Journal of Structural Geology, Vol. 14, No. 5, pp. 547 to 563, 1992
Printed in Great Britain

0191-8141/92 $05.00+0.00

Variscan transpressive inversion in the northwestern central Rhenohercynian belt of western Germany

WERNER FIELITZ

Geologisches Institut, Universität Karlsruhe, Kaiserstrasse 12, 7500 Karlsruhe, Germany

(*Received* 28 *January* 1991; *accepted in revised form* 19 *November* 1991)

Abstract—Important orogen-parallel strike-slip components during the convergence of the Variscan orogen of Central Europe have been increasingly recognized. These oblique-slip faults often develop from pre-existing tectonic boundaries and are influenced by pre-existing structural trends in the basement. In this paper evidence is presented for a 3–6 km wide distributed shear zone (here named the Monschau shear zone) in epizone metamorphic Cambro-Ordovician and Lower Devonian clastic rocks of the northwestern Rhenohercynian belt. Kinematic indicators including field observations of anomalous mullion and fold axes patterns, a set of two cleavages, stretching lineations, and shear and kink bands, microscopical observations of stretching-parallel asymmetric pressure shadows of quartz on prekinematic, epizone metamorphic pyrite porphyroblasts, and microfabrics of phyllosilicates by X-ray texture goniometry point to a sinistral tranpressive shear zone with emplacement of the southeast hangingwall under retrograde conditions. Coincidence of the anchizone–epizone metamorphic boundary with the change along strike from penetrative shearing to distributed thrusting suggests a ductile–brittle transition. The Monschau shear zone is situated on the backlimb of an allochthonous anticlinorium located above a basement ramp that corresponds roughly to a pre-orogenic synsedimentary normal fault. Its continuation shows up clearly in a DEKORP seismic reflection profile to a depth of 12–15 km.

INTRODUCTION

In recent years oblique shortening has been recognized increasingly as significant in the kinematics of orogenic belts (e.g. North American Cordillera, Coney *et al.* 1980, Oldow *et al.* 1989; Appalachians, Williams & Hatcher 1983, Rast 1989; North European Caledonides, Hutton 1987; New Zealand Alps, Korsch & Wellman 1988). The displacements connected with transpression are concentrated on discrete strike-slip faults or distributed throughout wider shear zones. Whereas faults are relatively easy to map, evidence for distributed oblique shear is difficult to document (Harland 1971, Sanderson *et al.* 1980, Sanderson & Marchini 1984, Woodcock 1990). It has become evident that oblique-slip faults often develop from pre-existing tectonic boundaries and are influenced by pre-existing structural trends in the basement. One important possibility for the reactivation of pre-orogenic structures is the inversion of synsedimentary extensional faults during orogenic shortening (Butler 1989, Cooper *et al.* 1989, Hayward & Graham 1989, McClay 1989, Williams *et al.* 1989).

The convergence of Gondwana and Laurasia between 400 and 280 Ma ago created the Variscan orogen of Central Europe. The orogen has been subdivided into several tectonometamorphic belts separated from each other by major faults (index map of Fig. 1). For many years this orogen was discussed as resulting from convergence perpendicular to its general NE–SW oriented structures. In recent years indications for important orogen-parallel strike-slip components have been found (e.g. Arthaud & Matte 1977, Berthé *et al.* 1979, Holder & Leveridge 1986, Matte 1986, Ziegler 1986, Krohe & Eisbacher 1988). Strike parallel tectonic movements have also been reported from the allochthonous Rhenohercynian belt of the Mid-European Variscan orogen. For example, a counterclockwise rotation of the paleostress field (Oncken 1988a,b) implies the possibility of dextral strike-slip displacements. At the southern boundary of the belt dextral strike-slip faults have been described by Schwab (1987). In addition, synsedimentary pre-shortening normal faults oblique to regional structural trends are widespread in the Rhenohercynian belt (Oncken 1988a,b, Meilliez 1989, Werner 1989, Meilliez & Mansy 1990).

This paper focuses on an area of the northwestern central Rhenohercynian belt which contains a major Lower Devonian synsedimentary normal fault zone that coincides with a several kilometre wide zone of subsequent penetrative deformation (Figs. 1 and 2). This zone, situated in interbedded pelitic and psammitic clastic rocks, also shows up clearly in a DEKORP deep seismic reflection profile (DEKORP Research Group 1991). This study attempts to document the evolution of the synsedimentary fault zone into a ductile–brittle oblique inversion structure, the large-scale regional kinematic significance of which is shown by reflection seismic work.

GEOLOGICAL SETTING

In western Germany the northernmost belt of the Variscan orogen consists of the Subvariscan Late Carboniferous clastic foreland basin which is situated in front of the allochthonous Rhenohercynian belt (index map in Fig. 1). In the Rhenohercynian belt, Lower Devonian to Lower Carboniferous sedimentary rocks rest unconformably on pre-Devonian Lower Paleozoic clastic rocks which crop out in several anticlinal struc-

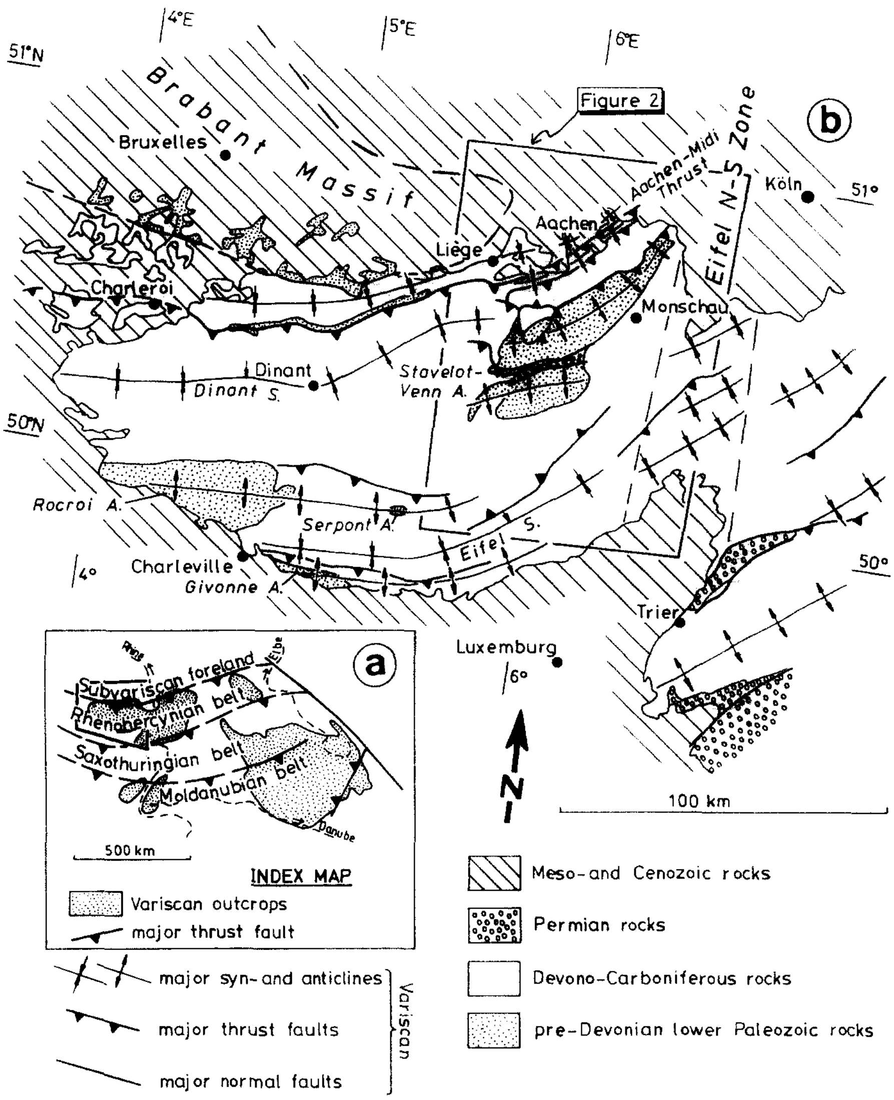

Fig. 1. Index map and major structures in the central Rhenohercynian belt of western central Europe and location of the study area.

tures (Fig. 1). The geometry and significance of earlier 'Caledonian' deformation in these rocks are still debated (e.g. Geukens 1961, Richter 1962, Albrecht 1971, Michot 1976, Delveaux de Fenffe & Laduron 1984, Fielitz 1987). The Late Carboniferous deformation of the Rhenohercynian rocks created NW-verging folds, a closely spaced slaty cleavage and thrust faults (Bless *et al.* 1980, Weber & Behr 1983, Walter & Wohlenberg 1985, Oncken 1988a,b, Meilliez & Mansy 1990). Intensity of deformation decreases northwestward. The age of penetrative deformation ranges from 327 Ma in the south to 305 Ma in the north (Ahrendt *et al.* 1983). The Rhenohercynian slates generally reflect diagenetic, anchizone to locally epizone metamorphic conditions (Weber 1976, Teichmüller & Teichmüller 1979, Teichmüller *et al.* 1979, Oncken 1984, Frank 1987). Relatively high pressure–low temperature metamorphism has been reported only from the southernmost Rhenohercynian domain along the tectonic border with the complex terranes of the Saxothuringian belt (Massonne & Schreyer 1983).

In the area investigated the Rhenohercynian belt contains the transverse Eifel North–South zone (Fig. 1), a paleogeographic feature that can be traced as far back as the Lower Devonian (Emsian) and which was intermittently active until at least Early Jurassic time, possibly also into Early Tertiary time (Meyer 1986, p. 506f). West of the Eifel North–South zone the central Rhenohercynian belt is separated from the Subvariscan belt along the Midi–Aachen Thrust fault zone, south of which there are several large anticlines with pre-Devonian Lower Paleozoic rocks in their cores. These are the Condroz, the Stavelot–Venn, the Rocroi, the Serpont and the Givonne anticlines and anticlinoria. Towards the west the general NE–SW trend of the Variscan structures in central Europe changes to an E–W trend. This change in the structural trend is probably related to the southern border of a pre-Variscan subsurface basement high in the foreland, the Brabant Massif (Michot 1980) (see Fig. 1). The Brabant Massif was affected by Caledonian deformation and is covered by a thin or incomplete Upper Paleozoic platform succession.

The Stavelot–Venn anticlinorium is the largest pre-

Devonian Lower Paleozoic outcrop area inside the western central Rhenohercynian belt (Figs. 1 and 2). It is bordered towards the southeast by Devonian rocks of the Eifel synclinorium, to the east by the axial depression of the Eifel North–South zone, to the northwest by Devono-Carboniferous rocks cut by several thrust faults, and to the west by the broad Dinant synclinorium. Stratigraphically, the core of the Stavelot–Venn anticlinorium consists of Lower Cambrian to Lower Ordovician quartzites, sandstones, banded shales and shales which underwent Caledonian, probably Early Caradocian (Michot 1976) deformation (e.g. Geukens 1961, Richter 1962, Albrecht 1971, Fielitz 1987). Small outcrops of rhyolites, basalts, and Middle Devonian granodioritic intrusives (Kramm & Buhl 1985) belong to an intracontinental domain with a tholeiitic differentiation trend (André *et al.* 1986). To the northwest of the anticlinorium an Upper Paleozoic clastic suite of conglomerates, arkoses, sandstones, siltstones and shales of Early Devonian to Eifelian age is overlain by limestones of Givetian to Frasnian age, sandstones of Famennian age, and a second carbonate suite of Early Carboniferous age. Synorogenic Upper Carboniferous shales, sandstones, conglomerates and coal seams of the foreland basin terminate the Paleozoic stratigraphic sequence. To the southeast of the Stavelot–Venn anticlinorium 3000–6500 m of Lower Devonian shale and greywackes contrast with 500–700 m of Lower Devonian sediments to the northwest (Knapp 1980).

Structurally, the Stavelot–Venn anticlinorium con-

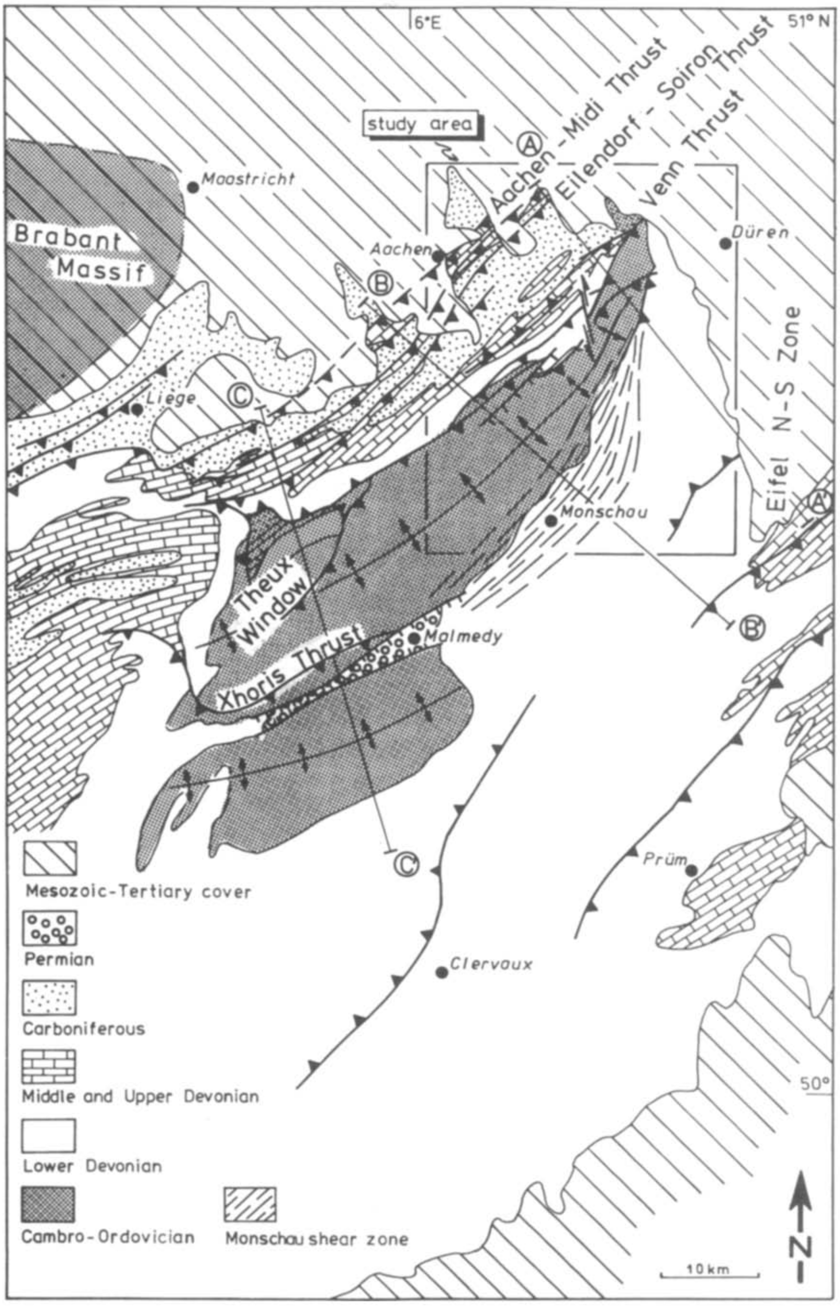

Fig. 2. Geologic sketch map of the Stavelot–Venn anticlinorium and adjacent areas with location of the study area. The lines A–A′, B–B′ and C–C′ indicate the location of the profiles in Fig. 12 and line B–B′ also the DEKORP 1A deep seismic reflection profile of Fig. 14. For tectonic symbols see Fig. 1.

sists of two anticlines separated by a narrow syncline and a small superimposed Permian graben structure (Fig. 2). The SE limbs of both anticlines are intensely deformed and metamorphosed (Fieremans & Bosmans 1982, Kramm *et al.* 1985b, Frank 1987). As indicated by prehnite–chlorite–epidote mineral assemblages in a metagranodiorite (Schreyer & Abraham 1979) and the occurrence of chloritoid and pyrophyllite in metapelites (Kramm *et al.* 1985b), epizone metamorphic conditions with up to 325/345–385/415°C and 1–2 kbar were reached in the area of the northern anticline. Kramm *et al.* (1985a) determined an age of 308–312 Ma for the last structural reorganization of the phyllosilicates. This deformed and metamorphosed zone which coincides with the area of dramatic thickness increase of the Lower Devonian succession is the site of a major oblique shear zone, here named the Monschau shear zone (Fig. 2). Because it has been imaged by deep crustal reflection seismic investigations (DEKORP Research Group 1991) to a depth of at least 4 s TWT its surface structure and kinematics were investigated in detail.

MINOR STRUCTURES IN THE MONSCHAU SHEAR ZONE

The Monschau shear zone along the SE limb of the Stavelot–Venn anticlinorium (Schmidt 1956, Spaeth 1979, Fielitz 1987) is characterized by the occurrence of well-developed small-scale folds, cleavage, stretching lineation, shear and kink bands, and quartz veins that originated in rocks showing epizone metamorphic mineral assemblages.

Small-scale folds

Small-scale folds in the Monschau shear zone generally trend and plunge easterly and verge to the NNW. Their trend differs distinctly from the general NE–SW trend of the major folds of this part of the Rhenish Massif, including the Stavelot–Venn anticlinorium (Knapp 1980). In detail the E–W-trending minor folds (Figs. 3 and 8b) occur mainly within the transition zone between pre-Devonian and Lower Devonian rocks

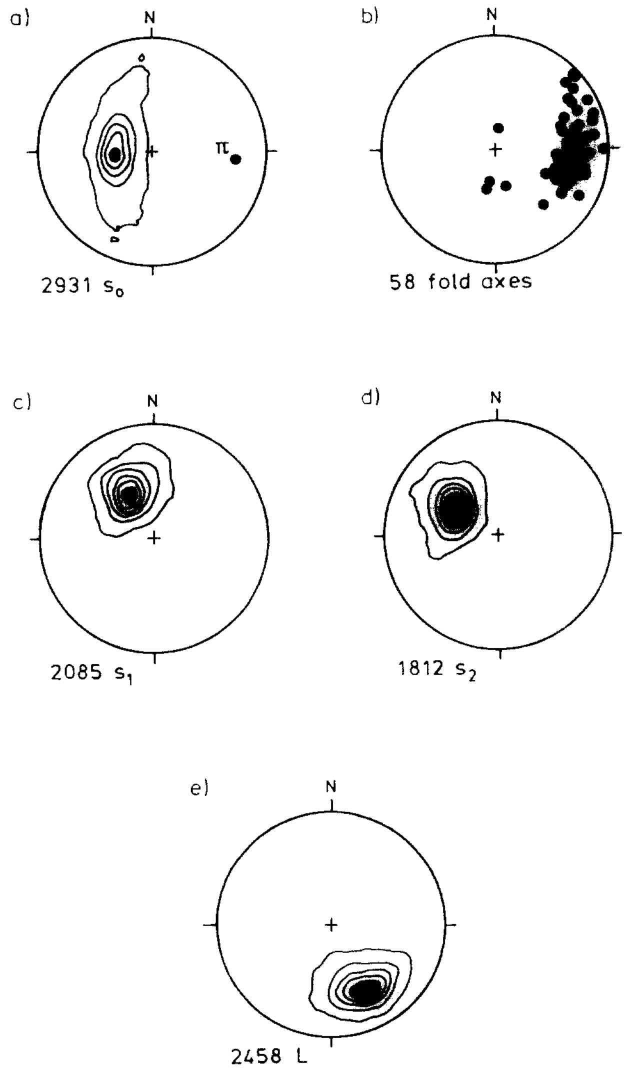

Fig. 3. Lower-hemisphere stereoplots of Lower Devonian rocks in the Monschau shear zone along the southeastern limb of the Stavelot–Venn anticlinorium: (a) poles to bedding (s_0), (b) fold axes, (c) poles to first cleavage (s_1), (d) poles to second cleavage (s_2), (e) stretching lineation (L). π is the average fold axis as derived from poles to bedding planes. Contours indicate 4% intervals starting with 1%.

Fig. 4). With increasing distance from the core of the anticlinorium, their trend changes progressively into the 'normal' NE–SW direction. Sinistral transpression in the transition zone would explain such a fold pattern. The trend of minor fold axes measured directly coincides with the average π fold axis, derived from a stereoplot of poles to the bedding surfaces (Fig. 3). Deviations from the pattern of minor fold axes are found in areas with an older Caledonian deformation, details of which are not entirely clear. Some N–S- and NW–SE-trending small-scale folds have previously been correlated with Caledonian deformation by Geukens (1975) and Fielitz (1987). Anomalous E–W-trending folds in the northeastern part of the area (between Hürtgen and Nideggen, Fig. 4) are possibly related to the adjacent axial depression of the Eifel North–South zone (Baum 1955). However, the exact relationship between the E–W-trending small-scale and the NE–SW-trending meso- to macro-scale folds in this area is not clear. The larger folds were partly deduced from scattered outcrops in a region with several cross-cutting Tertiary faults (Knapp 1980) and their trend could thus be incorrect. But, if it is correct, the meso- to macro-scale folds may post-date the small-scale shear zone and possible axe-ramp folds, because of the corresponding trend to the Stavelot–Venn anticlinorium and northwestward situated fold axes.

Besides minor folds the Monschau shear zone also contains numerous mullions (Brühl 1969) (see Fig. 8d), a special type of concentric folding typical for interlayered competent and incompetent beds with a given relationship of layer thicknesses and viscosity contrast (Smith 1977, Ramsay & Huber 1987 p. 391). The mullions are mostly parallel to the axes of small-scale folds. Nevertheless, in a few cases the mullions were refolded which indicates that they are older than some of the minor folds.

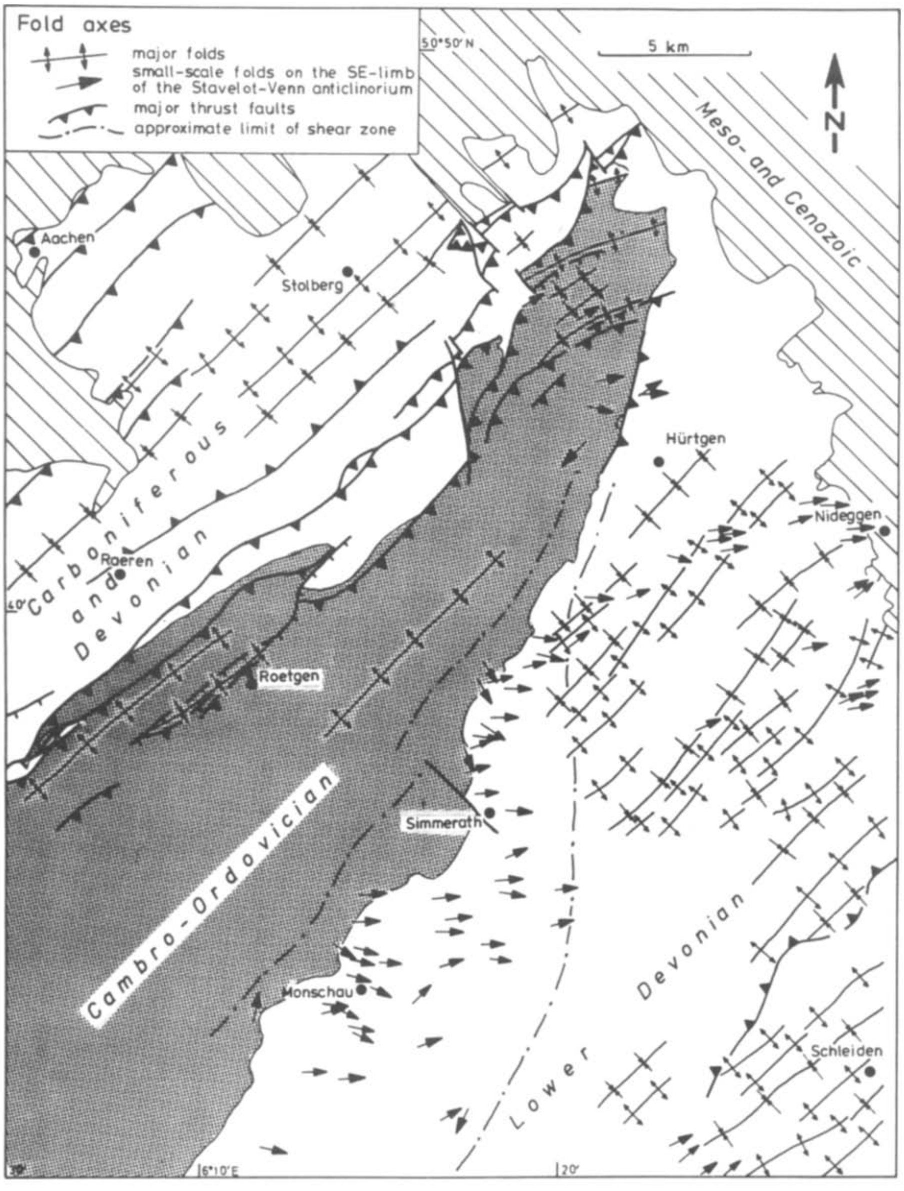

Fig. 4. Synoptic map with regional orientation of small-scale and major folds around the northeastern termination of the Stavelot–Venn anticlinorium (own measurements and data from Baum 1955, Knapp 1962, 1980, Spaeth 1979). The approximate limit of the Monschau shear zone is indicated. Major thrust and normal faults are added for orientation.

Cleavage

A closely spaced cleavage which is non-coaxial with the small- and medium-scale folds on the SE limb of the Stavelot–Venn anticlinorium (Fig. 8d) can be seen in all rocks along the SE limb of the anticlinorium (Figs. 5 and 8a & b). In the Monschau shear zone, however, there are two cleavages which are best developed in Gedinnian and Lower Siegenian silty shales (Fig. 8c). Since in the field age relationships between them are difficult to determine, oriented thin sections were inspected and yielded a crenulation cleavage (Fig. 9a). However, some thin sections suggest a contemporaneous development of the two cleavages. An interpretation as a *S–C*-type fabric (Lister & Snoke 1984, O'Brien *et al.* 1987) would then be possible which could be related to sinistral transpressive shearing. Scanning electron microscope studies also reveal growth of newly formed phyllosilicates within a matrix of older minerals (Figs. 9c & d) which implies upper anchizone to epizone metamorphic conditions for the cleavage formation (Weber 1981). Numerous field measurements show that both cleavages dip 30–40°, s_1 to the SSE and s_2 to the ESE (Fig. 3). However, the distribution of the pole maxima of s_1- and s_2-cleavages overlap and in coarser grained rocks often only one cleavage with such an overlapping distribution of the pole maxima is observed.

The two cleavages also occur in pre-Devonian Lower Paleozoic rocks, although the structural relationships are more complicated because of a possible even older cleavage which is commonly parallel to bedding (Spaeth *et al.* 1985, Fielitz 1987). Nevertheless, it appears that the younger cleavage s_2 does not penetrate very deeply from the shear zone into the Lower Paleozoic core of the

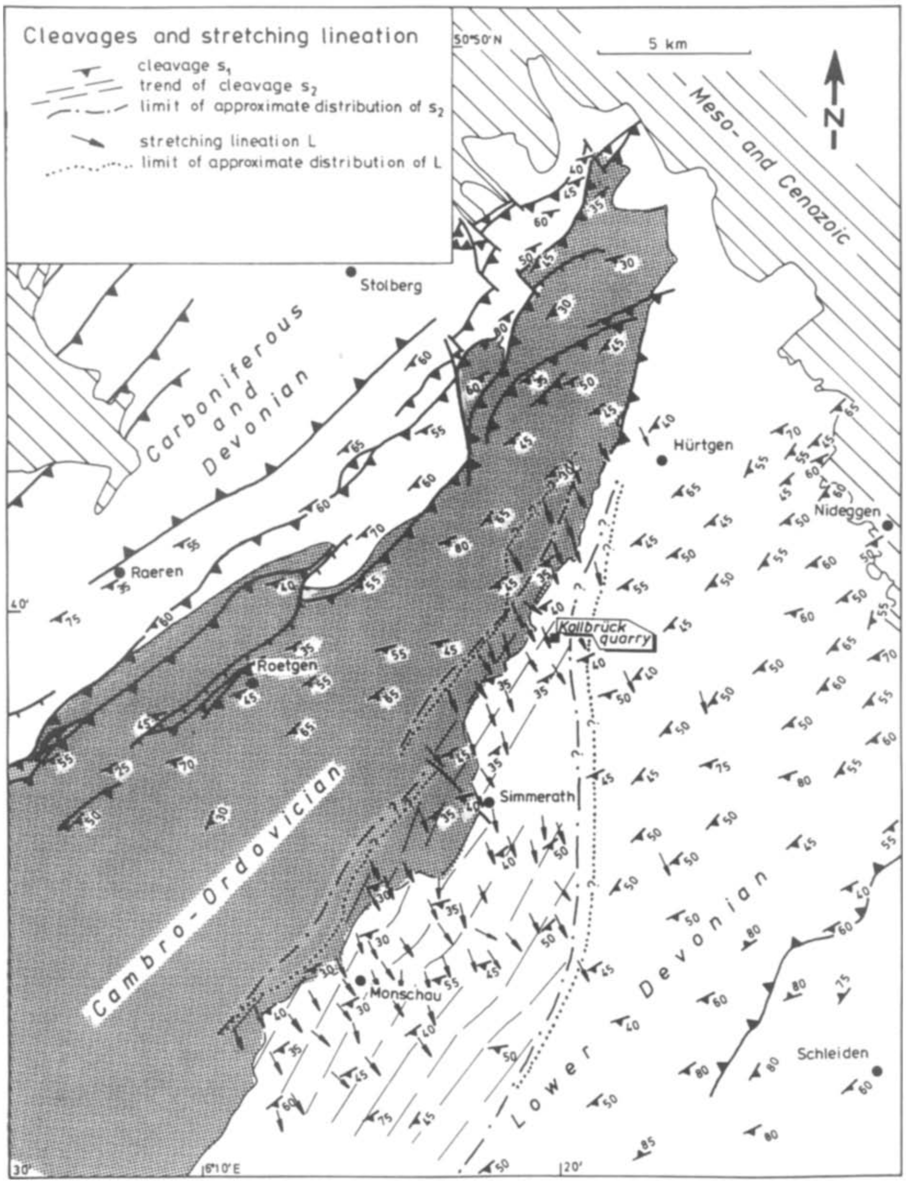

Fig. 5. Synoptic map with regional distribution of first cleavage s_1, trend of second cleavage s_2 and distribution of stretching lineation L around the northeastern termination of the Stavelot–Venn anticlinorium (own measurements and data from Baum 1955, Knapp 1962, Spaeth 1979, Frank 1987). The dash-dotted line indicates the limit of approximate distribution of s_2, the dotted line the limit of approximate distribution of L. Both delimit approximately the Monschau shear zone. The 'Kallbrück quarry' location refers to Fig. 10. Major thrust and normal faults are added for orientation.

Stavelot–Venn anticlinorium. Some shallow wells drilled in this outcrop-poor area helped to determine the northwestern limit of the s_2 distribution (Fig. 5). The southeast limit is not as well defined. Observations of Frank (1987) and the author suggest an approximate boundary that broadly coincides with the border of epizonal metamorphic conditions on the SE limb of the Stavelot–Venn anticlinorium. Macroscopically, this metamorphism is characterized by the growth of chloritoid- and pyrite-porphyroblasts. Their pre- to early synkinematic growth indicates that in the Monschau shear zone the peak of metamorphism preceded most of the cleavage formation (Fig. 9e). Chloritoid is restricted to a few rocks with an appropriate protolith chemistry and porphyroblasts reach only mm-size. Pyrite on the other hand is very common because of the widespread black-shale facies and porphyroblasts are up to 2 cm (mostly 0.5–1 cm) in diameter (Fig. 9e). Although normally not indicative of epizone metamorphism, the strong correlation between occurrence and size of the pyrite porphyroblasts and the extent of epizone metamorphism as defined by illite crystallinity leads to this conclusion. A Variscan age of the metamorphism is indicated because sedimentary rocks at least as young as Emsian and magmatic rocks of Givetian age were affected by it (Kramm & Buhl 1985, Kramm *et al.* 1985b, Frank 1987).

Stretching lineation

Within the Monschau shear zone the cleavage surfaces display a strong stretching lineation that plunges to the SSE (Figs. 3 and 8f). Parallel to this lineation pyrites are accompanied by well-developed pressure shadows filled by quartz fibres which can be used as kinematic indicators (Simpson & Schmid 1983, Passchier & Simpson 1986). Thin sections parallel to the stretching lineation show that most pressure shadows are asymmetric (Fig. 9e) indicating sinistral shear and tectonic transport of the hangingwall towards the NNW, oblique to the NE–SW-striking Stavelot–Venn anticlinorium. The oblique deformation can be related to sinistral transpression (Fig. 13). The distribution of the lineation is similar to that of the two cleavages (Fig. 5) and coincides with the area of epizonal metamorphic conditions (Fig 6). A block diagram (Fig. 13) illustrates the relationship between stretching lineation, pressure shadows and cleavages.

Quartz pressure shadows along pyrites (Fig. 9e) were also used to quantify strain with the method of Ramsay & Huber (1983, p. 265f). In the plane perpendicular to the cleavage and parallel to the stretching lineation (XZ) these quartz pressure shadows show sinistral shear with progressive rotation of either the quartz fibres or the suture lines between differently oriented face controlled fibre groups (Fig. 9e). A rigid fibre model was applied because a few attempts with the deformable fibre model gave only insignificantly different results and in most cases microscopic signs of fibre deformation appeared to be of little importance. Measurements yielded minimum values of stretching parallel to the lineation of 70–222% (mean of 123%) in slates and silty slates and of 17–191% (mean of 104%) in siltstones and sandstones. A regional plot of the strain ellipses in the XZ-plane indicates higher strain values towards the southeast, away from the immediate contact with the pre-Devonian Lower Paleozoic rocks (Fig. 6). If plane strain is assumed which seems reasonable because quartz pressure shadows are absent in the Y-direction (i.e. perpendicular to the cleavage planes and to the stretching lineation), minimum shortening between 50 and 60% can be deduced for the shear zone rocks.

Previous strain measurements on deformed quartz grains in Lower Devonian rocks from the Monschau shear zone yielded minimum values of stretching parallel to the lineation of 18–35% (mean of 26%) for greywackes (Spaeth 1969), 41–94% (mean of 71%) for pelitic rocks and 30–48% (mean of 36%) for psammitic rocks (Mukhopadhyay 1973). The difference betweeen these values and the quartz pressure shadow values are possibly due to strain partitioning which could have involved intense deformation of the rock matrix, sliding between quartz grains and removal of quartz by pressure solution from domains with apparent lower strain rates (Fig. 9b). If rigid objects such as pyrites were absent, only part of the strain would be registered by the quartz grains.

Mica (002) fabric diagrams for slate samples from the investigated area were derived by X-ray texture goniometry. With this method (e.g. Spaeth 1971, Lipshie *et al.* 1976, Weber 1981, Oertel 1983, O'Brien *et al.* 1987) microfabrics of phyllosilicate-rich rocks can be related to macroscopic cleavage planes. Taken together with a few samples from a local study by Spaeth (1971), the author's work indicated that a half-girdle distribution perpendicular to the lineation predominates (Fig. 7). This means that the crystallographic (002) planes of the phyllosilicates are in general parallel to the lineation, a result that is supported by scanning electron microscope investigations on sections perpendicular and parallel to the lineation (YZ- and XZ-sections, Figs. 9c & d). The regional variation of the pole diagrams shows a concentration of half-girdle fabrics in a narrow band near the NW border of the Monschau shear zone with a trend toward circular fabrics towards the SE and the NW (Fig. 7). This band with half-girdle fabrics is narrower than the distribution of the macroscopic stretching lineation (Fig. 5) and corresponds to the main zone of shearing. The circular fabrics of the pole diagrams of pre-Devonian Lower Paleozoic rocks which commonly show a high degree of preferred orientation intensities could be due to the superposition of a finely laminated bedding texture and a bedding-parallel slaty cleavage.

Shear bands

Most of the deformation in the Monschau shear zone probably occurred during retrograde metamorphic conditions. This is well illustrated by structures in the Kallbrück quarry (Fig. 10, location in Fig. 5). The

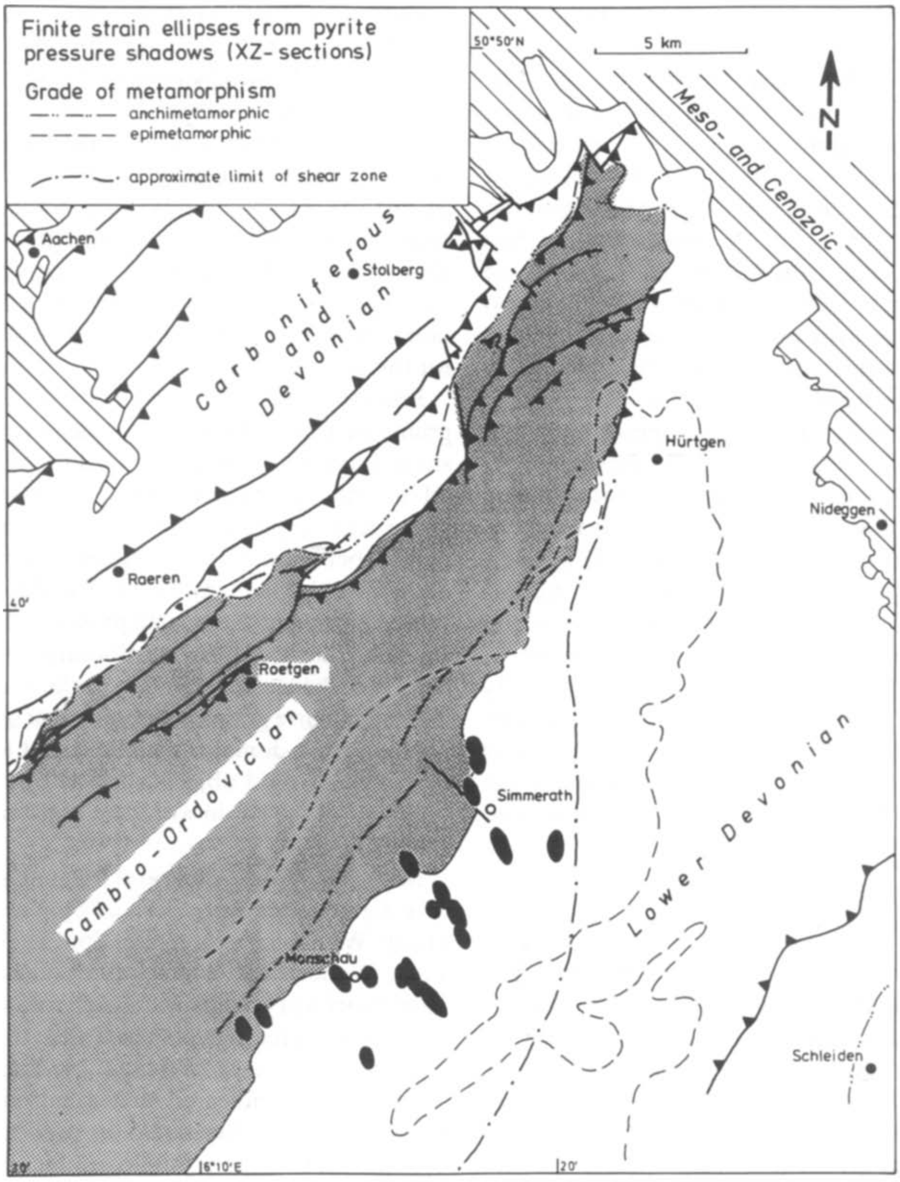

Fig. 6. Finite strain ellipses from pyrite pressure shadows (*XZ*-sections) in the Monschau shear zone. An increase in deformation intensity towards the southeast is clearly recognizable. The regional grade of metamorphism around the northeastern termination of the Stavelot–Venn anticlinorium is indicated by the dash-dotted line (transition from diagenesis to anchizone metamorphism) and the stippled line (transition from anchizone to epizone metamorphism). Both were determined by illite crystallinity and taken from Frank (1987). The approximate limit of the Monschau shear zone is indicated. Major thrust and normal faults are added for orientation.

quarry exhibits a N-verging anticline of Lower Devonian greywackes and silty shales. The sedimentary rocks were affected by a penetrative cleavage and later cross-cut by numerous shear bands. Although the shear bands seem to be contractional structures, some extensional faults exist as well. The most prominent zone of thrust-related shear bands is several metres thick. Locally, even coarse-grained greywackes were converted into phyllonites along the shear bands. The transition from penetrative cleavage to localized shear bands indicates strain concentration during retrograde conditions within a transition zone between ductile and brittle deformation. The presence of brittle conditions is supported by the fact that some of the contractional shear bands are clearly associated with synkinematic quartz veins which are subparallel to each other and are oriented at a moderate angle to the shear surfaces (Fig. 10, lower right). Their en échelon arrangement generally supports a contractional character of the shear bands.

Kink bands

The last major deformational structures in the Monschau shear zone are kink bands. In general they are less than 1 cm to several cm wide and occur either as single structures or are concentrated in cm- to dm-wide intervals. Although conjugate kink bands exist, mostly one set predominates. Their short limbs and moderate to steep NNE-plunge indicates hangingwall movement to the SSW which might be interpreted as related to a late backthrusting event.

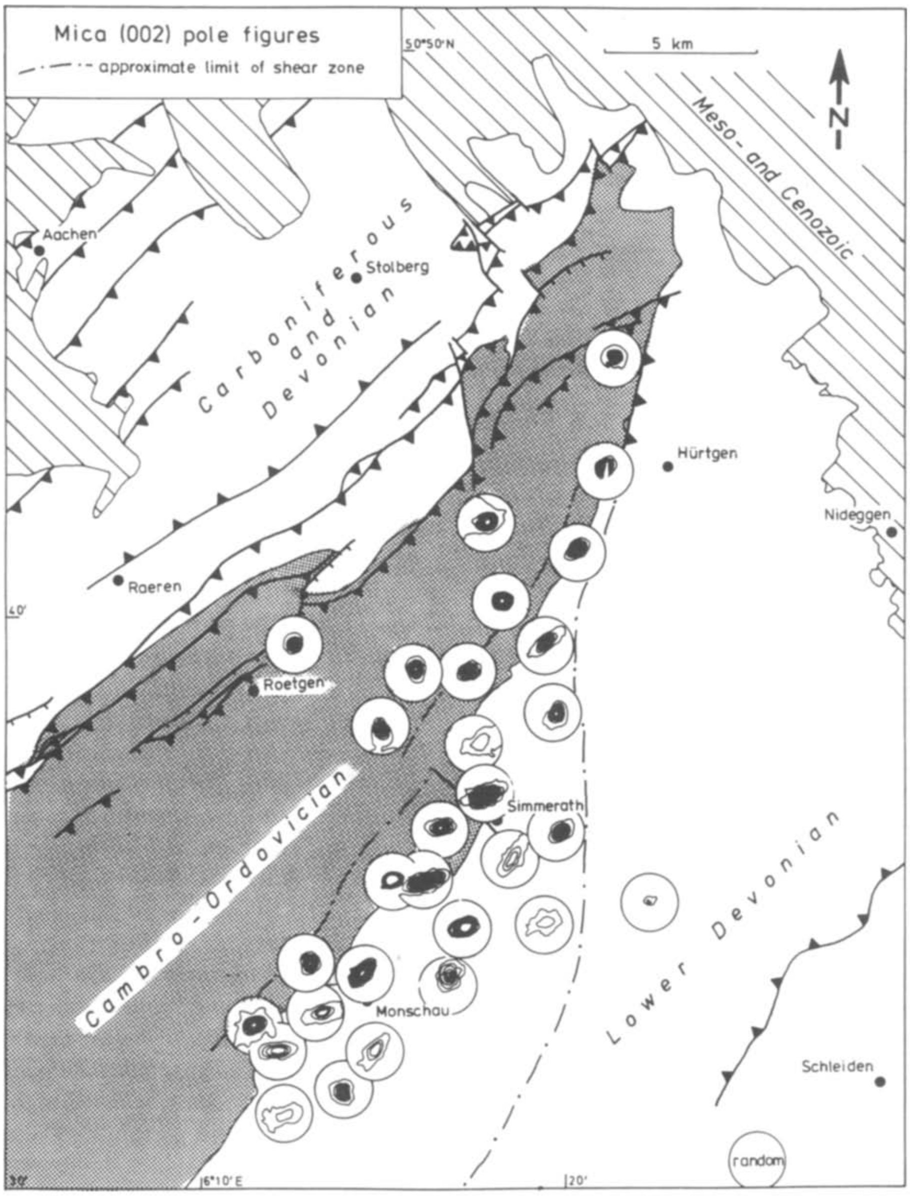

Fig. 7. Synoptic map with regional distribution of mica (002) pole figures from X-ray texture goniometry around the northeastern termination of the Stavelot–Venn anticlinorium (with additional data from Spaeth 1971). All are lower-hemisphere equal-area projections. The short axes of the half-girdles are parallel to the stretching lineation. The half-girdle distribution of the micas is characteristic of the most intensely deformed basal part of the Monschau shear zone. The approximate limit of the Monschau shear zone is indicated. Major thrust and normal faults are added for orientation.

Quartz veins

Quartz veins are numerous in the Monschau shear zone (see also Thome 1955, Hoffmann 1961, Breddin & Hellermann 1962). They are particularly common in sandstone beds and are either sigmoidal or boudinaged. The sigmoidal quartz veins are commonly restricted to single or a sequence of a few sandstone beds and indicate with their sense of bending and/or a more or less important displacement at the bed boundaries flexural slip in fold structures. A few quartz veins occur in slates where they are commonly parallel or subparallel to the cleavage. These often strongly boudinaged quartz veins (Fig. 8e) imply a formation before or at an early stage of cleavage development. Because of extensive transposition their dip is generally to the SE. Late quartz veins are much more restricted in occurrence and number. Some veins correlate with extension parallel to stretching or with shearing associated to late thrusts (Fig. 10), others seem to be associated with dilatation during kink folding. A few NW-trending cross joints contain thin fillings of undeformed quartz.

In summary, minor structures in the Monschau shear zone indicate that it developed shortly after regional folding in an area characterized by epizonal metamorphism. Kinematic indicators in the Monschau shear zone demonstrate oblique sinistral convergence along the southern limb of the Stavelot–Venn anticlinorium. The coincidence of the area of two cleavages, E-plunging minor folds, S-plunging stretching lineations and a higher grade of metamorphism, whose maximum was pre-kinematic to the penetrative deformation, suggests an emplacement under retrograde metamorphic conditions of the SE hangingwall. Towards the northeast contraction was increasingly accommodated by distributed thrust faulting. Coincidence of the anchizone–epizone metamorphic boundary with the change from penetrative shearing to distributed thrusting (Figs. 6 and 11) suggests a ductile–brittle transition. The occurrence of several generations of quartz veins suggests intense fluid movements early in the protracted phase of shortening.

REGIONAL TECTONICS AND REFLECTION SEISMIC EXPRESSION OF THE MONSCHAU SHEAR ZONE

Dramatic differences in the thickness of sedimentary rocks between the two limbs of the Stavelot–Venn anticlinorium (Figs. 2 and 12, profiles A–A′ and B–B′) suggest that the location of the Monschau shear zone was probably controlled by a major synsedimentary normal fault zone of Siegenian to Emsian age. Synsedimentary extensional faulting of Devonian age has been suggested also for other areas of the central Rhenohercynian belt (Oncken 1984, Meilliez 1989, Werner 1989).

The earliest compressional structures in the Monschau shear zone are mullions of sandstone beds. They might indicate the onset of shortening and folding and probably precede the peak of metamorphism represented by the emplacement of the majority of quartz veins and the growth of porphyroblasts of pyrite and chloritoid. Pyrites locally grew across the borders of quartz veins which in turn separate mullions from each other. Pressure shadows along pyrites show that the peak of metamorphism preceded penetrative cleavage formation on the SE limb of the Stavelot–Venn anticlinorium. Local non-coaxiality of cleavage and small-scale folds (Fig. 8d) suggests that some of the regional folds developed earlier than the regional cleavage. In general, folds transected by cleavage have been ascribed to protracted single transpressive deformations by distributed shear (Borradaile 1978, Sanderson *et al.* 1980, Gray 1981, Treagus & Treagus 1981, Soper 1986, Woodcock 1990). While cleavage is considered to have developed early in some cases (Gray 1981, Soper 1986) or synchronous with folding in others (Sanderson *et al.* 1980), most models favour a delayed initiation of cleavage development due to high fluid pressures early during folding (Borradaile 1978, Soper 1986, Woodcock 1990). Delayed cleavage formation is supported by data from the Monschau shear zone. There transpressive sinistral shear occurred in two stages: early folding with fold axes at a relatively high angle to the strike of the shear zone and a later superposition of a regionally uniformly oriented cleavage (s_1). Continuing transpressive deformation induced the formation of the second cleavage s_2.

The penetrative stretching lineation which formed contemporaneously with cleavages and asymmetric pressure shadows indicates tectonic transport of the hangingwall towards the NNW and oblique to the NE–SW-striking Stavelot–Venn anticlinorium (Fig. 13). The same direction of contraction can be inferred for an area of discrete thrusting and folding on the northern limb of the Stavelot–Venn anticlinorium characterized by ENE–WSW trends (Fig. 4, NNW of Hürtgen, and Fig. 13). The lateral borders of the thrusts seem to be nearly N–S-trending tear faults which cause a northward deflection of the northeasternmost Stavelot–Venn anticlinorium. Two generations of small-scale folds in Upper Carboniferous rocks north of the Aachen-Midi thrust also have been interpreted as being related to this late phase of thrusting (Wrede 1987). A late clockwise rotation of the direction of regional SE–NW convergence into S–N contraction during progressive forelandward propagation would be in accordance with observations from other parts of the central Rhenohercynian belt (Oncken 1988a).

A tectonometamorphic scenario for the Monschau shear zone can thus be proposed. The area of the shear zone as defined by the distriubtion of s_2, the stretching lineation, and the half-girdle of the X-ray pole figures (Figs. 11 and 13) probably underwent prograde metamorphism near a synsedimentary normal fault as indicated by the pre-kinematic chloritoid and pyrite porphyroblasts. The subsequent transpressive shearing, because post-kinematic to the porphyroblasts, was probably already under beginning retrograde metamorphic conditions. It followed mainly a layer of several hundred metres of Gedinnian and Lower Siegenian shales near the Lower Paleozoic–Lower Devonian unconformity (Fig. 12). To the northeast the shear zone passes into discrete thrusts (Fig. 11). Since the domain between the shear zone and thrust coincides with the passage from epizone to anchizone metamorphic conditions (Fig. 6), the ductile–brittle transition here probably corresponds to a temperature of approximately 330°C (Frank 1987). Shearing, once concentrated within the Monschau shear zone, continued and terminated along discrete shear bands and kink bands which document retrograde metamorphic conditions.

The suggested strike-slip component does not necessarily imply a significant sinistral displacement along the Monschau shear zone. As pointed out by Simpson (1990), strong fabric modifications do not necessarily mean large displacements. A more or less strike-parallel orientation of the shear zone and the absence of markers do not allow a quantitative estimate of the displacement. It is probably not very large, particularly in the central segment where the deformation increments are distributed over a broad area. As shown by the widespread cleavage s_1, the main shortening was oriented SE–NW, but was then reoriented into a NNW direction.

The probable prolongation of the Monschau shear zone to the southwest is assumed to be in an area covered by Permian sediments in a graben-like structure situated between two major anticlines of the Stavelot–Venn anticlinorium (Fig. 2). North of this graben the Xhoris Thrust might be a branch of the Monschau shear zone. Locally observed stretching lineations northeast of the Permian graben support this interpretation. Recumbent folding associated with a horizontal cleavage was described at the bottom of the Grand Halleux

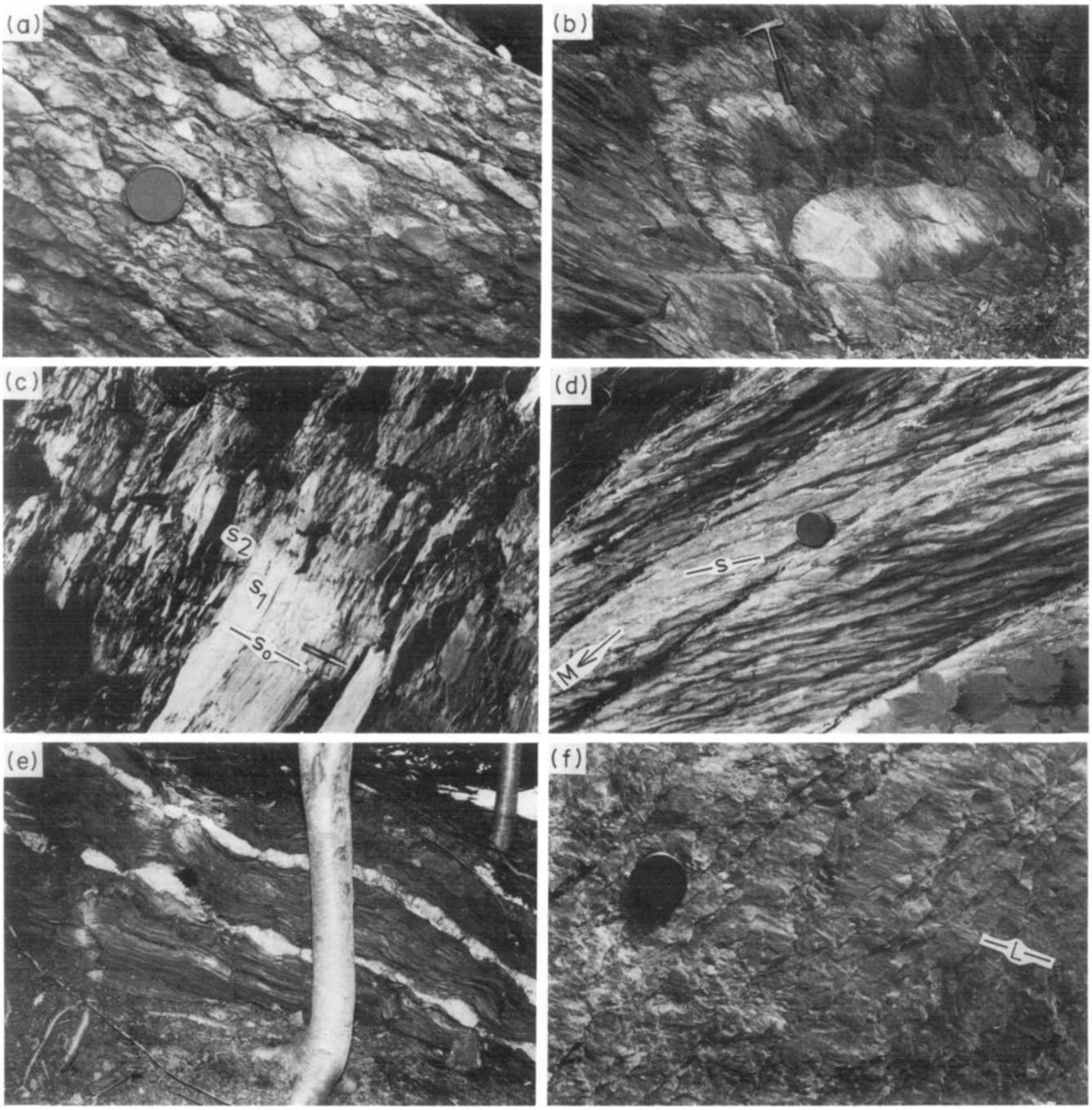

Fig. 8. (a) Gedinnian basal conglomerate from the southeastern limb of the Stavelot–Venn anticlinorium affected by an intensive cleavage. (b) Gedinnian arkoses and shales from the southeastern limb of the Stavelot–Venn anticlinorium with a N-verging (left) inclined small-scale fold affected by an intensive cleavage. (c) Siegenian silty shales from the southeastern limb of the Stavelot–Venn anticlinorium affected by two intensive cleavages (s_1 and s_2, their intersecting edges plunging steeply to the left). Bedding (s_0) dips moderately to the lower right (ENE). (d) Bottom surface of a nearly upright Siegenian sandstone bed with mullion structures from the southeastern limb of the Stavelot–Venn anticlinorium. The axes of the mullions (M) are parallel to the fold axis and dip to the east. A non-coaxial cleavage (s) intersects the mullion (and fold) axes. (e) Siegenian shales from the southeastern limb of the Stavelot–Venn anticlinorium with boudinaged quartz veins. The quartz veins are oriented nearly parallel to an intensive cleavage which indicates transposition. (f) Gedinnian shales from the southeastern limb of the Stavelot–Venn anticlinorium affected by an intensive stretching lineation (L).

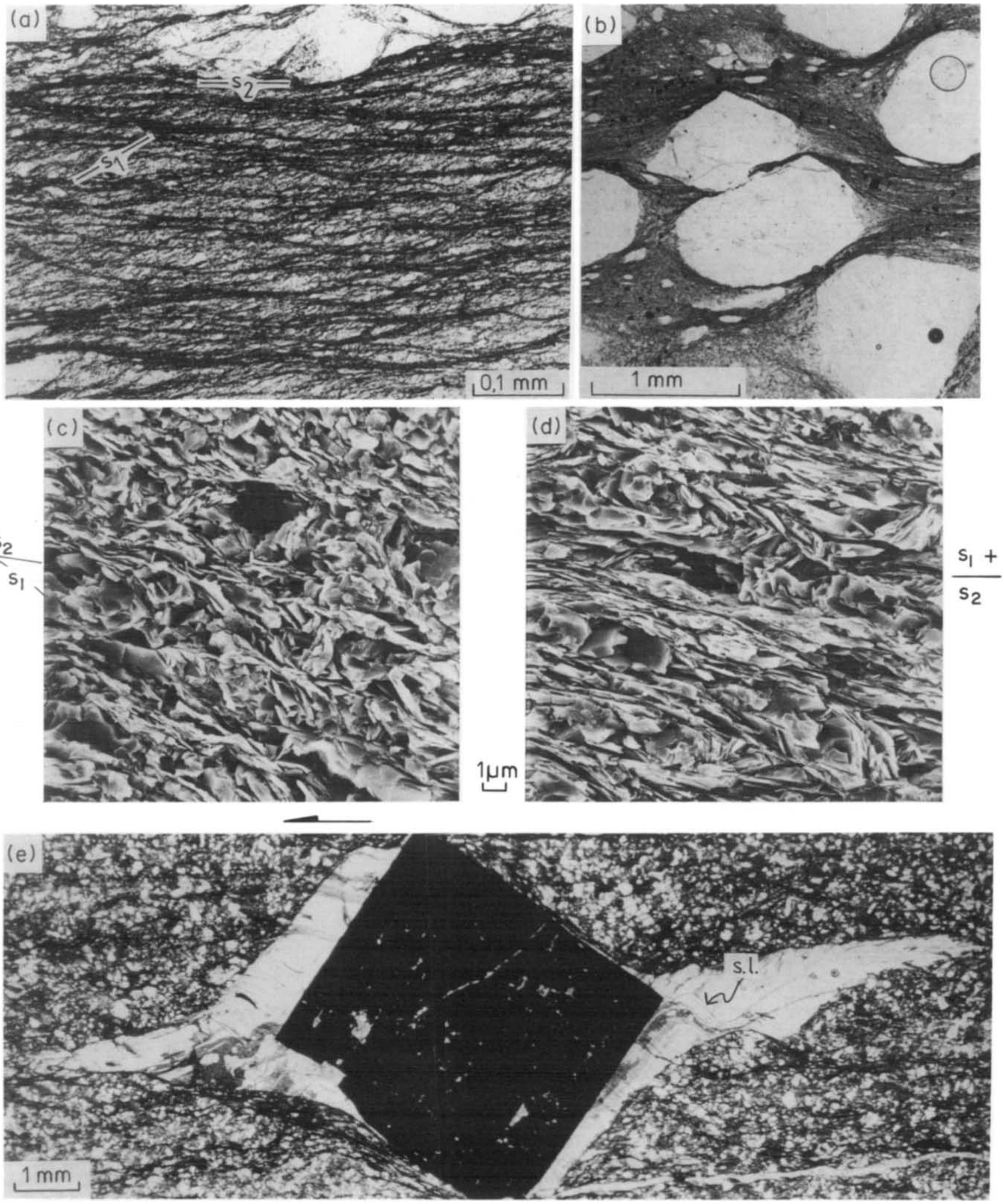

Fig. 9. (a) Siegenian silty shale from the southeastern limb of the Stavelot–Venn anticlinorium affected by two intensive cleavages (s_1 and s_2). Plane polarized light, YZ-section. (b) Gedinnian sandy–silty turbidite from the southeastern limb of the Stavelot–Venn anticlinorium with intensive cleavage in the finer grained matrix and pressure solution of quartz grains. Small quartz grains in the matrix are stretched while larger grains are not. Plane polarized light, YZ-section (c) & (d) Siegenian silty shale from the southeastern limb of the Stavelot–Venn anticlinorium (same sample as in a, left YZ- and right XZ-section). The preferred phyllosilicate orientation parallel to the stretching lineation (XZ-section) is clearly recognizable. It is associated with new formation of phyllosilicates. Scanning electron micrographs. (e) Siegenian sandstone from the southeastern limb of the Stavelot–Venn anticlinorium containing an idiomorphic pyrite with asymmetric, rotated quartz pressure shadows which indicate sinistral shear. s.l. = suture line between differently oriented face controlled fibre groups. Plane polarized light, XZ-section parallel to the stretching lineation.

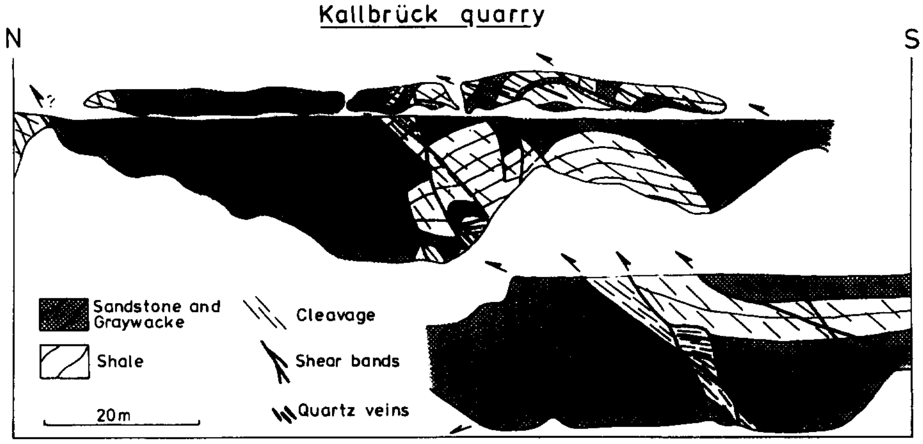

Fig. 10. Lower Siegenian greywackes and shales from the southeastern limb of the Stavelot–Venn anticlinorium with an intensively sheared N-verging anticline. Some of the shear bands are accompanied by extensional quartz veins (e.g. to the lower right) indicating brittle conditions during their formation. Although most shear bands and associated quartz veins are contractional, some extensional faults exist as well. Kallbrück quarry with three levels in their respective correct position (location in Fig. 5).

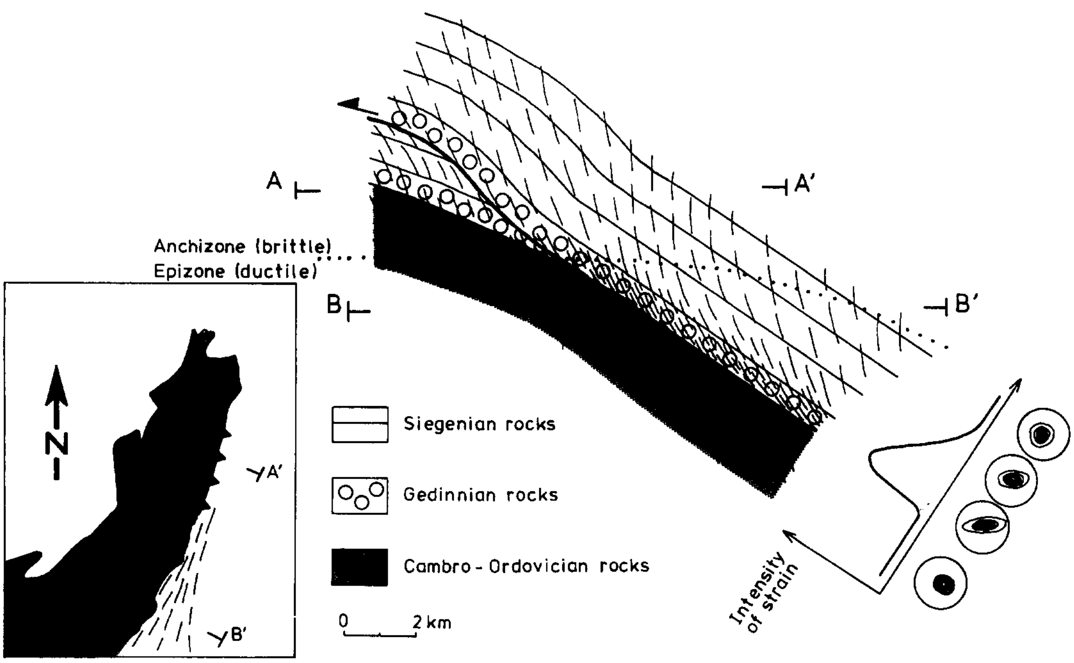

Fig. 11. Schematic profile through the Monschau shear zone at its northeastern end showing the transition from ductile to brittle deformation. The broad band of penetratively sheared rocks (profile B–B′) grades northeastward parallel to its strike into a narrow discrete thrust plane (profile A–A′) at the limit between epizone to anchizone metamorphic conditions. The maximum intensity of strain is located at the base of the shear zone and diminishes rapidly within the basal Cambro-Ordovician rocks and slowly within the overlying Lower Devonian rocks. The fabric evolution as indicated by typical mica (002) pole figures from X-ray texture goniometry follows closely this trend (see Fig. 7).

deep well (Breddin 1973) and has been projected into profile C–C′ of Fig. 12. It can be interpreted as connected with the continuation or a branch of the Xhoris thrust zone, thus supporting the interpretation of the Monschau shear zone as a brittle–ductile transition.

Three interpretative cross-sections through the Stavelot–Venn anticlinorium demonstrate the geologic situation (Fig. 12). Sections A and B are bed-length balanced in their northwestern part, where ductile deformation is negligible and marker beds are well developed. Pre-orogenic synsedimentary normal faults, whose exact configuration is unknown, considerable ductile deformation and/or missing reliable continuous marker beds leave the southeastern part of the sections as well as section C unbalanced. However, section B is also constrained by a recent deep crustal reflection seismic profile (DEKORP Research Group 1991) (Fig. 14, for location see line B–B′ in Fig. 2). The Stavelot–Venn anticlinorium thus would be a ramp fold whose apparent hinge line is displaced to the northwest. This ramp fold is complicated by a rather complex duplex involving in-sequence and out-of-sequence thrusts (sections A and B). Their presence is required by a combination of NE down-plunge projection of the surface

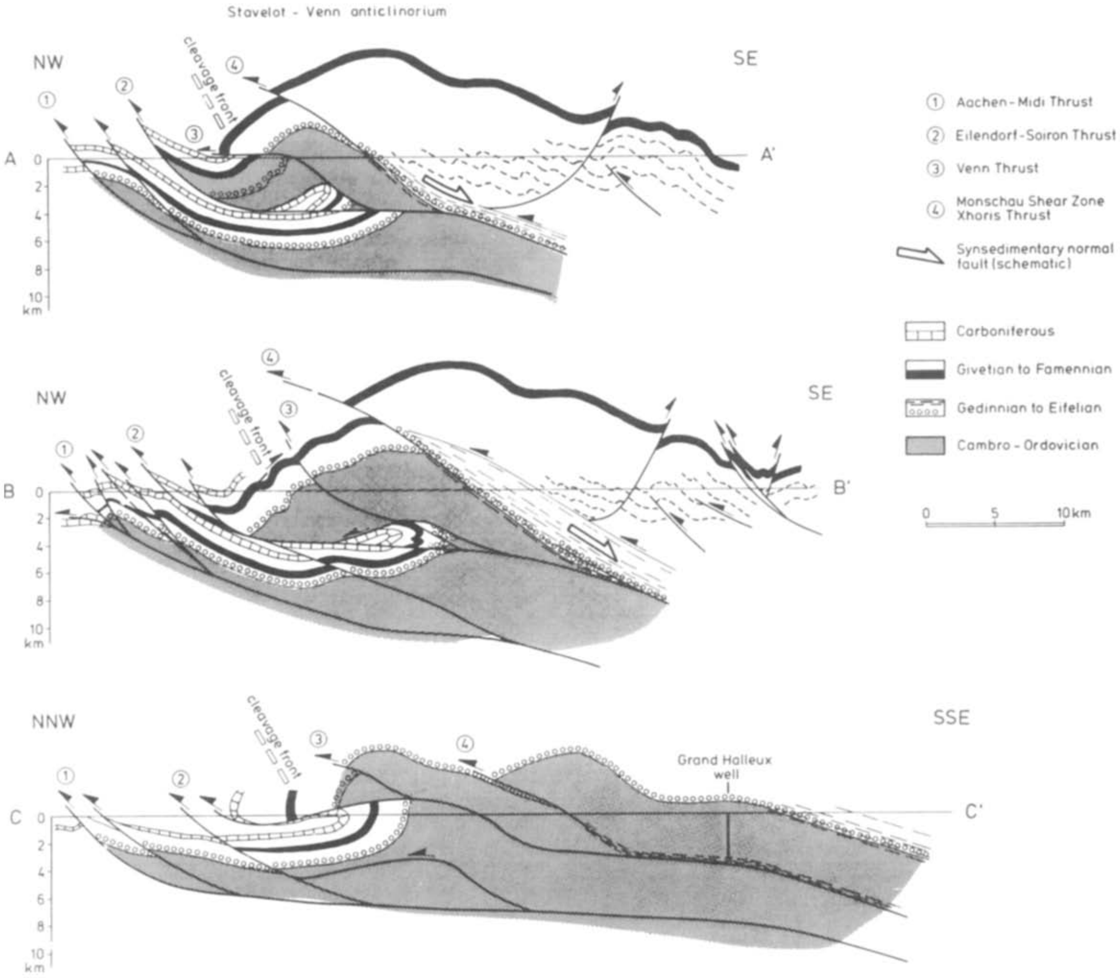

Fig. 12. Three partly balanced cross-sections through the Stavelot–Venn anticlinorium (location shown in Fig. 2). For reason of simplification minor Mesozoic to Tertiary deposits as well as the small Permian graben in profile C are not represented. The profiles are characterized by a ramp–flat geometry and a rather complex duplex in sections A and B, originating from an inverted Lower Devonian synsedimentary normal fault. The latter corresponds to the location of the Monschau shear zone.

geology from section C (area of Theux Window, for location see also Fig. 2), seismic reflectors (Fig. 14) and bed-length balancing.

The Monschau shear zone follows mainly the Early Devonian synsedimentary normal fault zone which was upthrusted together with subjacent rocks of the Stavelot–Venn back limb. This would bring the highest grade metamorphic rocks of the structure to the surface along the oblique out-of-sequence Monschau structure. The DEKORP seismic profile displays in this sector a broad set of relatively shallow dipping reflectors which in projection to the surface coincide with the shear zone (Fig. 14). Since to the surface the shear zone is characterized by a combination of subparallel bedding, cleavage and shear planes, it is thought, that the latter is also responsible for the numerous reflectors observed in the seismic profile. At depth the shear zone joins a strong reflector which to the northwest forms a ramp and then continues nearly horizontal for at least about 15 km. This strong reflector is commonly attributed to the Aachen–Midi Thrust (Meissner *et al.* 1981, Durst 1985, Betz *et al.* 1988). However, a geologic interpretation of the seismic profile suggests that it correlates with the Eilendorf–Soiron Thrust (Fig. 12, profile B–B′) located farther southeast.

In contrast to many discussed zones of dextral transcurrence and transpression along the Variscan belt (e.g. Arthaud & Matte 1977, Berthé *et al.* 1979, Holder & Leveridge 1986, Matte 1986, Ziegler 1986, Krohe & Eisbacher 1988) which are mostly late-orogenic, the Monschau shear zone is mainly an early- to synorogenic structure. The origin of the nappe geometry in this part of the Rhenohercynian belt as well as the NNE deflection of the described late tectonic transport direction might be sought in the presence of the nearby stable cratonic block of the Brabant Massif (Fig. 1) which was consolidated in pre-Variscan times and seems to have been relatively unaffected by later penetrative deformations (Michot 1980). The Variscan deformation was blocked by the Massif to the west, but could advance northward further to the east. The southeast termination of the Brabant Massif is the approximate boundary between these two domains, separated by a transfer zone to which belongs also the described shear zone.

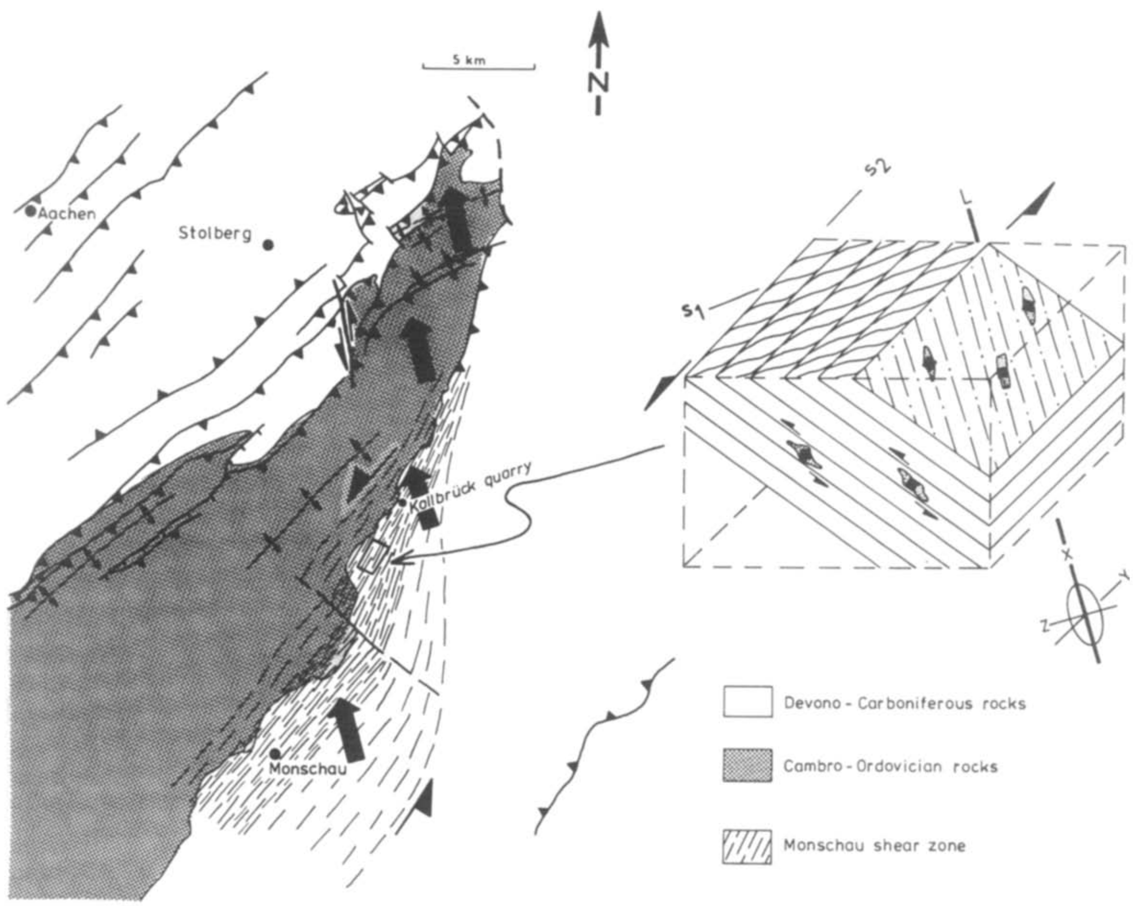

Fig. 13. Limit and regional context of the Monschau shear zone on the southeastern limb of the Stavelot–Venn anticlinorium. The strongest deformation is concentrated in a narrow band inside the larger shear zone. The inset block diagram shows the combined microstructural observations (relationship between the two cleavages s_1 and s_2, the stretching lineation L, and the sense of shearing as indicated by quartz pressure shadows on pyrites) which led to the interpretation of the Monschau shear zone as a sinistral transpressive structure whose hangingwall displacement was towards the NNW. Note also the ENE–WSW-trending folds and thrust faults bordered laterally to the west by an assumed nearly N–S-trending sinistral tear fault. The offset of the shear zone by a younger NW–SE-trending normal fault is probably of Tertiary age.

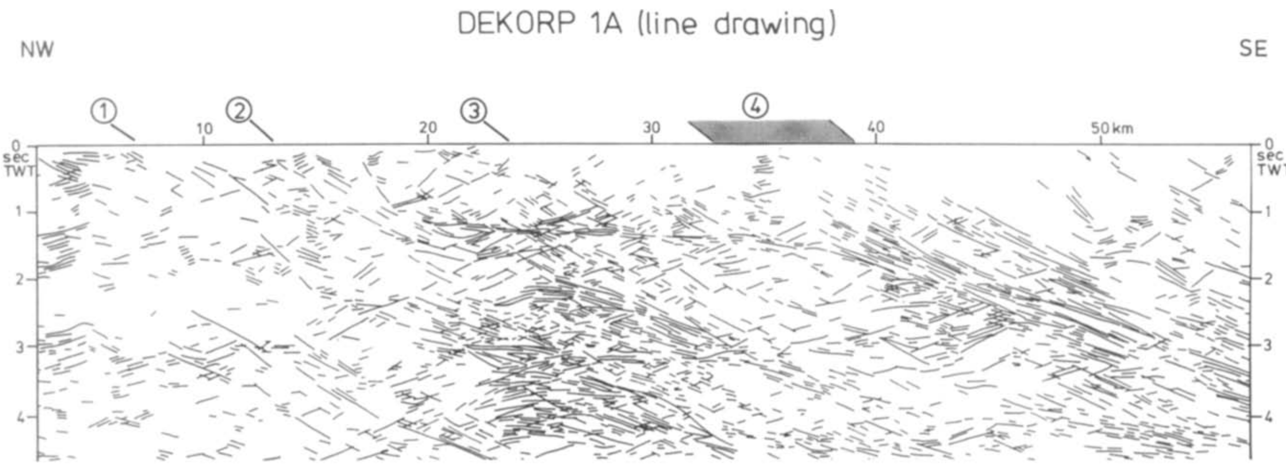

Fig. 14. Line drawing of a section from the DEKORP 1A migrated reflection seismic profile crossing the Stavelot–Venn anticlinorium (DEKORP Research group 1991). For the profile mainly finite-difference migration was used, in some critical regions with complex structures also frequency–wavenumber migration. Its geologic interpretation is seen in Fig. 12 (profile B–B′, same numeration of thrusts). The shear zone (zone 4) appears clearly in the seismic profile as a marked broad set of relatively shallow dipping reflectors.

CONCLUSIONS

In the northwestern central Rhenohercynian slate belt the 3–6 km wide Monschau shear zone situated on the southeastern limb of the major Stavelot–Venn anticlinorium is interpreted as the result of inversion along an Early Devonian synsedimentary normal fault system. After a phase of prograde metamorphism considerable penetrative deformation affected the predominantly fine-grained clastic rocks. Shortening was achieved first by folding accompanied by sediment dewatering, phyllosilicate rotation, pressure solution and mineral growth (=cleavage formation), and finally by retrograde oblique thrust faulting. Internal shortening within the shear zone was in the order of 50–60%. Sinistral oblique thrusting along the Monschau shear zone occurred within a brittle–ductile transition zone at about 330°C on the backlimb of an allochthonous anticlinorium located above a pre-orogenic basement ramp. Its continuation shows up clearly in a seismic reflection profile to a depth of 12–15 km.

Acknowledgements—Much of the field work was carried out during the site selection phase for the deep continental drilling project of Germany and was partly funded by the Deutsche Forschungsgemeinschaft (DFG). The author wishes to thank the DFG for their financial support. The reflection seismic data have been made available by the DEKORP research group. The paper greatly profited from discussions with B. Frank, H. Siemes, G. Spaeth, R. Walter (all at Rheinisch-Westfälische Technische Hochschule Aachen, Germany), and O. Goffette, J.-L. Mansy, F. Meilliez (all of the Université de Lille Flandres-Artois, France). G. H. Eisbacher (Universität Fridericiana Karlsruhe, Germany) critically read and improved several versions of the paper which is greatly appreciated. R. O. Greiling and an unknown reviewer are thanked for their helpful comments.

REFERENCES

Ahrendt, H., Clauer, N., Hunziker, J. C. & Weber, K. 1983. Migration of folding and metamorphism in the Rheinische Schiefergebirge deduced from K–Ar and Rb–Sr Age Determinations. In: *Intracontinental Fold Belts* (edited by Martin, H. & Eder, F. W.). Springer, Berlin, 323–338.

Albrecht, K. 1971. Quantitativ-geometrische Untersuchungen des Faltenbaues im Massiv von Stavelot–Venn bei Eupen (Belgien). *Geol. Mitt.* **11**, 1–137.

André, L., Hertogen, J. & Deutsch, S. 1986. Ordovician–Silurian magmatic provinces in Belgium and the Caledonian orogeny in middle Europe. *Geology* **14**, 879–882.

Arthaud, F. & Matte, P. 1977. Late Paleozoic strike-slip faulting in southern Europe and northern Africa: Result of a right-lateral shear zone between the Appalachians and the Urals. *Bull. geol. Soc. Am.* **88**, 1305–1320.

Baum, F. 1955. Tektonische Untersuchungen am Westrand der Eifeler Nord–Süd-Zone zwischen Untermaubach und Heimbach. *Geol. Rdsch.* **44**, 148–167.

Berthé, D., Choukroune, P. & Jegouzo, P. 1979. Orthogneiss, mylonite and non-coaxial deformation of granites: the example of the South Armorican Shear Zone. *J. Struct. Geol.* **1**, 31–42.

Betz, D., Durst, H. & Gundlach, T. 1988. Deep structural seismic reflection investigations across the northeastern Stavelot–Venn Massif. *Annls Soc. géol. Belg.* **111**, 217–228.

Bless, M. J. M., Bouckaert, J. & Paproth, E. 1980. Pre-Permian around the Brabant Massif in Belgium, the Netherlands and Germany. *Meded. Rijks geol. Dienst* **32-1**, 1–179.

Borradaile, G. J. 1978. Transected folds: A study illustrated with examples from Canada and Scotland. *Bull. geol. Soc. Am.* **89**, 481–493.

Breddin, H. 1973. Tiefentektonik und Deckenbau im Massiv von Stavelot–Venn (Ardennen und Rheinisches Schiefergebirge). *Geol. Mitt.* **12**, 81–130.

Breddin, H. & Hellermann, E. 1962. Petrogene Mineralgänge im Paläozoikum der Nordeifel und ihre Beziehung zur inneren Deformation der Gesteine. *Geol. Mitt.* **2**, 197–224.

Brühl, H. K. 1969. Boudinage in den Ardennen und in der Nordeifel als Ergebnis der inneren Deformation. *Geol. Mitt.* **8**, 263–308.

Butler, R. W. H. 1989. The influence of pre-existing basin structure on thrust system evolution in the Western Alps. In: *Inversion Tectonics* (edited by Cooper, M. A. & Williams, G. D.). *Spec. Publs geol. Soc. Lond.* **44**, 105–122.

Coney, P. J., Jones, D. L. & Monger, J. W. H. 1980. Cordilleran suspect terranes. *Nature* **288**, 329–333.

Cooper, M. A., Williams, G. D., de Graciansky, P. C., Murphy, R. W., Needham, T., de Paor, D., Stonely, R., Todd, S. P., Turner, J. P. & Ziegler, P. A. 1989. Inversion tectonics—a discussion. In: *Inversion Tectonics* (edited by Cooper, M. A. & Williams, G. D.). *Spec. Publs geol. Soc. Lond.* **44**, 335–347.

DEKORP Research group 1991. Results of the DEKORP 1 (BELCORP–DEKORP) deep seismic reflection studies in the western part of the Rhenish Massif. *Geophys. J. Int.* **106**, 203–227.

Delveaux de Fenffe, D. & Laduron, D. 1984. Analyse structurale au bord sud du Massif de Rocroi (Ardennes Francaises). *Bull. Soc. géol. Fr.* **13**, 257–271.

Durst, H. 1985. Interpretation of the Reflection–Seismic Profile across the Northeastern Stavelot–Venn Massif and its Northern Foreland. *Neues Jb. Geol. Paläont. Abh.* **171**, 441–446.

Fielitz, W. 1987. Schieferung und Ausmaß der Innendeformation der paläozoischen Gesteine am Südost-Rand des Stavelot-Venn-Antiklinoriums in der Nordeifel (Rheinisches Schiefergebirge). Unpublished Ph.D. thesis, Rheinisch-Westfälische Technische Hochschule Aachen, Germany.

Fieremans, M. & Bosmans, H. 1982. Colour zones and the transition from diagenesis to low-grade metamorphism of the Gedinnian shales around the Stavelot Massif (Ardennes, Belgium). *Schweiz. miner. petrogr. Mitt.* **62**, 99–112.

Frank, B. 1987. Bestimmung des Metamorphosegrades der paläozoischen Schichten des Venn-Großsattels (Linksrheinisches Schiefergebirge) mit Hilfe der Illit-Kristallinität und Untersuchung der Zusammenhänge zwischen dem Metamorphosegrad und den regionalen tektonischen Verhältnissen. Unpublished Ph.D. thesis, Rheinisch-Westfälische Technische Hochschule Aachen, Germany.

Geukens, F. 1957. Contribution à l'étude du massif cambro-ordovicien de Stavelot en territoire allemand. *Mém. Inst. géol. Univ. Louvain* **20**, 165–210.

Geukens, F. 1961. De geologische struktuur rond het devillium massief van Falize-Ligneuville. *Meded. K. Vlaam. Acad.* **23**, 1–7.

Gray, D. R. 1981. Cleavage–fold relationships and their implications for transected folds: an example from southwest Virginia, U.S.A. *J. Struct. Geol.* **3**, 265–277.

Harland, W. B. 1971. Tectonic transpression in Caledonian Spitsbergen. *Geol. Mag.* **108**, 27–42.

Hayward, A. B. & Graham, R. H. 1989. Some geometrical characteristics of inversion. In: *Inversion Tectonics* (edited by Cooper, M. A. & Williams, G. D.). *Spec. Publs geol. Soc. Lond.* **44**, 17–39.

Hoffmann, K. 1961. Die Tektonik am Südrand des Hohen Venns. *Neues Jb. Geol. Paläont. Abh.* **111**, 341–367.

Holder, M. T. & Leveridge, B. E. 1986. Correlation of the Rhenohercynian Variscides. *J. geol. Soc.* **143**, 141–147.

Hutton, D. H. W. 1987. Strike-slip terranes and a model for the evolution of the British and Irish Caledonides. *Geol. Mag.* **124**, 405–425.

Knapp, G. 1962. Der varistische Gebirgsbau im Bereich der Oleftalsperre (Eifel). *Geol. Jb.* **79**, 399–422.

Knapp, G. 1980. Erläuterungen zur geologischen Karte der nördlichen Eifel 1:100,000. Geologisches Landesamt Nordrhein-Westfalen, Krefeld.

Korsch, R. J. & Wellman, H. W. 1988. The geological evolution of New Zealand and the New Zealand region. In: *The Ocean Basins and Margins Vol. 7B: The Pacific Ocean* (edited by Nairn, A. E. M., Stehli, F. G. & Uyeda, S.). Plenum Press, New York, 411–482.

Kramm, U. & Buhl, D. 1985. U–Pb Zircon dating of the Hill tonalite, Venn–Stavelot Massif, Ardennes. *Neues Jb. Geol. Paläont. Abh.* **171**, 329–337.

Kramm, U., Buhl, D. & Chernyshev, I. V. 1985a. Caledonian or Variscan metamorphism in the Venn–Stavelot Massif, Ardennes? Arguments from a K–Ar and Rb–Sr study. *Neues Jb. Geol. Paläont. Abh.* **171**, 339–349.

Kramm, U., Spaeth, G. & Wolf, M. 1985b. Variscan metamorphism in the NE Stavelot–Venn Massif, Ardennes: A new approach to the question of regional dynamothermal or contact metamorphism. *Neues Jb. Geol. Paläont. Abh.* **171**, 311–327.

Krohe, A. & Eisbacher, G. H. 1988. Oblique crustal detachment in the Variscan Schwarzwald, southwestern Germany. *Geol. Rdsch.* **77**, 25–43.

Lipshie, S. R., Oertel, G. & Christie, J. M. 1976. Measurement of preferred orientation of phyllosilicates in schists. *Tectonophysics* **39**, 91–99.

Lister, G. S. & Snoke, A. W. 1984. S–C mylonites. *J. Struct. Geol.* **6**, 617–638.

McClay, K. R. 1989. Analogue models of inversion tectonics. In: *Inversion Tectonics* (edited by Cooper, M. A. & Williams, G. D.). *Spec. Publs geol. Soc. Lond.* **44**, 41–59.

Massonne, H. J. & Schreyer, W. 1983. A new experimental phengite

barometer and its application to a Variscan subduction zone at the southern margin of the Rhenoherzynicum. *Terra Cognita* **3**, 187.
Matte, P. 1986. Tectonics and plate tectonics model for the Variscan belt of Europe. *Tectonophysics* **126**, 329–374.
Meilliez, F. 1989. Tectonique distensive et sédimentation à la base du Devonien, en bordure NE du Massif de Rocroi (Ardenne). *Annls Soc. Géol. N.* **CVII**, 281–295.
Meilliez, F. & Mansy, J.-L. 1990. Déformation pelliculaire différenciée dans une série lithologique hétérogène: le Dévono-Carbonifère de l'Ardenne. *Bull. Soc. géol. Fr.* **6**, 177–188.
Meissner, R., Bartelsen, H. & Murawski, H. 1981. Thin-skinned tectonics in the northern Rhenish Massif, Germany. *Nature* **290**, 299–401.
Meyer, W. 1986. *Geologie der Eifel*. E. Schweizerbartsche, Stuttgart.
Michot, J. 1976. Le segment varisque et son antécédent calédonien dans le cadre de la Belgique et des régions limitrophes. *Nova acta Leopold* **45**, 201–228.
Michot, P. 1980. Belgique. In: *Géologie des Pays Européens* (edited by Dunod-Bordas). 26e Congrés Géologique International, Paris, 483–577.
Mukhopadhyay, D. 1973. Strain measurements from deformed quartz grains in the slaty rocks from the Ardennes and the northern Eifel. *Tectonophysics* **16**, 279–296.
O'Brien, D. K., Wenk, H.-R., Ratschbacher, L. & You, Z. 1987. Preferred orientation of phyllosilicates in phyllonites and ultramylonites. *J. Struct. Geol.* **9**, 719–730.
Oertel, G. 1983. The relationship of strain and preferred orientation of phyllosilicate grains in rocks—a review. *Tectonophysics* **100**, 413–447.
Oldow, J. S., Bally, A. W., Avé Lallemont, H. G. & Leeman, W. P. 1989. Phanerozoic evolution of the North American Cordillera; United States and Canada. In: *The Geology of North America—An Overview* (edited by Bally, A. W. & Palmer, A. R.). Geological Society of America, 139–232.
Oncken, O. 1984. Zusammenhänge in der Strukturgenese des Rheinischen Schiefergebirges. *Geol. Rdsch.* **73**, 619–649.
Oncken, O. 1988a. Aspects of the reconstruction of the stress history of a fold and thrust belt (Rhenish Massif, Federal Republic of Germany). *Tectonophysics* **152**, 19–40.
Oncken, O. 1988b. Geometrie und Kinematik der Taunuskammüberschiebung—Beitrag zur Diskussion des Deckenproblems im südlichen Schiefergebirge. *Geol. Rdsch.* **77**, 551–575.
Passchier, C. W. & Simpson, C. 1986. Porphyroclast systems as kinematic indicators. *J. Struct. Geol.* **8**, 831–843.
Ramsay, J. G. & Huber, M. I. 1983. *The Techniques of Modern Structural Geology, Volume* 1: *Strain Analysis*. Academic Press, London.
Ramsay, J. G. & Huber, M. I. 1987. *The Techniques of Modern Structural Geology, Volume* 2: *Folds and Fractures*. Academic Press, London.
Rast, N. 1989. The evolution of the Appalachian chain. In: *The Geology of North America—An Overview* (edited by Bally, A. W. & Palmer, A. R.). Geological Society of America, 139–232.
Richter, D. 1962. Zur Baugeschichte der Ardennen II. Der geologische Bau des südwestlichen Teiles des Massivs von Stavelot (Belgien) unter besonderer Berücksichtigung seiner tektonischen Prägung. *Geol Mitt.* **2**, 283–364.
Sanderson, D. J., Andrews, J. R., Phillips, W. E. A. & Hutton, D. H. W. 1980. Deformation studies in the Irish Caledonides. *J. geol. Soc. Lond.* **137**, 289–302.
Sanderson, D. J. & Marchini, W. R. D. 1984. Transpression. *J. Struct. Geol.* **6**, 449–458.
Schmidt, W. 1956. Neue Ergebnisse der Revisions-Kartierung des Hohen Venns. *Beih. Geol. Jb.* **21**, 1–146.
Schwab, K. 1987. Compression and right-lateral strike-slip movement at the Southern Hunsrück Borderfault (Southwest Germany). *Tectonophysics* **137**, 115–126.
Simpson, C. 1990. Strain pathways in progressive deformation—a review. *Terra Abs.* **2**, 17.
Simpson, C. & Schmid, S. M. 1983. An evaluation of criteria to deduce the sense of movement in sheared rocks. *Bull. geol. Soc. Am.* **94**, 1281–1288.
Smith, R. B. 1977. Formation of folds, boudinage, and mullions in non-Newtonian materials. *Bull. geol. Soc. Am.* **88**, 312–320.
Soper, N. J. 1986. Geometry of transecting, anastomosing solution cleavage in transpression zones. *J. Struct. Geol.* **8**, 937–940.
Spaeth, G. 1969. Das Deformationsbild planolinear geschieferter Grauwacken aus dem Unterdevon am SE-Rand des Hohen Venns (Rheinisches Schiefergebirge). *Geol. Mitt.* **8**, 309–326.
Spaeth, G. 1971. Untersuchungen der Mineralregelung in Tonschiefern vom Südostrand des Hohen Venns (Rheinisches Schiefergebirge) mit Hilfe des Röntgen-Texturgoniometers. *Neues Jb. Geol. Paläont. Abh.* **139**, 99–123.
Spaeth, G. 1979. Neuere Beobachtungen und Vorstellungen zur variszischen Tektonik der westlichen Nordeifel (Rheinisches Schiefergebirge). *Z. dt. geol. Ges.* **130**, 107–121.
Spaeth, G., Fielitz, W. & Frank, B. 1985. Caledonian deformation and very low-grade metamorphism in the northeastern part of the Stavelot–Venn Massif. *Neues Jb. Geol. Paläont. Abh.* **171**, 297–310.
Teichmüller, M. & Teichmüller, R. 1979. Ein Inkohlungsprofil entlang der linksrheinischen Geotraverse von Schleiden nach Aachen und die Inkohlung der Nord–Süd-Zone der Eifel. *Fortschr. Geol. Rheinld. Westf.* **30**, 223–239.
Teichmüller, M., Teichmüller, R. & Weber, K. 1979. Inkohlung und Illit-Kristallinität. Vergleichende Untersuchungen im Mesozoikum und Paläozoikum von Westfalen. *Fortschr. Geol. Rheinld. Westf.* **27**, 201–276.
Thome, K. N. 1955. Die tektonische Prägung des Vennsattels und seiner Umgebung. *Geol. Rdsch.* **44**, 266–305.
Treagus, J. E. & Treagus, S. H. 1981. Folds and the strain ellipsoid: a general model. *J. Struct. Geol.* **3**, 1–17.
Walter, R. & Wohlenberg, J. 1985. Proposal for an ultra-deep research borehole in the Hohes Venn area (West Germany). *Neues Jb. Geol. Paläont. Abh.* **171**, 1–16.
Weber, K. 1976. Gefügeuntersuchungen an transversalgeschieferten Gesteinen aus dem östlichen Rheinischen Schiefergebirge (Ein Beitrag zur Genese der transversalen Schieferung). *Geol. Jb.* **D15**, 3–98.
Weber, K. 1981. Kinematic and metamorphic aspects of cleavage formation in very low-grade metamorphic slates. *Tectonophysics* **78**, 291–306.
Weber, K. & Behr, H. J. 1983. Geodynamic interpretation of the mid-European Variscides. In: *Intracontinental Fold Belts* (edited by Martin, H. & Eder, F. W.). Springer, Berlin, 427–469.
Werner, W. 1989. Synsedimentary faulting and sediment-hosted submarine-hydrothermal mineralization in the Late Palaeozoic Rhenish basin (Germany). *Geotekt. Forsch.* **71**, 1–305.
Williams, G. D., Powell, C. M. & Cooper, M. A. 1989. Geometry and kinematics of inversion tectonics. In: *Inversion Tectonics* (edited by Cooper, M. M. & Williams, G. D.). *Spec. Publs geol. Soc. Lond.* **44**, 3–15.
Williams, H. & Hatcher, R. D., Jr. 1983. Appalachian suspect terranes. In: *Contributions to the Tectonics and Geophysics of Mountain Chains* (edited by Hatcher, R. D., Williams, H. & Zietz, I.) *Mem. geol. Soc. Am.* **158**, 33–54.
Woodcock, N. H. 1990. Transpressive Acadian deformation across the Central Wales Lineament. *J. Struct. Geol.* **12**, 329–337.
Wrede, V. 1987. Der Einfluß des Brabanter Massivs auf die Tektonik des Aachen-Erkelenzer Steinkohlengebietes. *Neues Jb. Geol. Paläont. Mh.* **1987**, 177–192.
Ziegler, P. A. 1986. Geodynamic model for the Palaeozoic crustal consolidation of western and central Europe. *Tectonophysics* **126**, 303–328.