed Late Precambrian and/or Lower Paleozoic age occur in this allochthonous basement (Bryhni & Sturt 1985), probably representing remnants of a sedimentary continental platform sequence continuous with the Lower Paleozoic foreland sediments to the southeast.

The décollement zone

A major décollement zone developed within the Late Precambrian to Lower Paleozoic sediments deposited
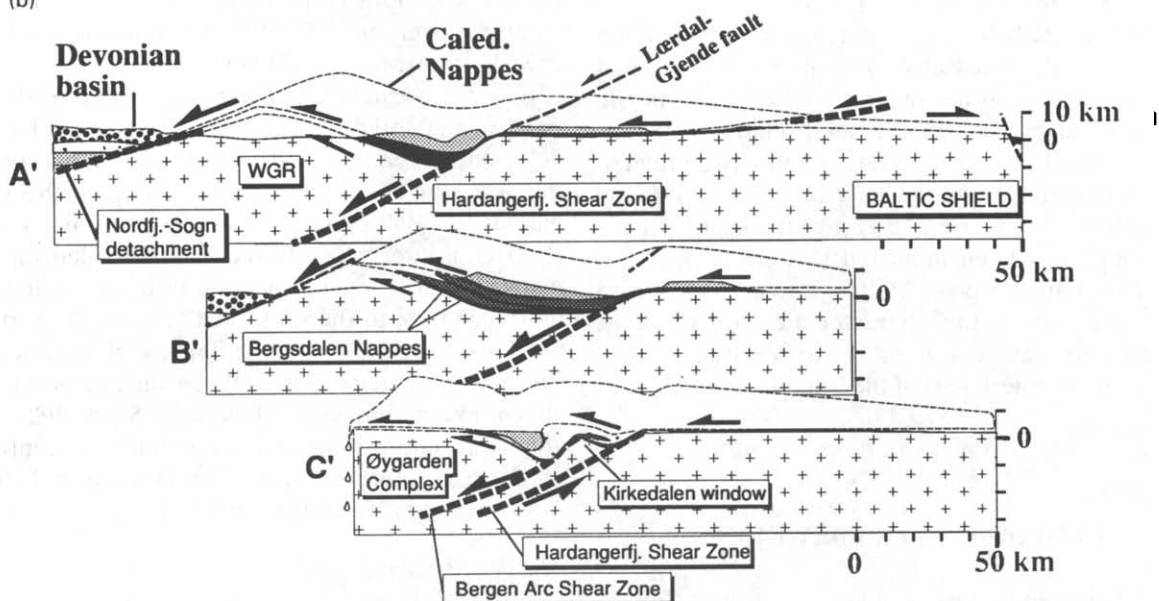
(a)
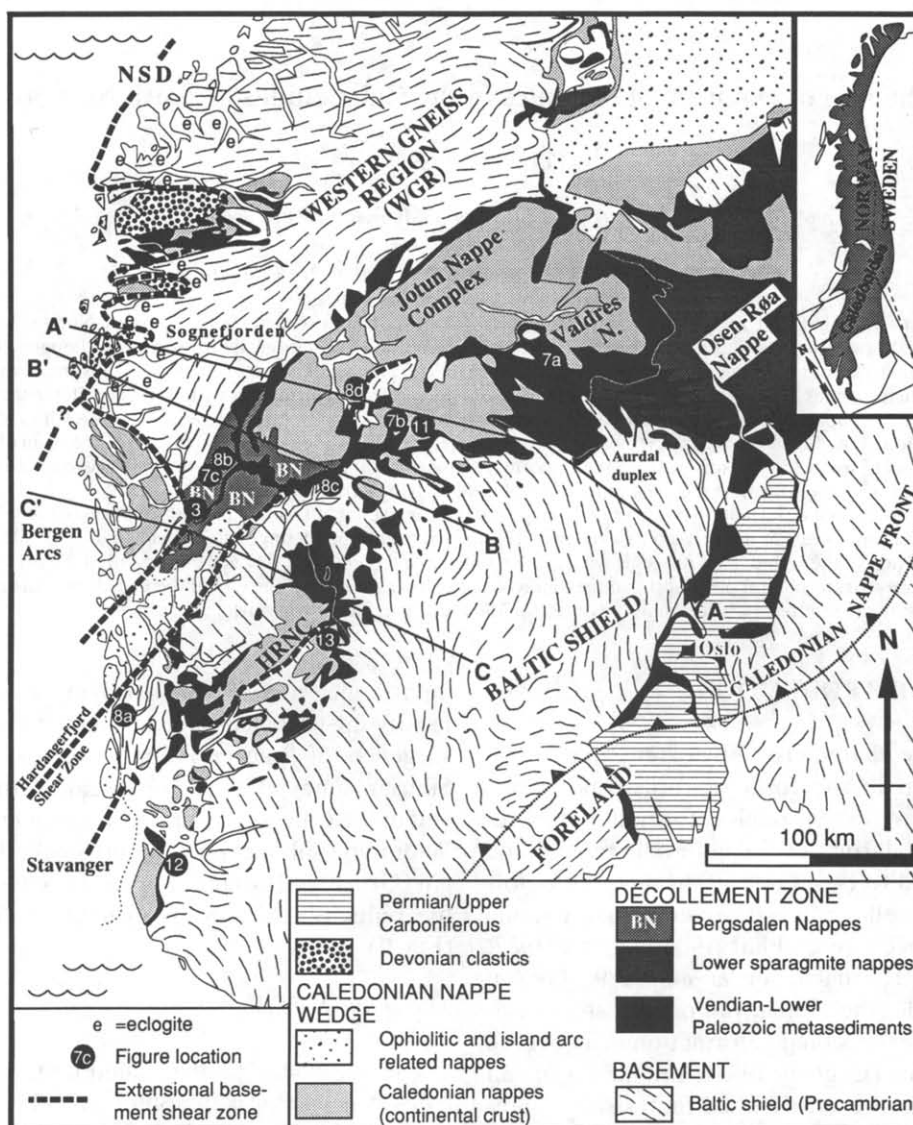
(b)

Fig. 1. (a) Simplified geologic map of southern Norway. HRNC = Hardanger-Ryfylke Nappe Complex. NSD = Nordfjord-Sogn Detachment. (b) Profiles indicated in (a).
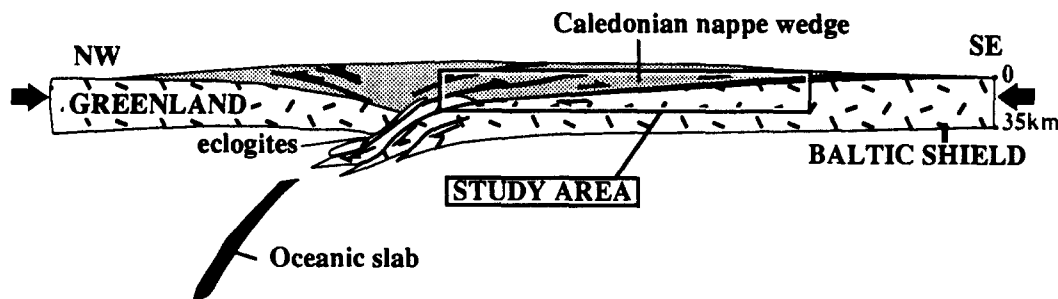
Fig. 2. Schematic section across the Baltic–Laurentian plate margin at the time of the continent–continent collision. The section studied in this paper is indicated.

on top of the Baltic Shield. These sediments were converted to a generally 100–200 m thick zone of mostly phyllites and phyllonites, which in the central and western regions have accommodated very high shear strains. The metasediments form a continuous zone of mechanically weak rocks which extends from the foreland (Oslo) in the southeast to the coastal areas in the west (Fig. 1b). In the western part of the décollement zone two large slivers of basement rocks named the Bergsdalen Nappes (Kvale 1947), appear within the phyllites. The Bergsdalen Nappes are sheets of Precambrian (Pringle et al. 1975, Gray 1978) crystalline and supracrustal rocks which probably represent detached fragments of the western margin of the Baltic Shield. They became significantly involved in Caledonian deformation and metamorphism, and are presently sandwiched between the Baltic Shield below, and the far-traveled Jotun Nappe Complex above (Fig. 1b, profile B–B').

The Caledonian nappe wedge

Remnants of a series of Caledonian nappes are exposed above the décollement zone. A general increase in the Caledonian metamorphic grade in both the décollement zone and the underlying basement indicates a SE-tapering, wedge-shaped Caledonian geometry of the nappe pile. Going from the unmetamorphosed foreland sediments to the northwest, the biotite and garnet zones are successively entered in the décollement zone under the Jotun–Hardanger-Ryfylke Nappe Complex (10–20 km), and indications of crustal thickness of 70–100 km in the northwestern portion of the WGR suggest that the maximum crustal thickness was close to or to the west of the present-day coastline.

The *Jotun Nappe Complex* (Bryhni & Sturt 1985) is the largest preserved allochthonous unit in this orogenic wedge (Fig. 1a). It represents a large portion of Proterozoic crust thrust above the Baltic Shield, and includes a variety of Proterozoic gneissic and plutonic rocks, e.g. large masses of anorthosite. In addition, some of the sheared metasediments and mylonitic gneisses in the lower part of the Jotun Nappe Complex represent the Late Precambrian 'sparagmites', with a locally preserved primary contact with the crystalline rocks (Milnes & Koestler 1985). However, the metasediments were partly detached from the crystalline parts of the Jotun

Nappe Complex, and have been described as individual nappes, e.g. the Valdres Nappe (Hossack et al. 1985), but no major slip plane appears to exist between the metasediments and the overlying crystalline parts of the Jotun Nappe Complex (Milnes & Koestler 1985).

Although the internal parts of the Jotun Nappe have escaped most of the Caledonian deformation, intense mylonitization along its base is clearly related to the Caledonian deformation in the underlying detachment zone. Considerable Caledonian deformation is also evident where Late Proterozoic sediments occur in the Jotun Nappe Complex (e.g. Hossack 1968).

The other major complex of Caledonian nappes, the *Hardanger-Ryfylke Nappe Complex*, is found to the south of the Jotun Nappe Complex. This complex consists of a variety of deformed Precambrian rocks (Naterstad et al. 1973), and shows evidence for more internal deformation than does the Jotun Nappe Complex.

The western part of the Caledonian orogenic wedge also contains more exotic rocks, including ophiolitic and island arc-related rocks of mostly Ordovician age (Dunning & Pedersen 1988).

CALEDONIAN CONVERGENCE

The construction of the composite Caledonian nappe wedge on top of the Baltic Shield is related to the convergence between Laurentia and Baltica in Silurian times. The deformation related to the Caledonian thrusting is termed D_1 in this paper, whereas the following extensional deformation is called D_2. D_1 structures are well displayed in deformed para-autochthonous, Cambro-Silurian sediments in the nappe-front region and within the Osen-Røa Nappe (Fig. 1a), where D_2 structures seem to be absent. Extensive imbricate and duplex structures in this region are consistent with translations to the southeast or south-southeast (Nystuen 1983, Morley 1986), and the imbricated sediments in the foreland region continue as a décollement zone to the northwest, overridden by successively more far-travelled nappes. This thin-skinned, contractional deformation can be traced northwestwards towards the hinterland for more than 100 km before being obscured and overprinted by the extensional D_2 deformation (see below). The westernmost preserved contractional duplexes known so far are the Aurdal-Synnfjell duplexes (Fig. 1a) (Hossack et al. 1985).
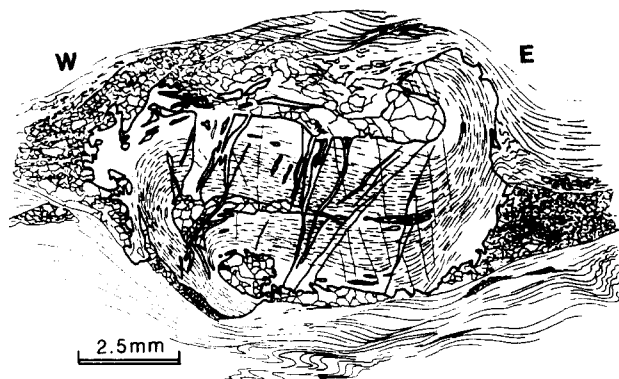
Fig. 3. Garnet porphyroclast with curved inclusion pattern indicating dextral rotation (top-to-the-east sense of shear) related to D_1 deformation and metamorphism. Old strain shadow-related quartz is preserved in the upper right and lower left parts of the garnet. Later (D_2) pressure shadows are asymmetric and indicate sinistral (top-to-the-west) shear sense, consistent with microfolds and shear bands in the matrix. Conjugate fractures developed during the D_2 back movement. Note the favored opening of the set sub-parallel to the maximum extension direction during D_2. Micaschist, Børdalsfjellet near Kvamskogen.

Further west the Cambro-Ordovician sediments turn into phyllites and phyllonites which show abundant evidence for polyphase deformation. The most striking fabrics and structures in many of these rocks formed during the extensional deformation (D_2) discussed below. Structures related to thrusting are, however, present in the form of variously modified planar and linear fabrics. Generally, the contractional structures (D_1) are better preserved in the more competent lithologies in the basal parts of the Jotun Nappe Complex and within the Bergsdalen Nappes.

D_1 fabrics have been identified and mapped using kinematic indicators. A variety of such indicators occur in the region, including S-C structures (Berthé et al. 1979), shear bands, garnets with curved inclusion trails (Fig. 3), asymmetric tails and pressure shadows of porphyroclasts, oblique grain shape fabrics, and quartz crystallographic fabrics. Intrafolial, asymmetric folds with axis oblique to the D_1 lineation indicate top-to-the-southeast (thrust-related) shear deformation. Southeast-verging close folds with sub-horizontal axes and gently W-dipping axial planar cleavage occur in the Upper Bergsdalen Nappe and a few places in the phyllites. These folds deform earlier thrust-related structures, and are interpreted as having been formed at a late stage during thrusting. In addition, a number of minor shear zones in heterogeneously deformed portions of the base of the Jotun Nappe Complex and the Bergsdalen Nappes show displacements consistent with SE-directed thrusting.

The SE-directed kinematic indicators in the décollement zone and the nappes are associated with a continuous set of stretching and mineral lineations defined by extended minerals and mineral aggregates in the mylonitic rocks. The most direct relationship between the top-to-the-southeast kinematic indicators and the lineation is found in conglomerates in the Bergsdalen Nappes (Fossen 1992) and in the Vågåmo conglomerate (Brunel 1980) where the quartz c-axis fabrics of stretched quartz-
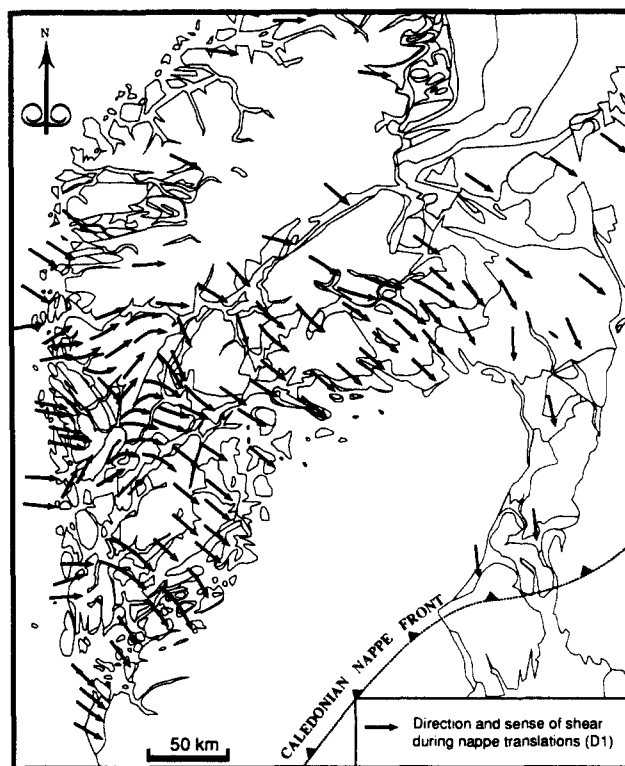

Fig. 4. Shear sense and thrusting directions during the Caledonian contraction (D_1). Data from Kvale (1953), Hossack (1968), Nystuen (1983), Morley (1986), Séguret et al. (1989), and own observations.

ite pebbles indicate top-to-the-east sense of shear, and in S-C mylonites where the D_1 lineation is an integral part of the S-C structure. In general, the reorientation of the D_1 lineation pattern by the later D_2 deformation seems to have been weak, as the orientation of D_1 lineations within and outside the regions unaffected by D_2 are very similar. It is therefore inferred that the regional lineation pattern (Fig. 4) roughly indicates the direction of nappe translation.

POST-COLLISIONAL EXTENSION

The D_1 fabrics and structures are to a large extent overprinted and locally obliterated by D_2 structures. The eastern limit of the extensional structures is approximately indicated in Fig. 5. During D_2, the sense of vorticity was basically reversed, and new structures and fabrics formed at various scales. The extensional deformation can be separated into two closely related modes of deformation. The first is characterized by structures indicating a plain reversal of the nappe translation direction (Mode I), whereas the other also involves the development of major oblique extensional shear zones (Mode II) (Fig. 6).

Mode I extension

Within the décollement zone (including the base of the orogenic wedge and the Bergsdalen Nappes) the D_1 fabrics are consistently overprinted by a new set of microscopic and mesoscopic fabrics. The new fabrics are
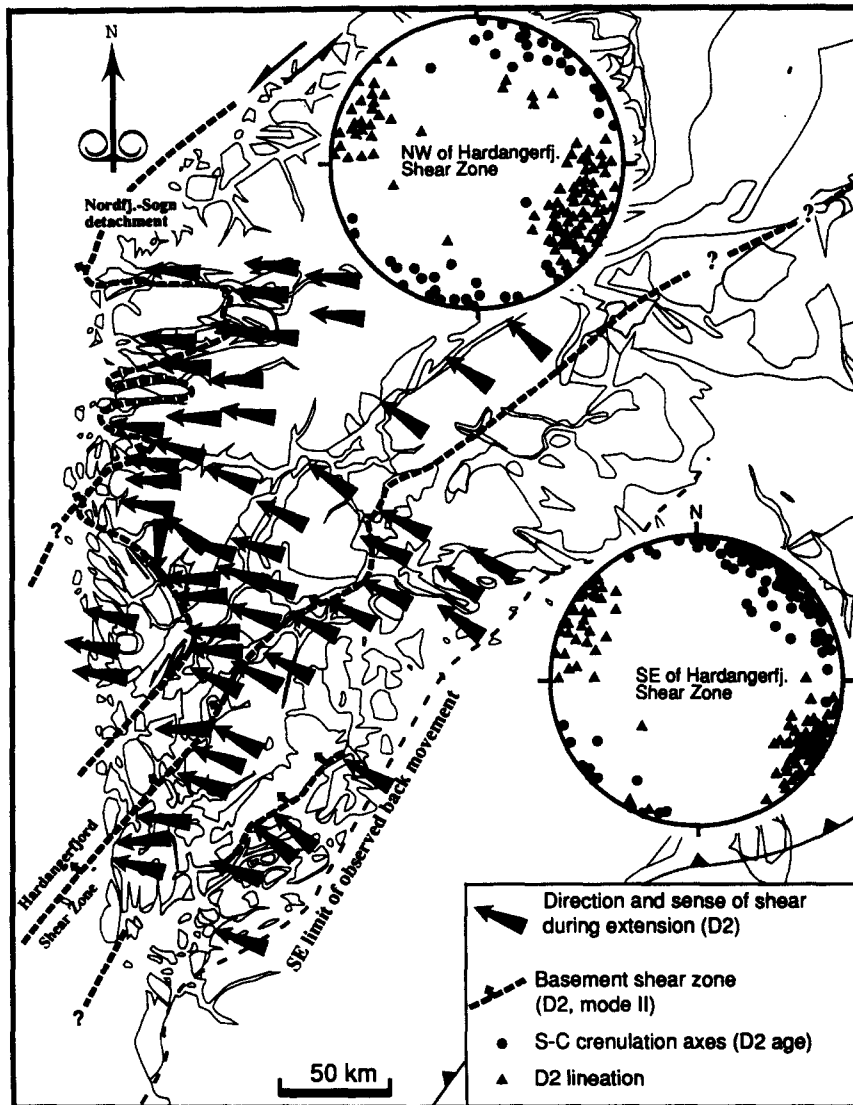
Fig. 5. Shear sense and shear directions during D_2 (mostly during Mode I). Data from Séguret *et al.* (1989) from the western Sognefjord area has been added to own observations.
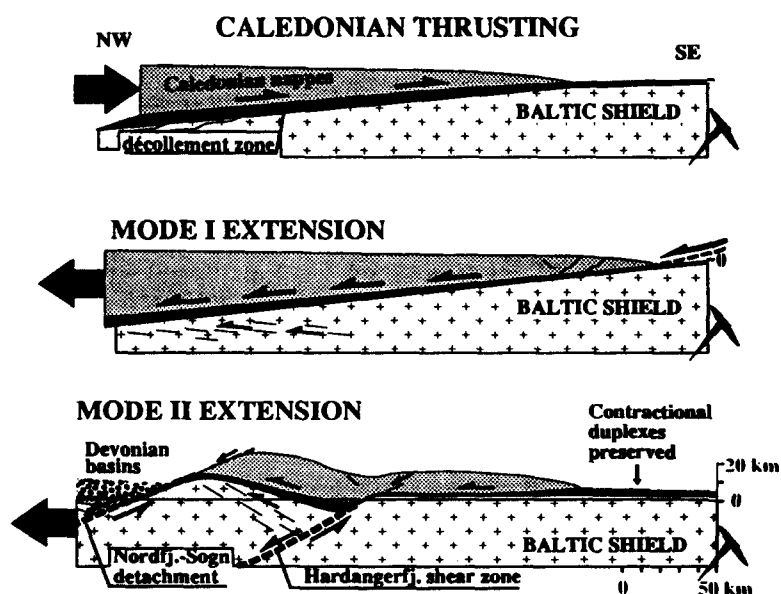

Fig. 6. Schematic development of the Caledonian wedge in south Norway.

in general very similar to the D_1 fabrics, although the metamorphic conditions seem to have been slightly lower during D_2. However, they are distinguished by their different (W-verging) geometries and overprinting nature.

The D_2 fabrics are dominated by shear bands and asymmetric folds and related axial planar fabrics. *Shear bands* are generally very well developed in the phyllites (Figs. 7a & b), in micaceous gneisses, in quartzschists and phyllonitic layers in the Bergsdalen Nappes (Figs. 7c & d) and the lower part of the Jotun Nappe Complex, and in the lower parts of the ophiolite-related rocks in the west. The foliation affected by shear bands is locally a reactivated and rotated S_1 foliation, but is more commonly an S_2 crenulation or pressure solution cleavage (Fig. 7a), or an S_2 transposition foliation. Typically, the D_2 deformation in the phyllites started out with shear-related folding and crenulation of the D_1 fabric to form a progressively stronger S_2 spaced cleavage, and at some point shear bands started to form. The resulting structure contains a D_2 fabric which anastomoses in and out of the shear bands (Fig. 7a).

The dip of the shear bands depends on the dip of the associated foliation. In the region northwest of the Hardangerfjord shear zone the foliation (S) generally dips about 30–40° to the southeast, and the shear bands (C) are close to horizontal (Fig. 9). The average S–C angle is 34° in this region (Fig. 10a), which compares well with S–C angles reported from other micaceous tectonites (e.g. Platt & Vissers 1980, Dennis & Secor 1987). The C surfaces are inclined to the E-dipping décollement zone by 10–20° (Fig. 10a). Calcite has precipitated in some of the shear bands, indicating extension across the structures. In the region southeast of the Hardangerfjord shear zone, the foliation is SE-dipping, whereas most of the shear bands dip gently to the northwest. S–C angles in this region are on average about 40°, but with a skewness towards higher angles (Fig. 10b). High S–C angles occur where the S-surface is an S_2 cleavage. In some places two generations of S–C structures of D_2 age have been observed; i.e. smaller S–C structures that form an integral part of the composite S foliation of larger S–C structures (Fig. 11). In this case, high-angle shear bands transect lower angle shear bands.

Top-to-the-(north)west S–C *structures* similar to those discussed by Berthé *et al.* (1979) occur in crystalline rocks which escaped the heterogeneous D_1 deformation but were sheared during the back movement. In most cases, however, the D_2 S–C fabric is superposed on an older D_1 S–C fabric.

Asymmetric D_2 folds, typically with northeasterly trend and sub-horizontal axes, are readily observed, and occur from millimeter scale up to kilometer scale. They are characterized by their NW vergence and SE-dipping axial surfaces, and fold D_1 planar and linear structures throughout the décollement zone (Fig. 12). The style is open to close, and a spaced S_2 cleavage is normally related to those folds. The S_2 cleavage dips to the southeast, and consistently overprints NW-dipping,

thrusting-related planar structures (Fig. 13). Similarly, r-efolding of D_1-related folds by D_2 folds is observed. Centimeter- or meter-thick zones of NW-dipping shear bands commonly alternate with zones of asymmetric folds, and the fact that the shear bands in most places are not directly affected by the folding, or vice versa, is taken as evidence for their synchronous development. Hence the folds and the shear bands may be considered as reverse and normal slip crenulations, respectively (Dennis & Secor 1987). Larger-scale folds occur in the more competent parts of the Bergsdalen Nappes, and unequivocally fold the various top-to-the-southeast fabrics formed during D_1. The extent of the D_2 folding is quite large, as folds occur everywhere in the décollement zone where D_2 deformation is present. They are previously described from the base of the Hardangervidda-Ryfylke Nappe Complex (D_4 of Andresen 1974) where they also post-date structures related to the Caledonian thrusting.

Asymmetric boudins, many of which are similar to those described by Hanmer (1986) are widespread in many gneissic lithologies. Fish-shaped boudins of granitic layers indicating top-to-the-(north)west sense of shear occur in the 'caledonized' basement close to the décollement zone southeast of the Hardangerfjord shear zone (Fig. 8a), and are very common in the reworked basement to the west. Meter or centimeter-scale *thrust-faults* occur in layered rocks, where competent, typically granitic layers are imbricated. All these structures consistently indicate (north)westward translations, and overprint the SE-verging D_1 structures.

Oblique grain-shape fabrics formed in many quartz domains or aggregates during the back movement (Fig. 8b), and asymmetric girdles of quartz c-axes have been detected. The most common kinematic microstructures are, however, shear bands and microfolds (Fig. 3, matrix).

The Mode I deformation can be traced continuously down and northwest from the Bergsdalen Nappes and the phyllites into the WGR (Fig. 1a). Here the D_2 deformation is characterized by the asymmetric, NW-verging D_2 folds, and in places by asymmetric boudins and shear bands. Similar W-verging structures are described by Milnes *et al.* (1988) from Sognefjorden.

The shear direction during the back movement has been determined at a number of localities. The stretching and mineral lineations have been used as a direct indication of the movement direction where they are clearly of D_2 age, and not just modified D_1 lineations. Crenulation axes to S–C structures, which ideally should be perpendicular to the associated lineation, have also been measured to constrain the D_2 movement direction. The general trend is top-to-the-northwest, with a more westerly trend towards the Bergen Arcs and the Devonian basins (Fig. 5). A rough estimate of about 20 km of back movement has been made by integrating the shear strain recorded by asymmetric folds in the Bergsdalen Nappes (Fossen 1992); i.e. an order of magnitude less than the D_1 displacement.
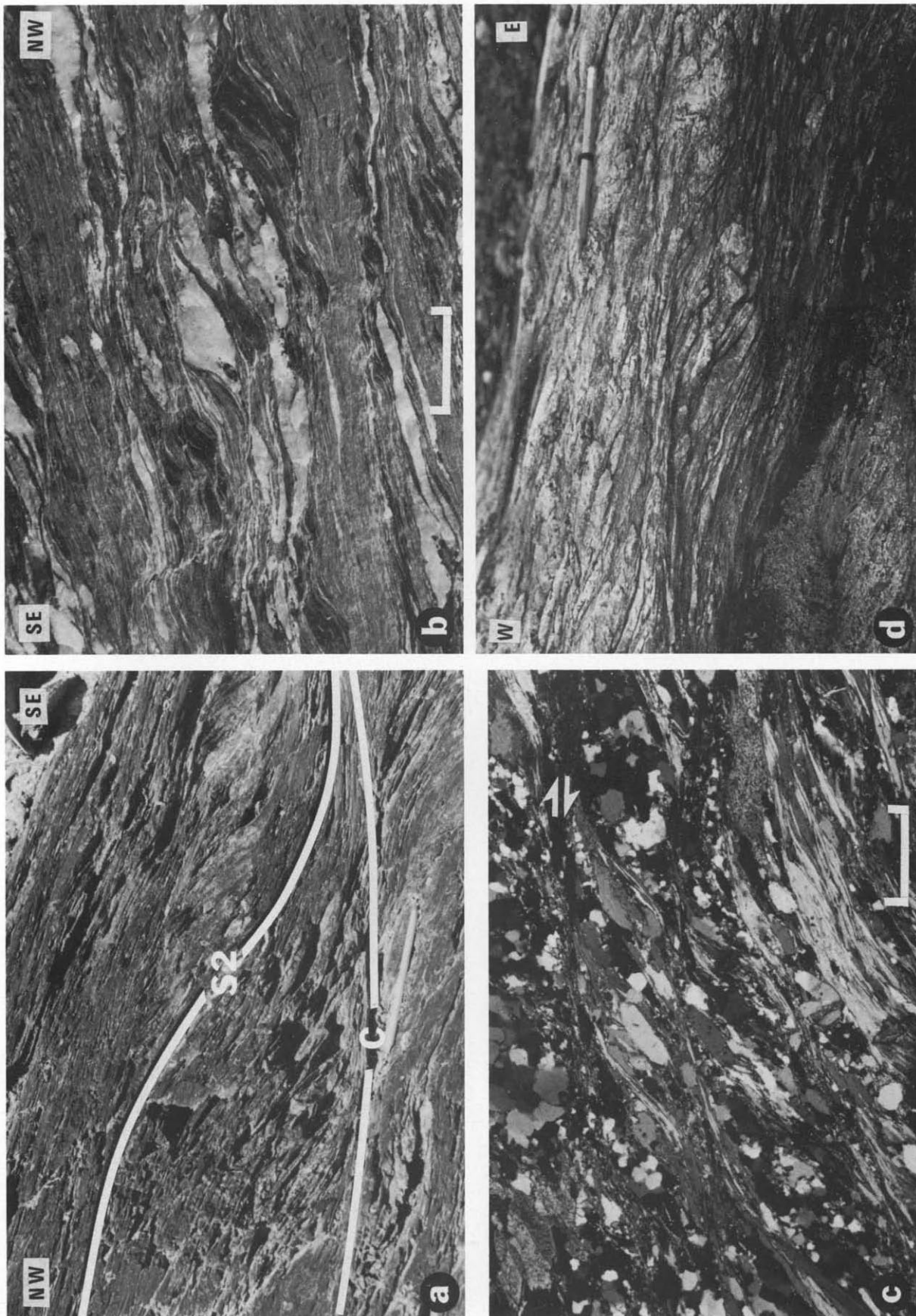
Fig. 7. D_2 S–C structures (Mode 1) from the décollement zone. (a) Phyllite under the Valdres Nappe, showing S_2 cleavage anastomosing in and out of shear bands. Note pencil for scale. (b) Shear band developed in strongly sheared phyllites under the Jotun Nappe. Aurlandsdalen. Scale bar = 2 cm. (c) Top-to-the-west S–C structures in phyllite between Upper and Lower Bergsdalen Nappes. Scale bar = 2 mm. (d) Shear bands affecting D_1 fabric in stretched conglomerate, Lower Bergsdalen Nappe.
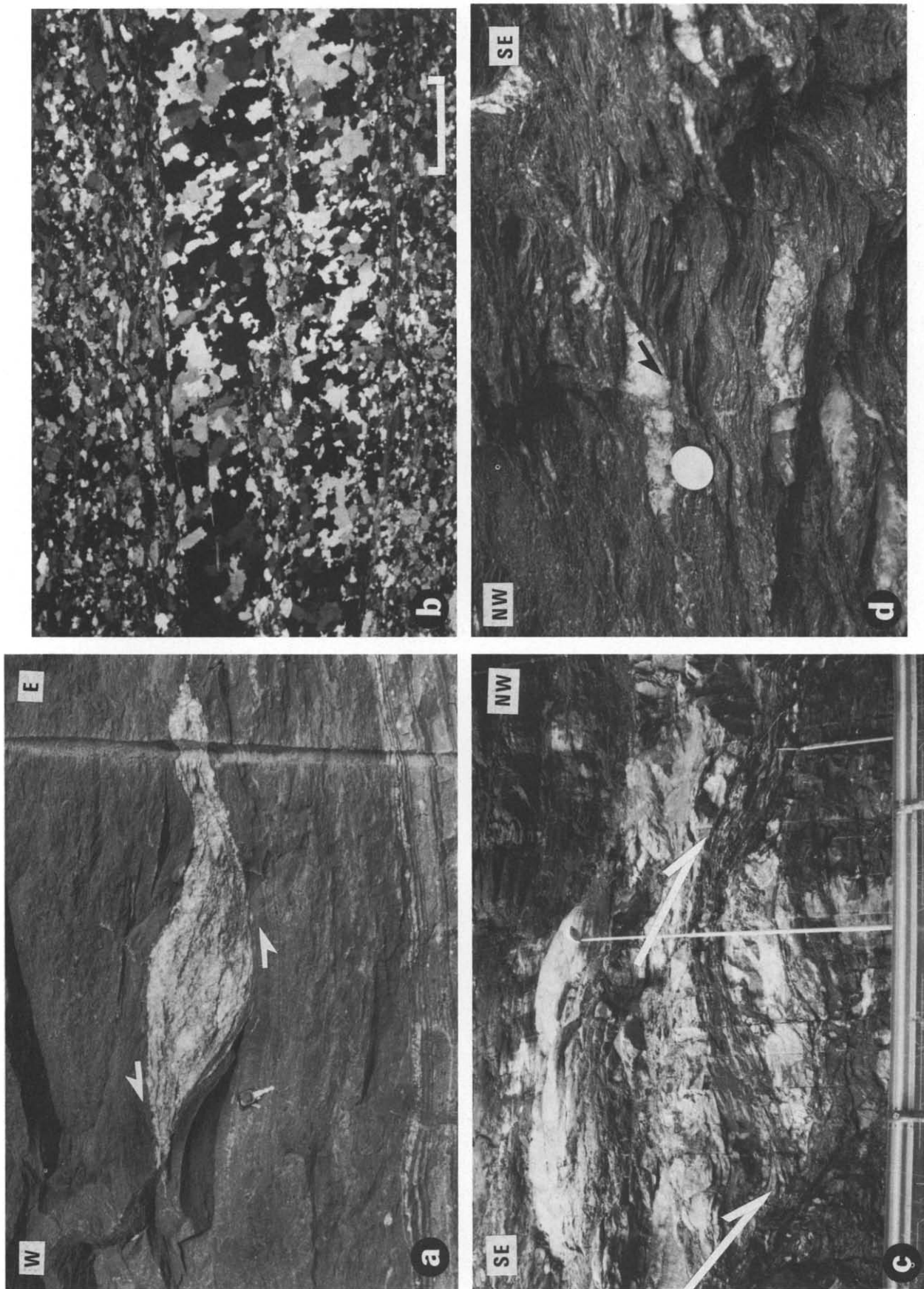
Fig. 8. (a) Asymmetric boudin (granite) in gneissic tectonites in the basement, Haugesundshalvøyen south of HFSZ. (b) Oblique shape fabric indicating top-to-the-west (left) in mylonite gneiss. Lower Bergsdalen Nappe. Scale bar = 2 mm. (c) Large-scale S–C structures in the basement close to the Hardangerfjord shear zone, Bruravik, indicating down-to-the-northwest shear. (d) Shear bands in phyllite, Hardangerfjord shear zone, near Aurland.
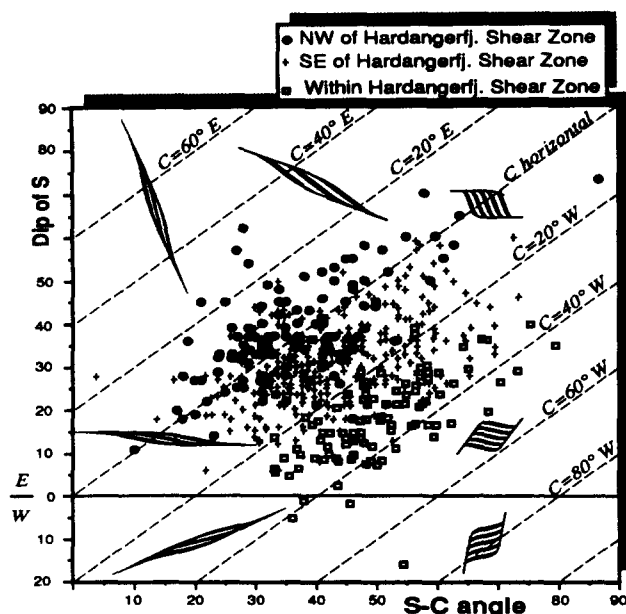
Fig. 9. Geometry of S–C structures measured in phyllite and phyllonite rocks.

Mode II extension

Two major NW- and W-dipping, normal-sense shear zones have been found in the western part of south Norway (Fig. 1). The Hardangerfjord shear zone transects the Baltic Shield, and provides preservation of portions of the Caledonian nappes to the northwest. The overlying nappes form a monoclinal flexure (the 'Fal-tungsgraben') above the basement shear zone, and the more brittle Lœrdal–Gjende Fault (Fig. 1b) may possibly be connected with this flexure. A small tectonic window to the northwest of the Hardangerfjord shear zone (Kirkedalen window) appears to expose the basement–cover contact (Ragnhildstveit 1987), which is off-set by about 2 km of vertical displacement, or about 4 km of slip along the shear zone itself. However, the displacement may be larger if the basement–cover contact in the Kirkedalen window is allochthonous.

Kinematic indicators along the Hardangerfjord shear zone include extensive S–C structures and intrafolial folds within sheared basement rocks (footwall) (Fig. 8c), well developed shear bands and related asymmetric folds in mica-bearing gneisses, phyllites (Fig. 8d) and micaschists, and oblique grain shape fabrics, asymmetric pressure shadows and S–C structures in allochthonous mylonitic gneisses above the phyllites. Such asymmetric structures may be Mode I structures rotated into the Hardangerfjord shear zone, and some of the D_2 S–C structures have been overprinted by folds or crenulations with axes parallel to the Hardangerfjord shear zone and gently SE-dipping axial planes. However, other shear bands are very planar and more likely formed as a result of movement along the shear zone.

Deep seismic reflection data across the off-shore extension of the Hardangerfjord shear zone have recently revealed the presence of a kilometer-thick, NW-dipping and NE-striking band of reflections interpreted as a
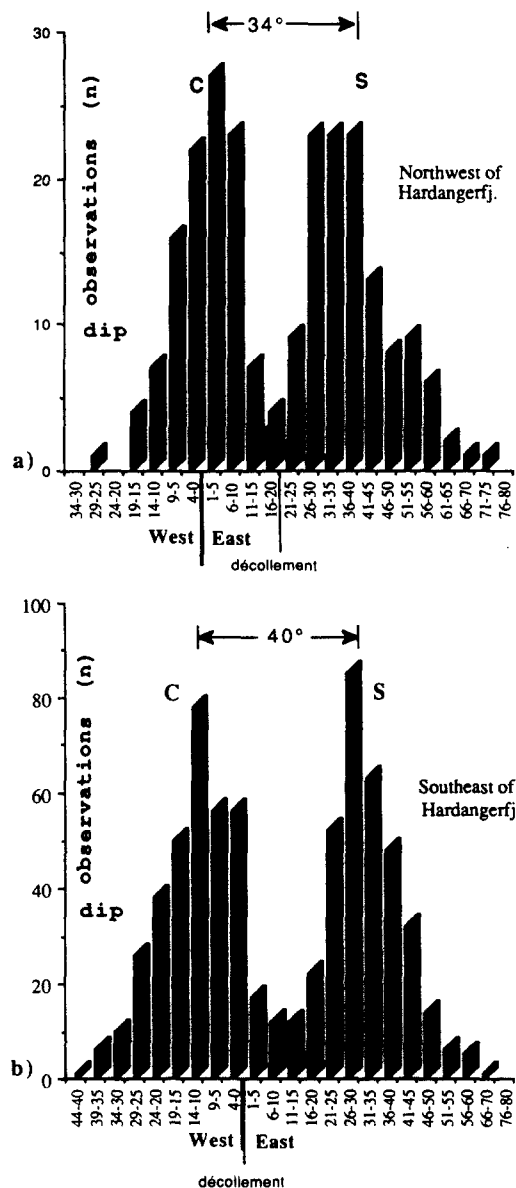

Fig. 10. Distribution of dip of S and C surfaces from field measurements in phyllites, (a) northwest of the Hardangerfjord shear zone, and (b) southeast of the Hardangerfjord shear zone.

mylonite zone penetrating to at least lower crustal depths (Hurich & Kristoffersen 1988). These data clearly support the present interpretation of the Hardangerfjord shear zone as a fundamental extensional shear zone. The existence of two smaller, less well defined shear zones near Stavanger is suggested by Hurich & Kristoffersen (1988) from similar off-shore data, possibly representing additional extensional shear zones in the basement.
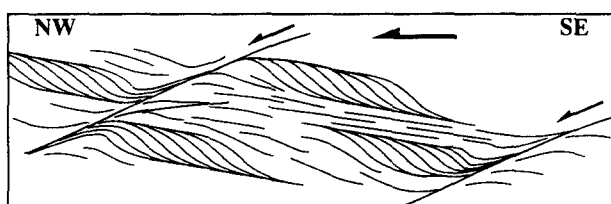

Fig. 11. Two generations of D_2 S–C structures in phyllite. Redrawn from field sketch (Aurlandsdalen near Geiteryggstunellen). Width of sketch is 15 cm.
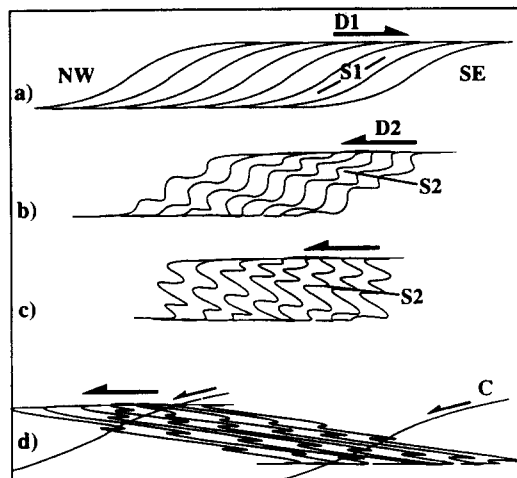
Fig. 12. Schematic illustration of the development of D_2 structures in phyllite by crenulation of D_1 S-C structure. Based on field sketches, Ullernhaug, Stavanger.

The largest shear zone, the Nordfjord–Sogn detachment (NSD) (Norton 1987, see also Hossack 1984) has already been the subject of detailed studies (Norton 1986, Séranne & Séguret 1987, Chauvet & Brunel 1988, Chauvet & Séranne 1988). This shear zone probably involves a larger displacement than the Hardangerfjord shear zone, since lowermost greenschist facies Devonian coarse-clastic sediments are in close contact with gneisses with Caledonian D_1-related eclogite facies metamorphism. West-dipping tectonites with kinematic indicators indicating normal-sense displacements are also found in the rocks along the contact between the Bergen Arcs (hangingwall) and the Bergsdalen Nappes and the southwestern part of the WGR (footwall). Numerous kinematic indicators along this zone including shear bands and related foliation fish, asymmetric grain shape fabrics in quartz layers and aggregates, asymmetric calcite shape fabrics in sheared Ordovician (Ashgill) limestones, and S-C fabrics in basement rocks in the hangingwall all indicate top-to-the-west sense of shear. Folds with gently E-dipping axial planes and shear zone-parallel, sub-horizontal fold axes are also related to the Mode II extension, overprinting top-to-the-east fabrics related to the earlier Caledonian D_1 thrusting. Local dextral and sinistral shear in the northern and southern part of the Major Bergen Arc, respect-

Fig. 13. Structures in mylonite at the base of the Hardanger-Ryfylke Nappe Complex close to the phyllite, Haukelidtunellen. Sigmoidal-shaped competent (stippled) lens in the weak (dark) mylonite indicate top-to-the-southeast sense of shear (D_1). Later asymmetric small-scale folding and a SE-dipping crenulation cleavage is developed in weak, mica-rich layers, indicating post-D_1 reversal of the shear direction. Redrawn from photograph.

ively, formed at a late stage, and possibly reflect three-dimensional Mode II accommodations to the curved shear zone geometry. Alternatively, they represent flexural slip related to Mid-Devonian folding of the Bergen Arc System. It is suggested that the Bergen Arc shear zone is connected with the NSD to the north.

Several pieces of evidence indicate that the development of Mode II shear zones continued after the back movement of the orogenic wedge ceased. One is the *en bloc* rotation of the WGR, the décollement zone, and the Jotun Nappe between the Hardangerfjord shear zone and the NSD (Fig. 1b). This rotation is reflected in the change in dip of the décollement zone and the mylonitic fabrics from sub-horizontal to 20–30° dip across the Hardangerfjord shear zone. This tilting is clearly associated with the development of the Mode II shear zones, and plausible explanations include *en bloc* rotation of the region due to a listric geometry of the Hardangerfjord shear zone and/or a footwall uplift effect related to the major NSD. The tilting prohibited further back movement of the nappes, which probably decreased and finally stopped as the rotation and slip along the Mode II shear zones continued.

Another argument is that the structures and fabrics related to the Mode I extension are clearly transected by, and rotated into the NSD (see Milnes *et al.* 1988). Evidence for this is found near the northeast margin of the Bergen Arcs, where the NW-verging D_2 folds formed during Mode I (above) are folded by a set of folds which is associated with, and restricted to, the extensional Mode II shear zone (cf. Fossen & Rykkelid in press).

The presence of Devonian clastics to the west of the NSD shows that the NSD transected and offset the *entire* nappe wedge (Milnes *et al.* 1988). This fact is hard to combine with continued slip along the décollement zone (Mode I), and is perhaps the most compelling evidence for the continued development of Mode II shear zones after the back movement of the nappes stopped.

DISCUSSION

Extensional deformation in orogenic belts may occur during as well as after the construction of the orogenic wedge, due to a variety of external or internal conditions. Extension may be the effect of local, internal adjustments during the construction of the wedge. An example is the domino-style extension of the Triassic Hauptdolomit west of the Tauern window in the Alps, caused by simultaneous overthrusting and extrusion of the footwall (Schmid & Haas 1989). However, the extension of the Caledonian wedge occurs on a regional scale, and clearly post-dates the contractional (D_1) deformation. Possible explanations for such extension may be discussed in the light of dynamic wedge theory (see Platt 1986 for a review). During contraction, a generally stable wedge can be assumed. The criteria for a stable wedge (no internal extension or contraction) being pushed up or sliding down above a weak décollement
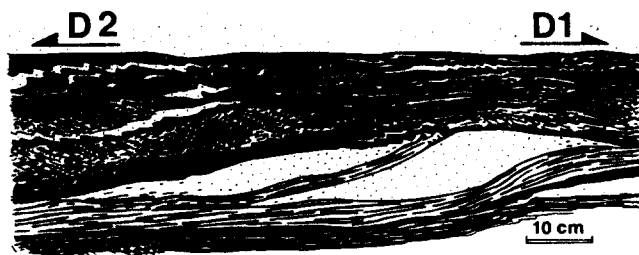
with slope β is identical to that of glaciers streaming downhill. The weight of a column of rock (ice) perpendicular to the base of the wedge has a component, $\rho g h \sin \beta$, parallel to the décollement, and this weight component is balanced at the base of the column by the shear stress τ_b (e.g. Paterson 1981):

$$\tau_b = \rho g h \sin \beta, \tag{1}$$

where ρ is density, g is gravitational acceleration, and h is the height of the column. An estimate for the dip of the décollement slope in south Norway during late stages of D_1 can be inferred from the change in metamorphic grade in the lowermost, autochthonous Cambrian cover. An increase from near-surface conditions in the foreland autochthon south of Oslo to garnet grade (ca 500°C) in the autochthonous Cambrian metasediments in the Hardangerfjord area (Solli et al. 1978) occurs over a distance of 215 km. Depending on the temperature gradient chosen, and assuming no inverted thermal gradient from overthrust nappes, an average β angle of 4–4.5° can be estimated. This is consistent with similar calculations from Jämtlan–Siljan by Hossack & Cooper (1986). For $\beta = 4$–4.5°, $\rho = 2700$ kg m^{-3} and $h = 15$–20 km, a basal shear stress in the range of 28–42 MPa can be calculated. It is interesting to note that preliminary paleo-stress estimates from recrystallized quartz grain size in quartzites in the Lower Bergsdalen Nappe give very similar values for the flow stress during thrusting (Fossen 1992).

Assuming a constant dip of the décollement during contraction, the geometry of the wedge may be described by the surface slope α. Expressing the equilibrium equation above in terms of α and using a small-angle approximation gives (Elliott 1976)

$$\tau_b = \rho g h \alpha. \tag{2}$$

Any change in the geometry of the wedge or in the basal shear stress will cause internal deformation of the wedge, in which case the basal shear stress must be balanced by two additional terms:

$$\tau_b = \rho g h \alpha + 2 \theta \tau_{xx} + 2 \frac{\partial \tau_{xx}}{\partial x} h, \tag{3}$$

where θ is the angle of taper, and where compressional stresses are considered positive (modified from Platt 1986). τ_{xx} in equation (3) is parallel to the shear direction and the décollement zone. The wedge geometry may change continuously by frontal accretion in the foreland region, leading to lengthening of the wedge and, consequently, to compressional deformation in the frontal part. The geometry of the wedge may also change periodically by subcretion (underplating) of rocks. For instance, the underplating of the Bergsdalen Nappes and possibly also of the WGR must have caused a local increase in the surface slope (α) which must have been balanced by negative terms in τ_{xx} and $\partial \tau_{xx}/\partial x$, and consequently caused extensional deformation at higher levels within the wedge (now removed by erosion). However, top-to-the-southeast deformation would at the same time govern the basal part of the wedge.

Hence, temporal changes in α seem inadequate in explaining the observed regional-scale extension in the décollement zone.

Changes in τ_b may have been important in the development of the Caledonian orogenic wedge in south Norway. Generally, an increase in τ_b will result in contraction and thickening of the wedge, whereas a sufficient decrease in τ_b will cause extension (Dahlen 1984). A decrease in τ_b may be caused by a reduction in subduction rate. Nd–Sm ages of about 410 Ma for eclogites in the western part of the WGR imply continent–continent collision and A-type subduction at this time (Griffin et al. 1985), leading to a decrease and finally vanishing of τ_b. Since there is no longer a subduction-related basal drag to counteract the gravitational stresses, the likely result is extensional collapse of the wedge. The collapse would lead to lateral expansion of the wedge, particularly towards the free (foreland) end where compressional structures therefore would continue to form. Within the wedge, conjugate or complex patterns of normal faults and shear zones are expected. There is, however, no obvious reason for the entire wedge to move back towards the hinterland unless 'space' is created in the central zone of the orogen.

A somewhat different and larger-scale model focuses on the central parts of the orogen which after the continent–continent collision must have been underlain by a thick root of cold, lithospheric mantle (Fig. 14a). The gravitationally unstable root may become detached and descend into the mantle; for example by delamination (Bird 1978) or convection (Houseman et al. 1981) similar to that inferred for the Himalayan orogen (England & Houseman 1988) (Fig. 14b). The resulting increase in surface elevation would generate an increase in vertical stresses which again would exceed the horizontal stresses associated with plate convergence. The result is horizontal extension of the elevated region. Such extensional collapse models have previously been applied to explain the NSD (Norton 1986, Séguret et al. 1989, Andersen & Jamtveit 1990). However, the effect on the foreland-thinning orogenic wedge would be a positive τ_{xx} ('push' from the orogenic center, see equation 3), i.e. continued thrusting rather than back movement. For instance, extensional collapse of a Paleogene collisional ridge in the present-day Alboran Sea is assumed to have driven the coeval, outwardly directed thrusting in the surrounding mountain chains (Platt & Vissers 1989). Similarly, thrusting takes place at the margins of the internally extending/collapsing Tibetan plateau. This model therefore does not explain the regionally uniform Mode I back movement of the Caledonian wedge, which requires a mechanism which can provide space for the nappe wedge to slide towards the central parts of the orogen. The only way this can be done at a regional scale is by a change from convergent to divergent plate motions. The direct reason for the plate divergence is unknown, but may possibly have been triggered or enhanced by the push exerted on the plates by an earlier (pre-D_2) extensional collapse in the central parts of the orogen (Fig. 14b). The back move-
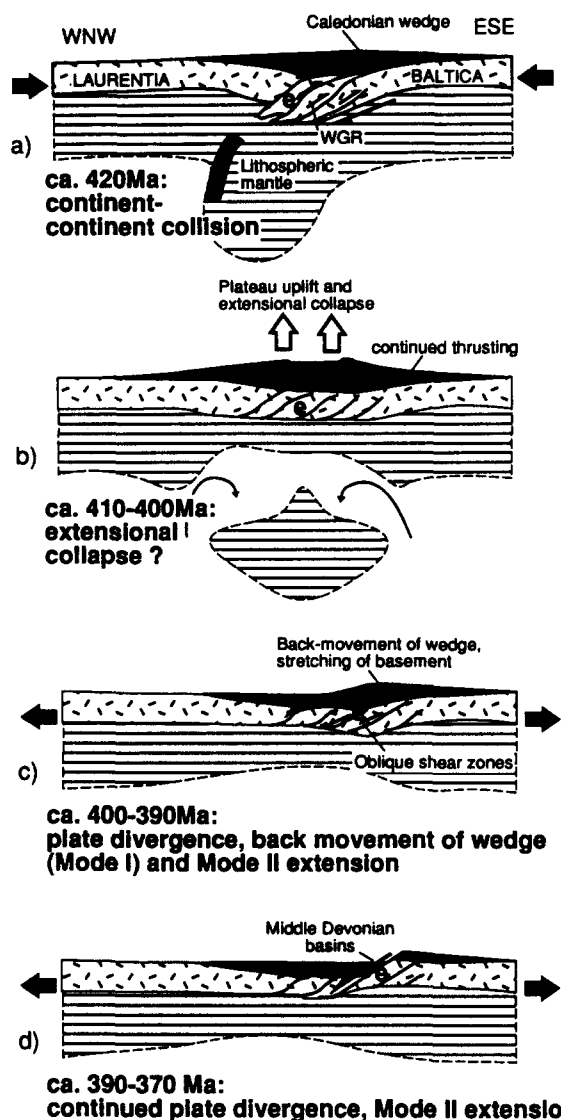
Fig. 14. Schematic model of the late and post-Caledonian development in the Caledonides. An extensional collapse of the central, overthickened part of the orogen (b) would have resulted in continued thrusting in the foreland-thinning orogenic wedge. Back-movement of the wedge (Mode I) extension, (b) and the following basement extension (Mode II, c and d) are here related to a subsequent stage, where divergence of the Laurentian and Baltic continental plates occur.

ment (Fig. 14c) would in this case take place after, rather than during an extensional collapse of the central parts. Finally, the Mode I extension stops, and further extension is taken up by the NSD which offsets the entire nappe wedge (Fig. 14d).

The *timing* of the extensional tectonics in south Norway may be seen in the light of available radiometric and stratigraphic age determinations. The youngest sediments to be affected by thrusting are Lower Silurian (Llandoverian) in the western (coastal) areas, and uppermost Silurian (Downtonian) in the foreland region (Ringerike Sandstone, Bockelie & Nystuen 1985). The collisional thrusting is bracketed upwards by the Middle Devonian flora and fauna (Hoeg 1936, Jarvik 1949) of the Devonian sediments in western Norway. Hence the extensional D_2 deformation must be of Lower to Mid-Devonian age.

Most Ar/Ar mineral ages from the WGR range from

385 to 410 Ma (Lux 1985, Berry 1991, Chauvet & Dallmeyer 1991), and may be interpreted as cooling ages related to the D_2 extension. Although the calibration of the Siluro-Devonian time-scale is not straightforward (e.g. Gale *et al.* 1980), these radiometric ages are consistent with the Lower to Mid-Devonian D_2 age as determined by structural–stratigraphic relationships above. It is interesting to note that sedimentary basins formed on E-dipping extensional or transtensional shear zones in east Greenland at around 385 Ma (K/Ar cooling ages) (Larsen & Bengaard 1991), suggesting that the extension, as well as the preceding contraction, was to a certain extent symmetric with respect to the suture zone of the Caledonian orogen (Fig. 14). It should also be mentioned that a post-contractional, extensional development very similar to the one described in this paper has been recognized in parts of northern Norway (Rykkelid & Andresen 1991, Fossen & Rykkelid in press), emphasizing the significant extent of the D_2 extensional event.

CONCLUSIONS

Extensional deformation in the basal parts of the Caledonian orogenic wedge in south Norway formed a regionally consistent set of structures which overprints earlier, contractional structures. Nowhere is the contractional deformation seen to post-date the extension. The extensional deformation is therefore clearly a post-contractional event, and indicates a regional change from a compressional to a tensional regime. Changes in the boundary conditions of the orogenic wedge, such as decreasing basal shear stress after the Silurian continent–continent collision do not provide a satisfactory explanation for the uniform back movement of the wedge. Furthermore, the D_2 deformation can not be explained by extensional collapse models of the type suggested by Andersen & Jamtveit (1990), since a collapse in the central part of the orogen would have caused continued thrusting towards the foreland, which is contrary to what is actually observed. Only a change from convergent to divergent plate motions, which must have happened in the lowermost Devonian, can explain the uniform transport of allochthonous rocks to the northwest (Mode I) and the accompanying Mode II stretching of the basement.

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