The Devonian basins of western Norway: tectonics and kinematics of an extending crust

Michel Seranne and Michel Seguret

Geological Society, London, Special Publications 1987; v. 28; p. 537-548
doi:10.1144/GSL.SP.1987.028.01.35
The Devonian basins of western Norway: tectonics and kinematics of an extending crust

M. Seranne & M. Seguret

SUMMARY: The Devonian basins of western Norway represent shallow to deep exposures of a synthetic extensional sedimentary basin and provide field evidence for ductile extensional deformation within the basin fill and for the evolution of a brittle low-angle fault and ductile shear zone along the basal contact. The motion along this low-angle (5–25°) detachment is synchronous with both deposition and tilting (25°) of the huge (up to 25-km thick) overlapping coarse detrital Middle Devonian series. Such a geometry requires a minimum dip-slip offset of 50 km.

The structural data are consistent with fault-rock associations along the basal contact and with the prograde greenschist metamorphism observed in the southern basin: deeper and deeper levels are observed from N to S. Except along the highly sheared and retrogressed basal shear zone, the footwall remained unaffected by deformation during basin development.

We discuss three crustal models for basin development and propose that the displacement along the basal contact of the basins is transformed into pervasive ductile flow within the lower crust both at some distance to the side of the basin and beneath the basin.

A key problem for large-scale extensional tectonics is to determine the relationships between brittle faulting near to the surface and ductile stretching at depth, which controls the structure and evolution of extensional sedimentary basins.

Using a theoretical approach, McKenzie (1978) proposed a model for lithospheric stretching. Recently, Kuznir and Park (1984) investigated mathematically the deformation of the intraplate lithosphere. They showed that it is closely dependent on the temperature distribution. Faugere & Brun (1984), analysed the stretching-induced structures in small-scale models. Deep seismic reflection profiles (COCORP; MOIST; SWAT; see Allmendinger et al. this volume; Cheadle et al. this volume) supplied control on the deep crustal structures. They display low-angle normal faults that vanish within the lower third of the crust.

In field examples, such as those in the Basin and Range Province, the accuracy of the observations of the deformation mechanisms should provide better constraints on the models. According to Miller et al. (1983) and Gans & Miller (1983), extensional basins are settled over a ductilely stretched crust and they are limited by a horizontal detachment. Wernicke (1981; 1984) and Bartley & Wernicke (1984) proposed a model of large lithospheric low-angle normal faults which allow shear between two large coherent sheets.

The Devonian basins of western Norway (DBWN) provide a new field example of continental extensional tectonics. They display a continuous section from undeformed sediments at the top of the basin fill, to ductile shearing at the basal contact, involving both Devonian metasediments and basement rocks (Seranne & Seguret 1985a). Observations made at different erosional levels aid determination of the geometry, tectonics and kinematics of these extensional basins.

Geological setting

The DBWN include from N to S, the basins of Hornelen, Hasteinen, Kvamshesten and Solund (Fig. 1). They are located on the western side of the Caledonian thrust belt. During the Caledonian Orogeny, Cambro–Silurian rocks related to a passive margin and oceanic origin were thrust eastwards (Hossack & Cooper, in press) above the Scandinavian craton consisting of Precambrian gneisses now outcropping within the More window (Santarelli 1977). The basal contact of the DBWN represents a re-working of the thrust plane as a westwards-sliding normal fault (Hossack, 1984; Norton, 1984; Seranne & Seguret 1985a) (Fig. 1).

Sedimentology

In the Hornelen and Kvamshesten basins, the coarse alluvial fan conglomerates outcropping along the present borders distally grade into alluvial/lacustrine sandstones in the axis of the basins (Steel & Gloppen 1980). In these basins both types of deposits are well organized into 50–200-m thick coarsening- and thickening-upward cycles. It has been demonstrated that sedimentation was largely controlled by tectonics related to the basin formation (Brynhi & Skjerlie 1975; Steel 1976).

The Solund and Hasteinen basins are mostly filled with unsorted and ungraded conglomerates, the grain size of which ranges from very coarse sandstone to cobble. Nilsen (1968) and Indreaver (1980) interpreted these deposits as being formed in humid
alluvial fans. The occurrence of sieve and stream flood deposits confirms this interpretation. The pebbles are derived from metamorphic rocks of Cambro-Silurian origin (greywacke, quartzite, siltstone and gabbro) or Precambrian origin (quartzites and gneisses). In addition, very rare pebbles of reddish siltstones are thought to be re-worked material from intrabasinal, fine distal facies of Devonian age.

The bodies of gabbro found in N Solund were interpreted as thrust sheets (Kildal 1970) or as intrusive (Nilsen 1968) but the relationship with the conglomerate argues for an olistolithic origin (Brynhi 1976).

**Geometry**

**The borders**

The basins represent a homogeneous structural setting (Figs 1 & 2). Along the western borders, the Devonian series lies unconformably on the Cambro-Silurian metamorphic rocks. The present eastern margins are low-angle (5–25°) westward-dipping tectonic contacts traditionally interpreted as eastward-moving thrusts (Nilsen 1968; Kildal 1970; Hoisaeter 1971; Steel & Gloppen 1980; Roberts 1983; Sturt 1983). The low-angle eastern contact evolves along the N and S borders into steeper faults (45–60°) that define a spoon-shape geometry to the basal contact of each basin. This geometry is well preserved in the Hornelen basin in which the present steep N and S borders probably closely fit the original margins (Steel & Gloppen 1980). In contrast, the low-angle faults presently limiting the Solund basin indicate a structure truncated by erosion. Therefore, the Hornelen basin represents a shallow exposure of the DBWN, whereas the Kvamshesten and Solund basins represent deeper exposures of the DBWN. Hossack (1984) suggested that the basal contact of the four basins could constitute a unique westward-sliding normal fault with lateral ramps related to E–W-trending culminations. Our field data support these
assumptions. The Hasteinen basin exhibits a slightly different structure; except for the western unconformable boundary, all the margins are sinistral strike-slip faults, as indicated by microstructures.

Internal organization of the basins

The Devonian strata generally dip towards the eastern normal fault and at a constant 25° along the axial sections of the basins. The dip decreases, however, close to the eastern normal fault, but increases close to the lateral margins where the strikes tend to be parallel to the faults (Fig. 1).

In the Hornelen and Kvamshesten basins, unconformities with local erosion have been identified in the steeply dipping marginal conglomerates, associated with a significant thickening of the cyclothems towards the basin axis (Bryhni & Skjerlie 1975). This demonstrates an inhomogeneous and syn-depositional subsidence. The syncline geometry does not require a later N-S compressional event. The present great stratigraphic thickness of the basin successions (25 km in Hornelen (Bryhni 1964), 7 km in Kvamshesten, 6 km in Solund) strongly contrasts with their small area (2500 km² for the larger one: Hornelen); the stratigraphic thickness does not represent the depth of the basin. The vertical section of sediments above the low-angle fault is observable at Kvamshesten Basin; it does not exceed 1500 m. It is likely that the depth of the basal contact beneath the depositional surface did not exceed 5–10 km (Fig. 2).

Motion along the basal contact

The geometrical features of the basins allow an estimate of the amount of motion along the basal contact. A dip of 25° E of the basal Devonian unconformity, a 65 km length for Hornelen basin and a fault plane dipping westward at an angle of 10°, integrating the deep structural data from the other basins, require a minimum dip-slip offset of 50 km for this basin (Figs 2 & 3). From the analysis of branch-lines and fault cut-off lines, Hossack (1984, Figs 4 & 5) found a similar displacement value.

During basin development, the syntectonic sedimentation occurred in a depocentre located along the eastern fault margin, overlapping the earlier sediments (Fig. 3). The rollover in the hanging wall caused tilting of the Devonian sediments and the motion along the very low-angle fault resulted in a westwards translation of the series and preserved their constant dip of 25° (Fig. 3).

Structural analysis

Hornelen and Kvamshesten basins

Hornelen and Kvamshesten sandstones and conglomerates display an intense fracturing and jointing. Three main orientations can be defined (N 40° E, N 110° E, N 170° E); generally no relative motion, dip-slip or strike-slip, can be determined at outcrop scale along these joints.
Some steeply dipping bedding planes reveal pebbles with long axes re-oriented parallel to the dip-slip direction (Brynhi 1978) and interbedded siltstones show striations on bedding planes in the maximum-dip direction. A fine cleavage often indicates interbed dip-slip movement. Such brittle deformation on a discrete plane results in a finite extension broadly parallel to the axis of the basin and an associated vertical shortening. In these basins, no penetrative deformation is observed; the pebbles are broken only very close to the basal contact.

**Solund basin**

The conglomerates of the Solund basin contain an outstanding basin-wide anisotropic fabric, interpreted by Nilsen (1968) as a sedimentary fabric. More detailed studies, however, show a relationship between the fabric evolution and that of the regional deformation. This allowed us to define four structural zones from the centre of the basin to the SE basal contact (Fig. 4).

**Zone 1**

In the centre of the basin, the bedding dips at 25° SE and the landscape is dominated by the repetition of ledges which interrupt the conglomerate scarps. A typical morphological sequence, 10–50 m thick, is composed of a basal division (D1) of poorly consolidated conglomerates, a few metres thick, grading up into a well-consolidated upper division (D2) (Fig. 5).

The uppermost part (D2b) of D2 is half a metre thick and is characterized by pebbles lying flat on the bedding plane which show no preferred orientation. Within the rest of the morphological sequence, that is in D1 and D2a, representing most of D2, the pebble long axes trend N 120° and plunge 15° steeper than the bedding (Nilsen 1968). The fabric is well exposed in the poorly cemented division (D1) and has been found from systematic measurements in D2a. Some considerations and observations argue against a sedimentary origin for this fabric; a sedimentary fabric would not display so constant direction through time and space, and the consistency of pebble orientation contrasts with the fluctuating palaeocurrent directions indicated by sedimentary structures in the scarce sandstone bodies (Nilsen 1968). In addition, the associated microstructures (Fig. 6) argue for a tectonically induced fabric.

A N 120° trending mineral lineation results from the re-orientation of phyllites and formation of chlorite–quartz–epidote. This lineation develops on the pebble surfaces as striated lineations, whose orientations vary according to the pebble shape. Systematic analysis reveals a mean N 120° orientation and a dip-slip motion on the upper and lower pebble surfaces. The continuation of the striated lineation at the tip of the pebbles shows horizontal conical pressure shadow like structures, trending N 120° and corresponding to a zone of cementation of the sand and gravels (Fig. 6). An horizontal cleavage developed in the sandy matrix and in the scarce reddish marl pebbles which are fine re-worked layers of the Devonian alluvial fan. In the sandy matrix the cleavage is marked by the orientated clasts and by the synkinematic growth of quartz–chlorite fibres in pressure shadows. In the mudstone pebbles, the schistosity, the initiation of which was clearly synchronous with that of the regional Devonian cleavage, is marked by the re-orientation of phyllites and orientated growth of chlorite, epidote and actinolite.

In addition, many of the large Precambrian or Caledonian pebbles contain vertical, N 30° striking, tension gashes (Fig. 4), filled with horizontal fibres of quartz and calcite (+ epidote, chlorite) and the conglomerates are cut-off by a number of dip-slip microfaults. Stress tensor determinations (Fig. 7) from microfault analysis employing the automatic method (Etchecopar et al. 1981) give a direction of minimum stress, $\sigma_3$, orientated horizontal and trending N 120°; a direction of maximum stress, $\sigma_1$, vertical and a ratio $R = (\sigma_1 - \sigma_3)/(\sigma_1 - \sigma_2) = 0.75$, i.e. $\sigma_1$ close to $\sigma_2$. The Mohr circle gives the relative values of $\sigma_1$, $\sigma_2$, $\sigma_3$ and the normal and shear stresses for each
The Devonian basins of western Norway

fault plane. The unusual distribution is a result of the un lithified state of the conglomerate (Seranne & Seguret, 1985b).

Zone 2
In this zone the morphological sequence is not expressed, as the poorly cemented division D1 diminishes laterally southeastwards. The bedding dip and the imbrication angle decrease (Fig. 8), but the preferred orientation of the pebbles strengthens. The same microtectonic structures as in zone 1 are evident from detailed field observation and thin-section studies.

Zone 3
The bedding dip is flatter and there is a horizontal foliation. There is no imbrication, but the pebble preferred orientation is stronger both on bedding planes and on sections striking N 120°. The change in pebble shape suggests internal deformation of the pebbles. This deformation is controlled by lithology and develops as boudinage, ductile deformation, and sliding along pre-existing internal heterogeneities, such as strata, cleavage and foliation.

Sets of low-angle NE-striking, metre-scale dip-slip shear zones, which dip gently to the NW or the SE and flatten downward, affect the general fabric.
FIG. 5. Morphological sequence and pebble fabric (see text for explanation). The orientations of about 50 pebble long-axes are plotted in each circular diagram referred to the N on the bedding plane and referred to horizontal in the vertical planes. The 25° tilting of the bedding has been removed.

FIG. 6. Microtectonic structures associated with the pebbles.

FIG. 7. Example of a stress tensor given by microfault population analysis employing an automatic method (Etchecopar et al. 1981) (Schmidt net, lower hemisphere) and Mohr circle. The $R$ ratio characterizes the shape of the stress ellipsoid (for location see Fig. 4).

FIG. 8. Schematic cross-section of the Solund basin showing the decrease of the bedding dip and imbrication angle toward the basal contact. Each circular diagram represents about 40 measurements and gives the mean imbrication (stippled line). (For location see Fig. 4.)
They are characterized by the re-orientation of pebbles to the dip-slip direction or by the deflection of the horizontal foliation and they are associated with asymmetrical pressure-shadows around quartz rods. They indicate a conjugated normal sense of motion and a horizontal NW–SE extension.

Some vertical quartz veins, striking N 30°, are folded symmetrically with respect to the horizontal foliation. Giant synkinematic porphyroblasts of epidote (0.20 m mean size, up to 2 m) lie within the horizontal foliation. A crack-seal origin for these porphyroblasts is suggested by horizontal epidote–quartz fibres developed between closely spaced vertical joints (J.P. Brun, pers. comm.). The long, intermediate and short axes of the porphyroblasts are respectively horizontal N 120°, horizontal N 30° and vertical.

Zone 4

A few metres above the basal contact, pebbles are difficult to recognize. There is a grain-size reduction, however, and an increase in clast elongation towards the Devonian/Cambro–Silurian contact (Fig. 9). The foliation, defined at outcrop scale by the preferred orientation of the deformed clasts, becomes parallel to the contact, which dips 10° NW, and some mylonitic bands occur parallel to the foliation. They are both affected by C' shear planes (Berthe et al. 1979), dipping 35° NW. In the foliation plane some clasts are broken up into fragmental trails, parallel to the N 120° stretching lineation. The contact itself is marked by a green cohesive schistose rock.

Beneath the contact, the Cambro–Silurian rocks display a foliation parallel to the contact. Some quartz ribbons may be folded into intrafolial recumbent folds. The foliation plane is slightly undulated by oblique shear planes (C'), which dip towards the NW (Fig. 10). The stretching lineation, trending N 120°, developed within the foliation and some NW-verging sheath folds are observed.

Further down, the foliation becomes horizontal and the S, C and C' structures become less important. The intensity of the foliation decreases southeastward and, a few 100 m below the contact, it appears to be superimposed on earlier structures related to a WNW-directed ductile shearing of high stress.

---

**Fig. 9.** Evolution of the clasts near the basal contact of the Solund basin. Each plot represents the mean values of about 20 measurements. (a) Section across the contact, (b, c) grain-size reduction, (LA=Long axis; SA=Short axis), (d) stretching of the clasts, (e) reduction of imbrication angle θ (for location see Fig. 4).

**Fig. 10.** Example of shear bands (C') within the basal contact of the Solund basin, that indicate a northwestward shearing.
strain. A detailed study of the basement deformation is not the subject of this paper, however, we assume that three major events can be inferred: (i) Caledonian nappe emplacement; (ii) late-Caledonian westward ductile shearing, homogeneous over the area; and (iii) heterogeneous Devonian extensional deformation with westward-directed brittle to ductile shearing restricted to a few hundred metres thick zone at the basement/Devonian interface with brittle fracturing below the detachment.

**Structural interpretation of the Solund section**

In zone 1 the brittle microstructures (microfaults and intrapebble tension gashes) give a stress tensor with $\sigma_3$ horizontal N 120°, $\sigma_2$ horizontal N 30° and $\sigma_1$ vertical. The ductile microstructures (lineation, pressure-shadows, cleavage) result from a finite strain with a N 120° horizontal extension axis (X), a N 30° horizontal intermediate axis (Y), and a vertical shortening axis (Z). The nature of the Y direction (shortening or extension) has not been determined and the strain-ellipsoid shape is not precisely known. There must, however, be a very long extension (X) axis.

The close relationships between the conglomerate fabrics and microtectonics demonstrate that the fabric is not of sedimentary origin. The flattening post-dates both the tilting of the bedding, as the strain and stress axes are not tilted, and the acquisition of the fabric, as the surface of imbricated pebbles are used as normal faults. In fabric-bearing conglomerates, in D1 and D2a divisions of the morphological sequence, the pebbles are undeformed apart from the tension gashes and the fabric has been attained by body rotation of the pebbles in an unconsolidated sandy matrix. The characteristics of the microfaults (Guiraud & Seguret, in press) fit this rheological interpretation. The interbedded levels in the $D2b$ division that do not exhibit a fabric are considered to be previously consolidated layers with a different rheological response to stress.

We propose that the fabric is linked to the tilting of the series and is the result of a southeastwards dip-slip shearing of soft conglomerates (Fig. 11). The shearing occurred parallel to the bedding and was controlled by the previously consolidated levels in the $D2b$ division, which acted as thin undeformable planes interbedded into the soft conglomerate (Seranne & Seguret 1985b). The flattening is superimposed on the shearing. The pebble imbrication angle results from an equilibrium between rotation synthetic to shearing, related to the tilting, and antithetic rotation, related to the flattening. Both shearing and flattening are induced by the vertical maximum stress and the N 120° horizontal minimum stress produces the extension. It has not been possible to establish a relationship between the morphological sequence and a possible sedimentary sequence due to the strong deformation and the poor knowledge of such deposits.

In zone 2, the same structural data as in zone 1 suggest the same deformation mechanisms. The decrease of bedding dip and angle of imbrication results from the down-section increase of flattening (Fig. 11). Bedding and pebbles are rotated and tend to parallel the finite flattening plane; the horizontal schistosity.

In zone 3, rotational deformation criteria have been observed in sections trending N 120°. At outcrop scale, however, the rotational effects are balanced in conjugate structures and the growth of epidote porphyroblasts and the symmetry of the folded quartz veins argue for a finite strain resulting from a coaxial deformation. All the structural data can be interpreted as resulting from a coaxial deformation.
The Devonian basins of western Norway 545
deformation with principal directions X, Y, Z orientated respectively; horizontal N 120°, horizontal N 30° and vertical.

In zone 4, the reliable sense of shear markers indicate that the strain regime is not coaxial. The deformation history of the rocks within the basal contact is of progressive shearing towards the NW, parallel to the contact between Devonian and Cambro–Silurian rocks, with maximum shearing along the contact and motion parallel to the N 120° trending lineation. The microstructures of zone 4 are therefore in agreement with it being a shear zone.

Evolution of the deformation along the DBWN basal contact

From N to S, the Hornelen, Kvamshesten and Solund basins show a different pattern of rock textures at the basal contact, between the Devonian series and the basement. In the northern basin the conglomerates are brecciated through a zone a few metres thick; the contact is marked by a schistose cataclasite (Chester et al. 1985) generated by cataclasism on densifying discrete shear planes. This brittle deformation, linked to the formation of the basin, has not affected the well-developed mylonitic fabric of the basement. There is a textural and metamorphic gap between footwall and hanging wall rocks. At the Kvamshesten eastern basal contact, the Devonian Series exhibits the same structures as at Hornelen. The schistose cataclasite is underlain by a 0.3 m thick black glassy rock that locally intrudes fractures, thus suggesting a pseudotachylite. The underlying basement consists of polydeformed chloritic gneisses with a consistent mylonitic foliation, showing plastic deformation and recrystallization (Tullis et al. 1982). This is overprinted by some discrete cataclastic shear planes that become denser towards the contact. The brittle deformation related to basin formation affected both Devonian and basement rocks. It is still dominated by brittle cataclasis but denotes a higher shear strain than in Hornelen. In contrast to the former examples, the conglomerates of Solund basin are epimetamorphic and ductilely deformed. The fault rocks still resemble the schistose cataclasites of the Kvamshesten basin, but the extremely fine-grained matrix contains bands of dynamically recrystallized quartz. In the basement, the same close relationships between ductile and brittle processes are observed. There is no tectonomorphic gap between hanging wall and foot wall.

At regional scale, the footwall consistently demonstrates the results of ductile deformation. Close to the contact this deformation is progressively overprinted by a younger event the effects of which are found both in the basement and in the Devonian Series. This younger event, thus related to the basin formation, displays fault rocks produced under increasing P/T conditions, from Hornelen to Solund. This allows us to observe the transition from a brittle to ductile mode of deformation (Sibson 1977; Mitra 1984). We conclude that the observed basal contact shows deeper and deeper exposures from N to S.

Metamorphism and magmatism

In Solund basin the deformation is associated with a synkinematic prograde metamorphism and hypovolcanic intrusions. The petrological studies are still in progress and we present only some preliminary results.

The conglomerate constituents are mostly Precambrian gneisses, and Caledonian metasediments and metabasites which have been metamorphosed into the amphibolites facies. Devonian metamorphism is expressed by; (i) an incomplete retrogression of the gneissic pebbles into the greenschist facies; and (ii) crystallization of index metamorphic minerals in the sandy matrix and into the scarce reddish siltstone pebbles of Devonian age. The blastesis is clearly synchronous with the regional Devonian foliation.

The occurrence of prehnite, zoisite, clinozoisite, chlorite, albite and quartz, with possibly pumpe-
tachylite, characterizes a prehnite–pumpellyte facies (200–300°C). The occurrence of actinolite marks the beginning of the greenschist facies (300–350°C). The succession of mineral assemblages is typically of low-pressure type (P < 3 Kb) but barometers are poorly defined at low pressure. The metamorphism in the Devonian basins is very unusual, being clearly synkinematic with the extensional tectonics. In that way it differs from other basin-related metamorphism (Muffler & White 1969; Guiraud & Seguret in press), in which minerals are mostly static. In addition, trachytic and rhyolitic lavas are interbedded with the Solund conglomerates (Sigmond et al. 1984; Furnes & Lippard 1983) and veins and small bodies of rhyolitic rocks occur within the conglomerates of the Solund basin. The veins and bodies have a granitic composition, with quartz and feldspars and a granophyric texture, typical of hypovolcanic rocks. The bodies (0.1 to 2 m long) have poorly defined and contorted margins. Along the bedding plane they are elongated and well orientated with a similar fabric to that of the pebbles. At the present stage of investigation, we interpret these granitic bodies as synkinematic hypovolcanic intrusions. Only the most differentiated types occur; this may indicate a crustal origin by partial melting but more geochemical data are needed to confirm this.
Kinematic interpretation

Integration of the data obtained from the four basins indicates that deformation of the Devonian sediments resulted from vertical flattening, accommodated by the westward motion of the hanging wall. It consisted of brittle deformation (i.e., interbed sliding) in the upper part of the section, now outcropping at the highest exposed levels (Hornelen, Kvamshesten). At deeper levels (Solund, zones 1 and 2), the conglomerates are ductilely sheared and flattened with pebble-body rotation in a ductile matrix; the eastward shearing is still controlled by the tilted bedding. Close to the bottom of the basin (Solund, zone 3) the flattening increases and the pebbles are ductilely deformed. The basal contact is associated with dip-slip shearing towards the NW (or W), involving both Devonian and basement material. Below this shear zone, the basement was brittlely deformed during the development of the basins. Metamorphism and fault-rock associations are consistent with the structural data: deeper and deeper levels are observed from N to S.

Discussion on crustal scale model

One of the main acquisitions of this study is the observation of ductile deformation in the basin fill and on the detachment, over a basement undeformed during basin development. The large offset of the low-angle basal shear zone needs to be accommodated somewhere in the crust or in the lithosphere.

A model associated with a horizontal detachment (Miller et al. 1983; Kligfield et al. 1984) is not realistic due to; (i) the uniform sense of shear along the detachment and (ii) the huge amount of extension that would be required (stretching factor $\beta = 4.4$, McKenzie 1978) without intense and generalized plutonism and magmatism.

In contrast, all the data support a model with a dipping detachment (Kligfield et al. 1984). In a first hypothesis (Fig. 12a), the basal shear zone deepens westward into the crust. There is no ductile deformation of the lower crust and therefore the displacement is preserved down-slip and the offset affects the Moho discontinuity. The deep structure

![FIG. 12. Crustal models of extension for the DBWN (see text for discussion). Large stipples = continental crust; irregular small dots = syntectonic deposits; dashes parallel to Moho discontinuity = continental crust ductilely stretched.](image)

The sketch illustrates the case of a unique low-angle detachment that implies the development of a large hanging wall syncline, W of the rollover anticline.
may be controlled by a decoupling at the Moho as proposed by Wernicke (1981). In a second hypothesis (Fig. 12b) the basal shear zone deepens westwards within the crust down to a depth where P/T conditions produce ductile deformation. The horizontal displacement observed in the upper part of the crust is transformed into pervasive ductile stretching of the lower part of the crust at some distance W of the basin. The crustal thinning does not occur directly beneath the basin. In a third hypothesis (Fig. 12c) the stretching is distributed both beneath the basin and further to the W. In this case, the basin and its undeformed basement overlie a stretched and thinned lower crust. The model with a dipping detachment is also supported by a comparison with deep seismic reflection profiles (Allmendinger et al. 1983; Allmendinger et al., this volume; Brewer & Smythe 1984). These profiles display low-angle reflectors that limit downward-tilted strata of sedimentary basins. Most reflectors are traceable into the upper crust and merge downward into horizontal reflectors, a feature of the lower third of the crust, which may represent the zone of ductile stretching. The transfer of stretching in the lower crust into shearing along the detachment is facilitated by reactivation of pre-existing planes of strength anisotropy, such as thrust planes (Brun & Choukroune 1983). The retrogressive dip-slip shear zone of the DBWN results from the re-working of a previous eastward-dipping Caledonian thrust plane (Hossack 1984; Norton 1984) and/or a late-Caledonian westward shearing. Finally the DBWN are likely to represent the eastern margin of the Devonian North Sea basin (Fig. 13) (Ziegler 1981), symmetrical to the N Scotland Devonian basin in which a similar geometry has been revealed by the MOIST profile (Brewer & Smythe 1984).

ACKNOWLEDGMENTS: This research was supported by Elf-Aquitaine Norge: we wish to thank J.M. Golberg and Ph. Laurent for their contribution concerning metamorphism and fault-rock associations, respectively. Fruitful discussions with colleagues at the laboratories of Petrology and Structural Geology and with J.P. Brun during a field trip, improved the paper. We are grateful for the comments of two anonymous referees.

FIG. 13. Proposed section of the North Sea during middle Devonian times. The symmetrical low-angle shear zones bounding the basins at each margin merge in the lower crust within a zone of pervasive ductile flow (not to scale).

References


CHEADLE, M.J., MCGEARY, S., WARNER, M.R. & MATTHEWS, D.A. This volume. Extensional
structures in the western UK continental shelf; a review of evidence from deep seismic profiling.


