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Tectonophysics 275 (1997) 15–40

TECTONOPHYSICS

The deep crust of the Southern Rhine Graben: reflectivity and seismicity as images of dynamic processes

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Received 7 January 1996; accepted 22 June 1996

Abstract

The Rhine Graben, part of the European Cenozoic rift system, deserves special attention because of its location in the foreland of the Alpine orogen. The Phanerozoic evolution of the lithosphere in this region is defined by a set of major geodynamic events ranging from the Variscan orogeny, late-orogenic crustal re-equilibration to the interference of rifting and Alpine orogeny in Tertiary times.

The Rhine Graben is one of the most detailedly studied continental grabens. Prospecting for hydrocarbons in its sedimentary fill and intensive pre-site studies for the Continental Deep Drilling Program (KTB) on its eastern crystalline flank (Black Forest) provided a comprehensive set of geoscience data. Seismic investigations ranging from a deep seismic network of near-vertical reflection to concurrent wide-angle refraction experiments were accompanied by seismological and geological surveys. Therefore, a direct observational comparison between the two depth definitions of crustal subdivisions, elastic vs. strength/rheological, was possible. In large parts of Western Europe deep seismic sounding (DSS) investigation distinguishes between a crystalline upper crust of comparatively low P-wave velocities (5.9–6.0 km/s) with little and discrete reflectivity and a strongly reflective lower crust with higher P-wave velocities (about 6.4–6.8 km/s). Both are separated by the Conrad discontinuity. In contrast, strength and rheological considerations define a brittle upper crust and a ductile lower crust. The boundary between these two realms is usually referred to as the brittle/ductile transition zone which in seismically active regions is imaged by the deepest crustal hypocentres. The present brittle/ductile transition in the Rhine Graben region, as defined by the deepest hypocentres, does not necessarily coincide with the structural separation in an upper and lower crust as seen in deep seismic studies. This suggests that the laminated lower crust developed during the crustal re-equilibration phase at the end of the Variscan orogeny.

The second part of this paper concentrates on evidence of recent tectonic activities related to rifting processes impacting on the Variscan lower crust. A remarkable concentration of deep crustal earthquakes reaching to the top of an asymmetrical thinned lower crust is observed at or near the eastern border fault of the southern Rhine Graben and is possibly induced by the young rifting process.

We suggest that this asymmetry could be a consequence of the uppermost mantle detaching from the upper crust in collision with the Alpine orogen. This allows us to propose the distinction between the influence of compression, extension, continental subduction and intra-crustal detachment during graben formation. The dynamic interference of

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0040-1951/97/$17.00
PII S0040-1951(97)00014-0

rifting and foreland compression shown by structural and seismological observations leads to the notion of small-scale collisional escape tectonics associated with an intra-continental transform fault at the southern end of the Rhine Graben, the Rhine–Saône Transform zone.

Keywords: seismicity; crustal structure; lower crust; European Cenozoic rift system; Black Forest; reflection seismics; Rhine Graben

1. Introduction

The seismically active Rhine Graben, located in the immediate northern foreland of the Alps, offers the chance of a direct observational comparison between two types of depth-dependent crustal subdivisions, as defined by the distribution of seismic velocities and defined by strength and rheology. Furthermore, the Rhine Graben is an outstanding place to trace recent deep crustal deformation resulting from two ongoing and interfering processes, namely rifting and Alpine collision, and to relate them to tectonic structures observed at the surface. Our investigations are based on a wealth of geoscience data which were collected during the past four decades, making the Rhine Graben one of the best studied continental rifts.

In many parts of Western Europe deep seismic sounding (DSS), including near-vertical reflection and wide-angle refraction surveys has led to a characteristic subdivision of the crust into an upper part characterized by low P-wave velocities (5.9–6.0 km/s) with little and discrete reflectivity and a high-velocity lower crust (about 6.4–6.8 km/s) with strong, evenly distributed reflectivity. Both zones are separated by the historically defined Conrad discontinuity.

Another way to image the crust is defined by its rheological structure and its response to near static stresses; the brittle upper crust is deformed by fracture on faults whereas the ductile lower crust deforms by creep. The boundary between these two zones is referred to as the brittle/ductile transition zone (B/DT). In seismic active areas this B/DT is supposed to be marked by the deepest crustal hypocentres.

In this paper we compare both types of subdivision of the continental crust on the basis of a dense set of observational data giving information on structures imaged by elastic waves and the strength structure as derived from the spatial distribution of earthquakes.

There is a consensus that the crust/mantle boundary was re-equilibrated at the end of the Variscan orogeny in Western Europe during Permo-Carboniferous times (Klemperer, 1987; Eisbacher et al., 1989; DEKORP Research Group, 1990; Bois et al., 1991; Echtler and Chauvet, 1992; Mooney and Meissner, 1992). Therefore, we attempted to evaluate whether this process was associated with the development of the lower crustal lamination, as seen in some areas of Western Europe and whether the distribution of laminated lower crust in the area of the Rhine Graben is concordant with the actual ductile part of the crust in the Rhine Graben rift. Furthermore, we tried to find evidence of recent deformation patterns, possibly related to the rifting process either in the velocity structure or in the depth distribution of crustal earthquakes or in a combination of both observations.

We have found that regions of active tectonics (extensions or deep faulting) appear to correlate well with unusually thinned lower crust and extremely deep hypocentres. These earthquakes reach in the southeastern part of the Rhine Graben a depth of only 5 km above the crust–mantle boundary. The unusual depth of the hypocentres, their asymmetric concentration close to the eastern border fault of the Rhine Graben and regions of extension lead us to propose a detachment surface separating the mantle from the upper crust probably related to collisional coupling of latter with the Alpine orogen.

2. Geological setting

The Rhine Graben between Basel and Frankfurt forms the central part of the European Cenozoic rift system (ECRIS) which extends from the shore of the North Sea via the Rhône–Bresse Graben to the Mediterranean Sea (Fig. 1). In the area of the

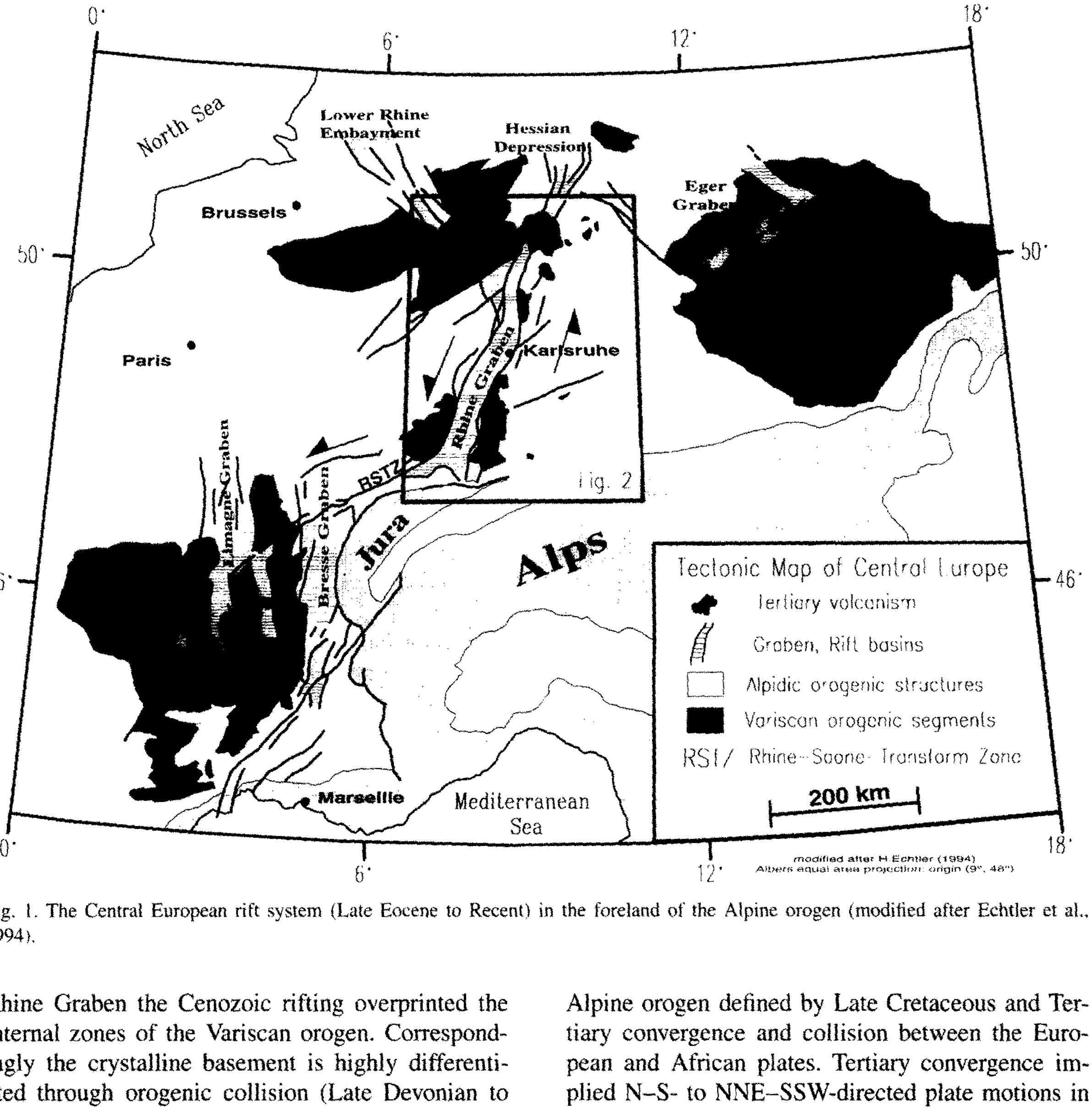


Fig. 1. The Central European rift system (Late Eocene to Recent) in the foreland of the Alpine orogen (modified after Echtler et al., 1994).

Rhine Graben the Cenozoic rifting overprinted the internal zones of the Variscan orogen. Correspondingly the crystalline basement is highly differentiated through orogenic collision (Late Devonian to mid-Carboniferous) and late-orogenic transtensional re-equilibration (Upper Carboniferous to Permian) with widespread granitization, uplift and exhumation and intra-montane basin formation (Eisbacher et al., 1989; Echtler and Chauvet, 1992). This Palaeozoic orogenic deformation generated major heterogeneities: first, a highly reflective lower crust (15–30 km), secondly, a transparent upper crust including some prominent reflective shear zones.

This crust was overprinted by the formation of the Rhine Graben in the northern foreland of the Alpine orogen defined by Late Cretaceous and Tertiary convergence and collision between the European and African plates. Tertiary convergence implied N–S- to NNE–SSW-directed plate motions in the Eocene and NW–SE-directed motion since the Miocene (Dewey et al., 1989). Southward subduction of European lower crust and lithospheric mantle, northward thrusting of decoupled upper crust, Moho-imbrication and southward backthrusting resulted in intense crustal stacking and thickening (Pfiffner et al., 1990). The Alps are still largely in a compressional regime. Recent seismicity and recent crustal uplift point to presently continuing collision and associated decoupling of upper crust (Deichmann and Baer, 1990; Valasek et al., 1991).

The Tertiary Rhine Graben trending roughly north-northeast between Basel (Switzerland) and Frankfurt (Germany) has a length of about 300 km and an average width of 36 ± 5 km. A northern and a southern segment can be defined based on asymmetric graben fill and differential subsidence (Doebl and Olbrecht, 1974). Horizontal extension does not exceed 10–15% (Illies, 1972, 1974; Meier and Eisbacher, 1991). Graben formation is mainly controlled by the reactivation of pre-existing faults.

3. Database

Since decades the Rhine Graben has received special attention from the international scientific community that led to several research programs. The major activities started with the *Upper Mantle Program* (Rothe and Sauer, 1967; Illies and Mueller, 1970) and the *Geodynamics Project* (Illies and Fuchs, 1974; Illies, 1981). The *International Lithosphere Program* continued the series of field observations and a number of reviews were published (e.g., Ziegler, 1992a). Most notable was the pre-site survey of two possible drilling sites for the *Deep Drilling Program* (*KTB*) in the Black Forest (the eastern shoulder of the Rhine Graben) and the Hohenzollerngraben (about 50 km south of Stuttgart). Within this survey a combination of CDP reflection surveys (supported by KTB/DEKORP) and wide-angle refraction experiments, supported by Deutsche Forschungsgemeinschaft (DFG) through the *KTB* and the *Special Research Program SFB 108* provided refined images of structural and physical properties of the lithosphere beneath a continental rift (Fuchs et al., 1987; Gajewski and Prodehl, 1987; Gajewski et al., 1987; Lüschen et al., 1987; Holbrook et al., 1988; Eisbacher et al., 1989; Emmermann and Wohlenberg, 1989; Lüschen et al., 1989; Meissner and Gebauer, 1989; Zeis et al., 1990). Furthermore, two combined French–German *ECORS/DEKORP* reflection traverses cross the Rhine Graben (Brun et al., 1991, 1992; Bois et al., 1991; Meier and Eisbacher, 1991; Wenzel et al., 1991; Campos-Enriquez et al., 1992; Rousset et al., 1993), supplemented by special reflection surveys of the SFB 108 in the southernmost part of the Rhine Graben (Echtler et al., 1994; Mayer et al., 1995) and furnished a detailed image of the crust and its regional variation. The locations of the reflection and the refraction profiles are shown in Fig. 2 with the epicentres of deep crustal earthquakes (depth range of hypocentres 15–24 km) concentrating at the southern end of the Black Forest. A dense network of seismological surveying in the same region provides a detailed inventory of seismogenic zones in the crust (Bonjer, 1997). The Alpine orogenic lithosphere has been studied during many years since 1956, last by the *European Geotraverse Project* of the ESF (Freeman et al., 1990; Blundell et al., 1992). A recent review of the Rhine Graben within the CREST group gives a detailed description of the geological history and the results and interpretations of the various DSS experiments (Prodehl et al., 1992, 1996).

3.1. Crustal velocity structure

Typical reflectivity patterns of the Variscan crust in extension are presented in Fig. 3 as CDP-reflection seismic sections from various profiles. In the three migrated sections a clear separation in depth between a transparent upper crust, a strongly reflective lower crust and a transparent upper mantle can be recognised. The first profile (Fig. 3a) runs parallel to the Rhine Graben along its eastern shoulder (Lüschen et al., 1987; Meissner and Bortfeld, 1990), the second one (Fig. 3b) cross-cuts the rift (Bartelsen et al., 1982; Lüschen et al., 1987; Brun et al., 1991) and the third one (Fig. 3c) is located more than 100 km to the east (DEKORP Research Group, 1985; Behr and Heinrichs, 1987).

At a first glance, one cannot recognise any effect of the graben tectonics at the top of the lower crust (Fig. 3b). The graben is much better seen in the topographic features than in the morphology of the top of the reflective lower crust. The classical Conrad discontinuity between upper and lower crust defined as a velocity increase by refraction observation appears as a '*foggy*' boundary in near vertical reflection observations. If the geography and topography are not considered in the sections, the position of the graben would be difficult to locate only from the structure of the lower crust. The graben could also be placed at any other site where strong lateral variations in Conrad topography are apparent (e.g., see southern part of DEKORP-2S in Fig. 3c). Only the presence of the sedimentary trough seen in connec-

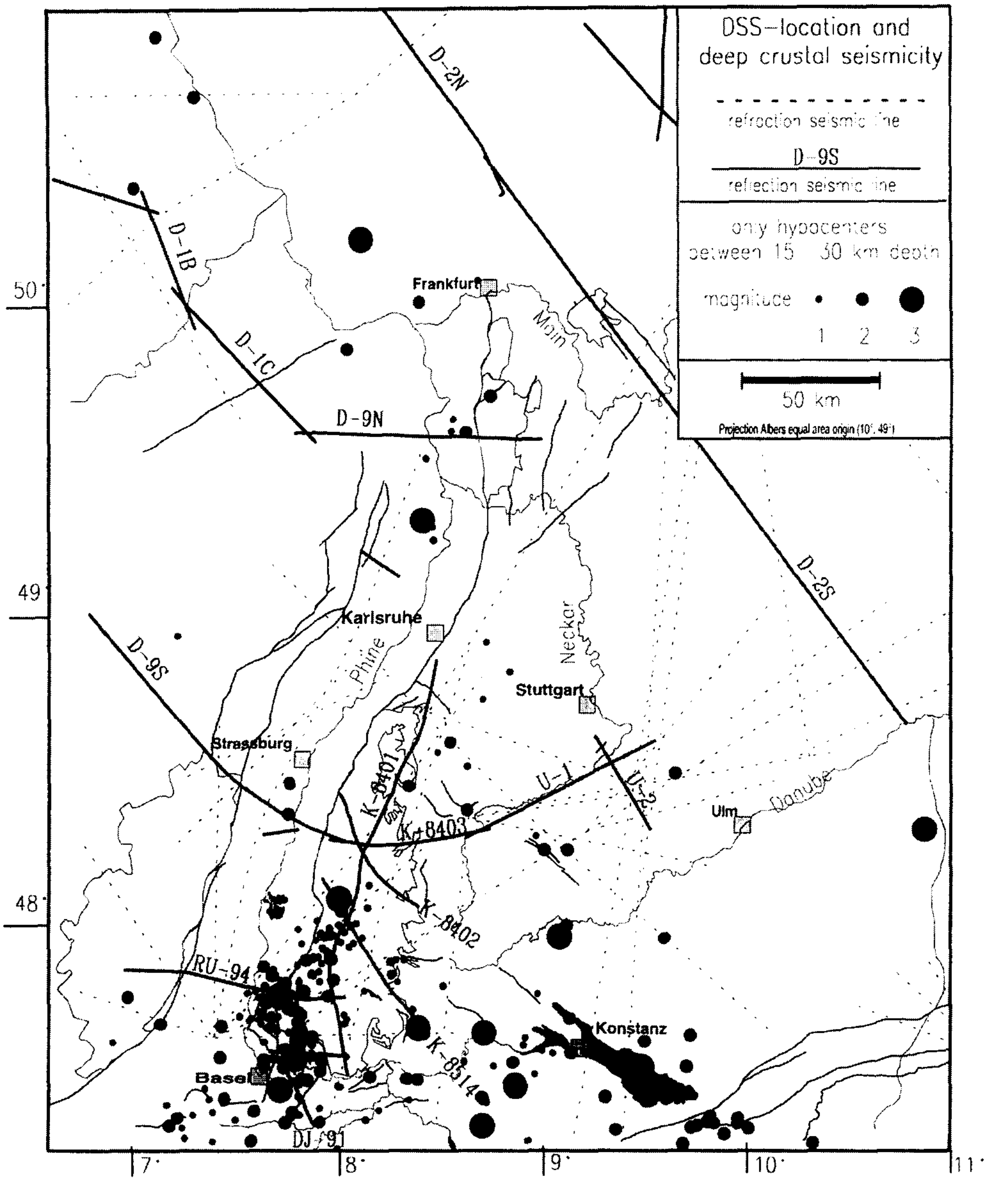


Fig. 2. Location map of the refraction (finely slashed) and reflection (labelled solid lines) seismic profiles in southwest Germany and eastern France (compare Fig. 1) including the deep crustal earthquakes (filled circles with radii growing with magnitude) with hypocentre between 15 and 30 km depth (see Bonjer, 1997). Note the concentration of deep crustal seismicity and the dense network of reflection seismic investigations in the southernmost part of the Rhinegraben and the southern Black Forest.

tion with a broad and weakly updoming of the Moho provides an indication of the location of the graben. Indeed, the comparison of the three profiles carries the message that tectonic processes in the lower crust are only a second-order phenomenon compared with those visible in the topography and in the brittle upper crust.

Despite this general message a detailed look at the southern traverse DEKORP-ECORS-9S (Fig. 4, interpreted linedrawing of the central part of Fig. 3b) exhibits apparently a lowering of the reflective lower crust, leading to the interpretation of a asymmetrical thinning (Brun et al., 1991; Wenzel et al., 1991). The near-vertical profiling was supplemented

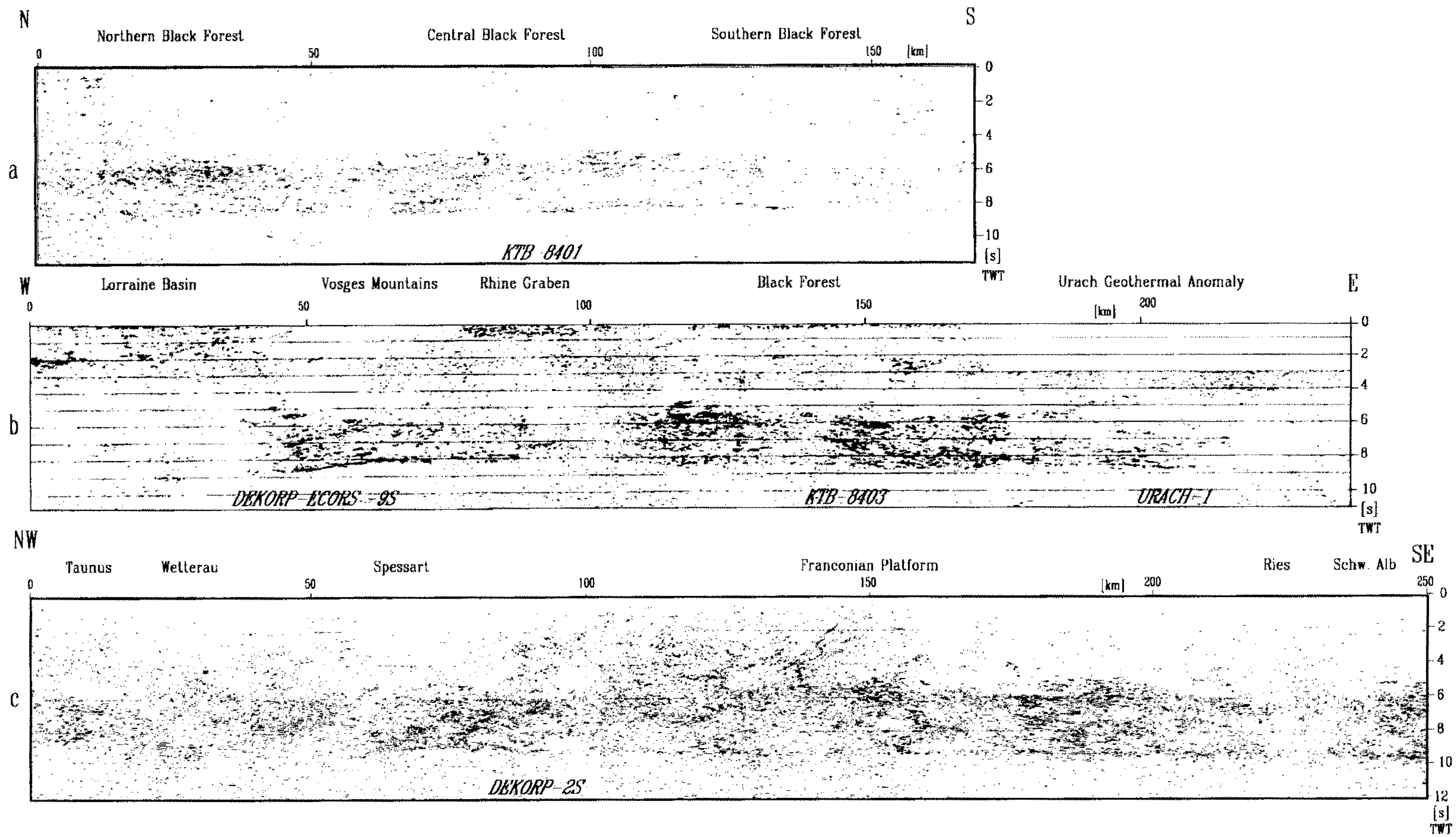


Fig. 3. Three observed seismic CDP-reflection sections in SW Germany. In all three migrated time sections from steep angle reflection observation a clear separation in depth appears between a more or less transparent upper part with some discrete reflectors, a strongly reflective lower part of the crust and again a transparent upper mantle. (a) N–S-striking profile KTB-8401. The upper crust contains mostly dipping, often cross-cutting local events. The reflective lower crust is characterised by horizontally layered reflections. In the central part of the Black Forest it starts at an almost constant level of 5 s TWT, corresponding to a depth of 14–15 km, slightly dipping towards the south (after Lüschen et al., 1987; Meissner and Bortfeld, 1990). (b) Combination of three migrated reflection seismic sections (ECORS-DEKORP 9S, KTB-8403, URACH1) (Bartelsen et al., 1982; Lüschen et al., 1987; Brun et al., 1991; compilation distributed by Lüschen, 1994). A broad, relatively flat updoming of the crust–mantle boundary and the strong reflections from sedimentary fill point to the position of the rift. (c) The NW–SE-striking profile DEKORP 2-S is located more than 100 km east of the rift. In the migrated section the top of the reflective lower crust varies between 4.7 s and 5.8 s TWT and is thus comparable to the observed variations of this boundary beneath the Rhinegraben. However, a large-scale updoming of the Moho is here not visible (DEKORP Research Group, 1985; Behr and Heinrichs, 1987). Try to find the location of the Rhine Graben without prior knowledge of geography. Note the fogginess of the top of the reflective lower crust (Conrad discontinuity) and the sharpness of the Moho as its lower bound.

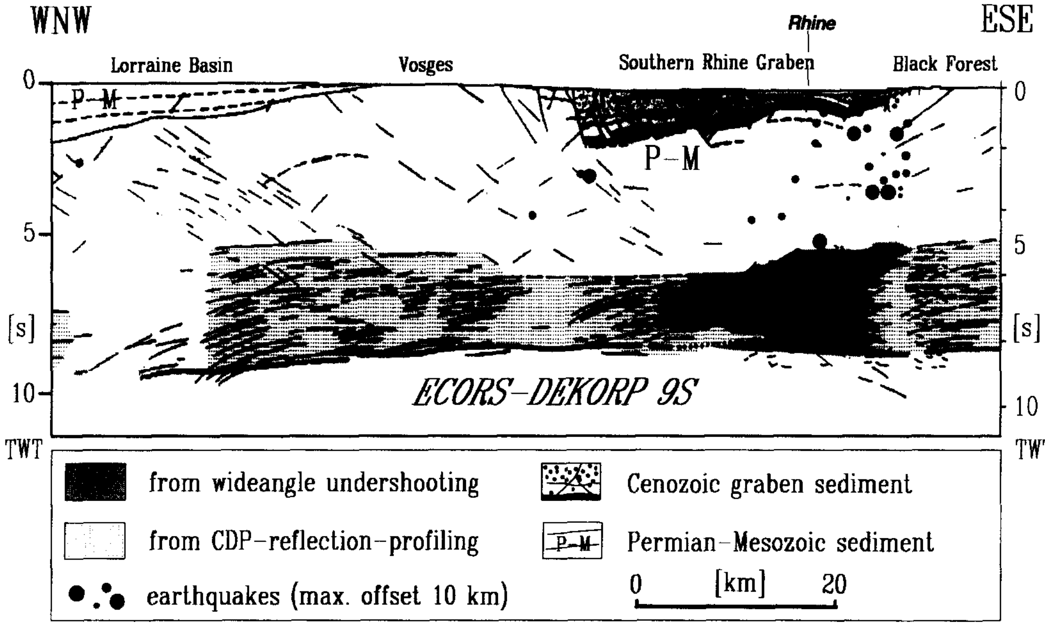


Fig. 4. Interpretative linedrawing of the ECORS-DEKORP 9S southern Rhinegraben seismic profile (modified after Brun et al., 1991; Wenzel et al., 1991), including hypocentres with a maximal lateral offset of 10 km from the section of the seismic line; for location see Fig. 2 and for comparison also Fig. 3b. The reflective lower crust appears beneath both rift shoulders between 14–15 km and 28–30 km depth and is overlain by a more or less transparent upper crust. Beneath the Rhinegraben a large-scale updoming of the crust–mantle boundary and a small-scale increase in the depth of the lower crust, documents a possible asymmetrical thinning. The strong reflections from the sediments in the Rhinegraben proper allow no clear imaging of the lower crust in this part of the section. On the profile ECORS-DEKORP-9S wide-angle undershooting experiments enabled the detection of the reflective lower crust reaching from the shoulders into the Graben proper emphasising the asymmetric shape of the thinned lower crust on its eastern side (Echtler et al., 1994).

by wide-aperture undershooting observations around the eastern border fault which support the former interpretation of a lower crustal thinning (Echtler et al., 1994). The results of steep and wide-angle observations show a clear and sharp lower boundary of the reflective lower crust — presumably the Moho — and a less distinct upper boundary — the '*foggy*' Conrad. So far, the reflectivity pattern is understood as an accumulation of elongated heterogeneities (lamellae) which back-scatter or reflect the seismic waves (Wenzel et al., 1987; Sandmeier and Wenzel, 1990).

These heterogeneities are thought to have developed during the post-Variscan (Permo–Carboniferous) regional re-equilibration of the Moho discontinuity; however, it cannot be excluded that Cenozoic rifting activity also contributed to their development (Klemperer, 1987; Eisbacher et al., 1989; DEKORP Research Group, 1990; Bois et al., 1991; Echtler and Chauvet, 1992; Mooney and Meissner, 1992).

The question arises whether the Cenozoic rifting processes also affected this lower crust and whether this could be recognized? Is the observed asymmetric thinning of the lower crust due to Cenozoic rifting? If this were the case, how would the lower crust reflectivity pattern be modified? Were new heterogeneities generated, or do we just see modified lamellae? How can we distinguish between recent and palaeo-lamellae?

There are hints from previous Vibroseis undershooting experiments (Damotte et al., 1987; Lüschen et al., 1987) that there are significant differences in the physical nature of the lower crust between the rift shoulders and the axial rift zone. Beneath the graben the reflective elements appear much stronger, continuous and concentrated in the uppermost part of the reflective lower crust. This points to a modification of palaeo-lamellae. Those questions will be further addressed below in the light of new seismic investigations in the southernmost part of the Rhine Graben.

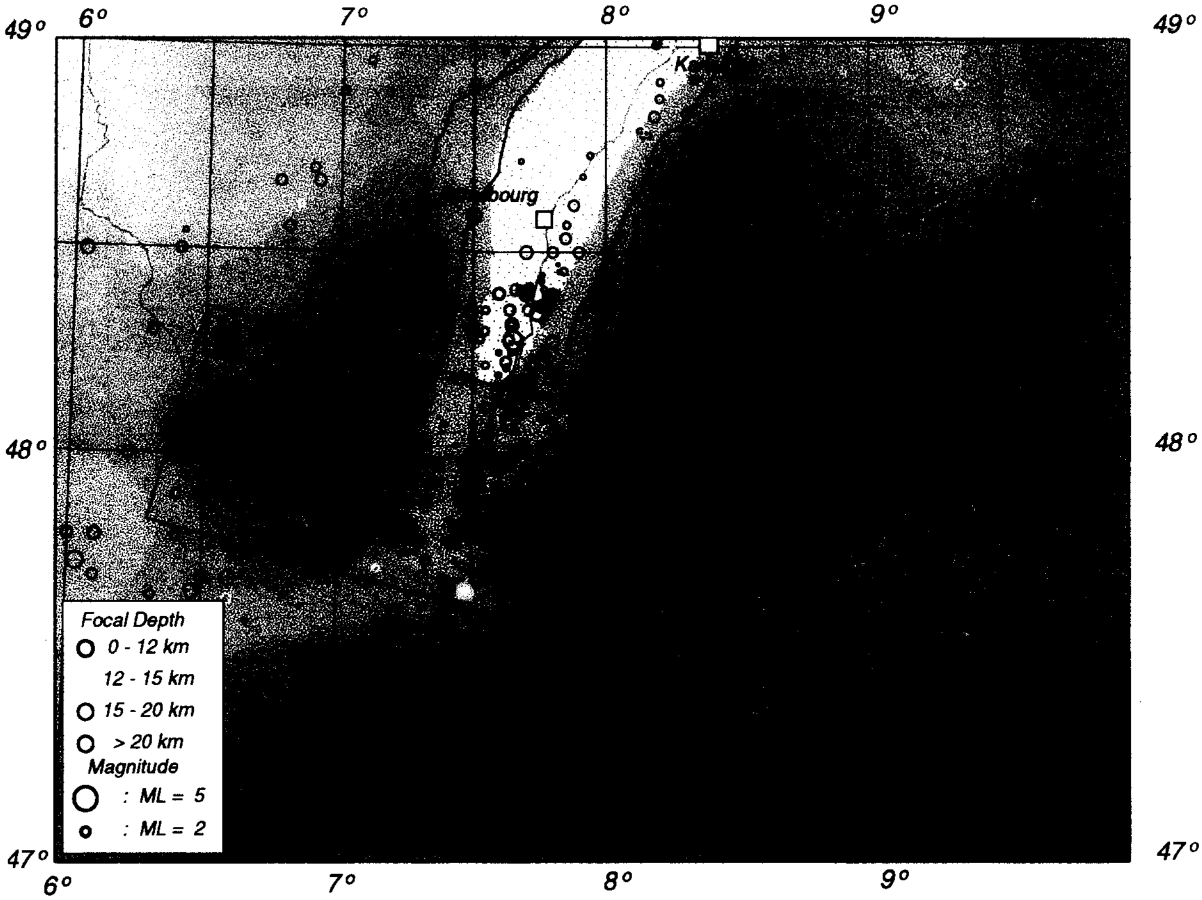


Fig. 5. Spatial distribution of earthquakes in the southern Rhinegraben, derived from a dense seismic network operating from 1971 until present. The depth ranges of hypocentres are colour-coded, circle-size indicates the magnitude of the earthquakes (modified after Bonjer, 1997). The rectangle overlain denotes the area for the hypocentre cross-section in Fig. 7 (bottom). Note the concentration of deep hypocentres around the eastern border fault.

3.2. Seismicity and stress

All active rifts are characterized by earthquake activity. Similarly the seismicity of active orogens provides an insight into ongoing dynamic processes. Therefore, the area of the Southern Rhine Graben is ideally located to evaluate the relative importance of dynamic processes related to rifting and collisional coupling of the foreland with the Alpine orogen. Fig. 5 provides a summary of the seismic activity in the southern Rhine Graben and the southward adjacent Jura Mountains and Molasse Basin.

An interference of these two seismicity patterns is observed with an obvious concentration of activity in the southernmost part of the Rhine Graben. It is characterised by a distinct asymmetry in the spatial distribution of epicentres both in historical records and recent activity (Bonjer et al., 1984, 1989; Bonjer, 1992, 1997). A crustal cross-section of earthquake foci extending from the Vosges across the Rhine Graben to the Black Forest shows a drastic increase in the maximum depth of hypocentres from west to east (Fig. 5). Along the eastern border fault hypocentres reach almost down to the crust–mantle boundary.

Furthermore, a dense network of seismographic stations in the region of the southern Black Forest allowed not only precise earthquake localisation, but also the determination of focal mechanisms (Deichmann and Baer, 1990; Ebel and Bonjer, 1990; Dahm and Bonjer, 1992; Bonjer, 1997; Plenefisch and Bonjer, 1997). In the European part of the World Stress Map, based on fault plane solutions, hydraulic fracturing, overcoring measurements and breakout analysis of borehole data, the Rhine Graben area exhibits practically no deviation from the general Western European direction of S_{Hmax} (N145°E) (Müller et al., 1992, 1997; Müller, 1993). As a second-order effect we can observe a counterclockwise rotation of the S_{Hmax} direction of about 30° from the Swiss Jura in the south (N160°E) to the lower Rhenish embayment in the north (N130°E); this observation derived from fault plane solution is probably caused by the geometry of the Cenozoic rift system (Pavoni, 1980; Ahorner et al., 1983; Deichmann, 1992b; Plenefisch and Bonjer, 1997).

The distinct seismicity throughout the Rhine Graben and the analysis of borehole data provide important information on stress and stress releases and therefore about the rheology of the crust. The first astonishing finding is that the general Western European direction of S_{Hmax} (N145°E) is hardly affected by the Rhine Graben. Secondly, only for the minimum principal stress axes σ_3 a dominant NE–SW-directed trajectory could be defined. Thirdly, hypocentres reach to unexpected depths compared to classical rheological notions (Müller et al., 1997; Bonjer, 1997).

A detailed look at the fault plane solutions from the Southern Rhine Graben shows an interchange between strike-slip (SS) and normal faulting (NF) events with very few exceptions of thrust faulting (TF). Therefore only σ_3, the smallest principal stress axis, displays a constant direction (NE–SW). Analysing the fault plane solutions at the southern end of the Rhine Graben, including the deep events around the eastern border fault, a systematic change with depth is indicated. The earthquakes in the upper crust down to 15 possibly 18 km show dominantly a strike-slip regime, whereas in the deeper parts normal faulting mechanisms prevail slightly (Plenefisch and Bonjer, 1997).

The question arises whether the observed asymmetry in earthquake locations and the tendency of normal faulting mechanism to become more important at depth is possibly caused by the combined effect of collisional coupling of the upper crust and its detachment from the uppermost mantle.

4. Velocity structure versus strength of the crust — a comparison

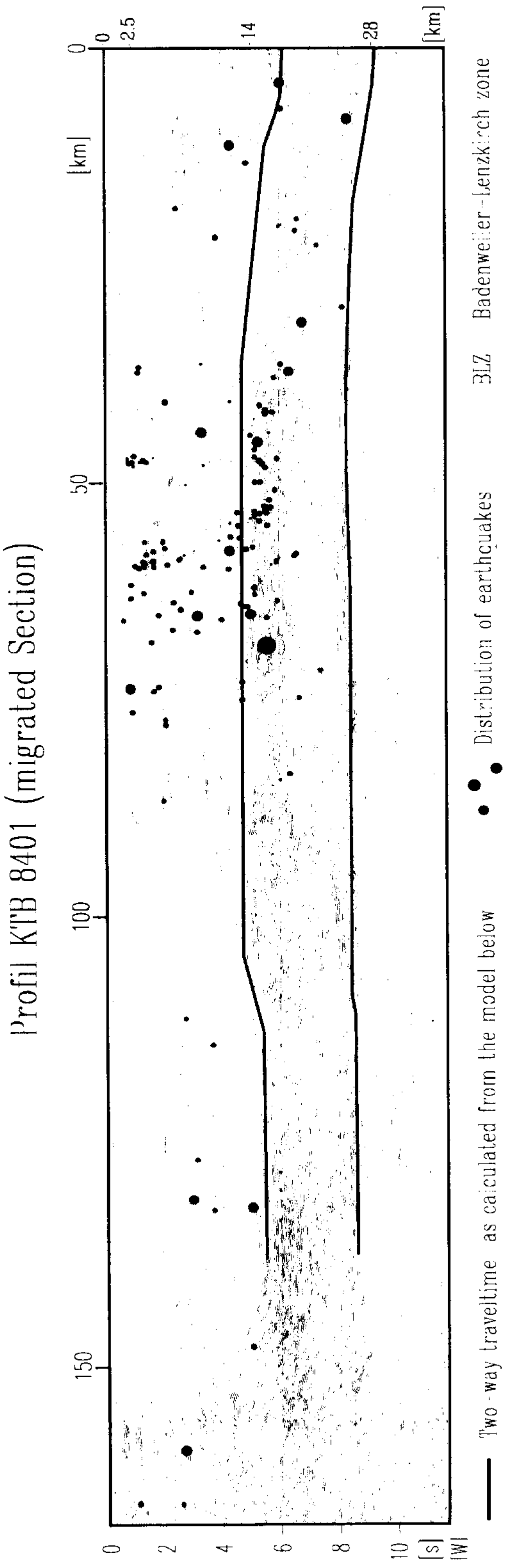
In this section we compare the velocity structure of the crust and its strength/rheology as imaged by the deep crustal seismicity. Although it has been repeatedly pointed out that the reflective lower crust with its strongly elongated heterogeneities images deformation processes in a ductile regime (Wever et al., 1987; Green et al., 1990), the deepest crustal earthquakes are interpreted as originating from the brittle–ductile transition zone. In the Southern Rhine Graben we can directly compare both notions on the lower crust.

Firstly, we will compare the observations on the crustal velocity structure and seismicity along two crustal transects and secondly we discuss strength models compatible with the observed seismicity.

Lessons from this section are twofold. Firstly, it will be shown that the reflectivity of the lower crust observed in seismic experiments corresponds to palaeo-deformations. Deep crustal hypocentres within the reflective lower crust show that the present brittle regime reaches far into the reflective lower crust, at least below the southern Black Forest. Secondly, classical strength models of the crust have on the one hand difficulty in moving the brittle–ductile transition as close to the Moho as observed beneath the eastern flank of the Rhine Graben and on the other hand to explain the asymmetric appearance of deep crustal seismicity.

4.1. Crustal structure — seismicity

The crustal structure and seismicity of the Southern Rhine Graben area was analysed along two transects. The first transect strikes N–S and is located on the eastern, uplifted rift shoulder of the Black Forest; it is controlled by the KTB-8401 reflection line and a coincident refraction survey (Section 4.1.1, Fig. 6). The second transect crosses the southern most part of the Rhine Graben 25 km to the north of Basel; it is based on a compilation of several refraction/reflection surveys, but is not controlled by a through-going combined reflection/refraction line (Section 4.1.2, Fig. 7).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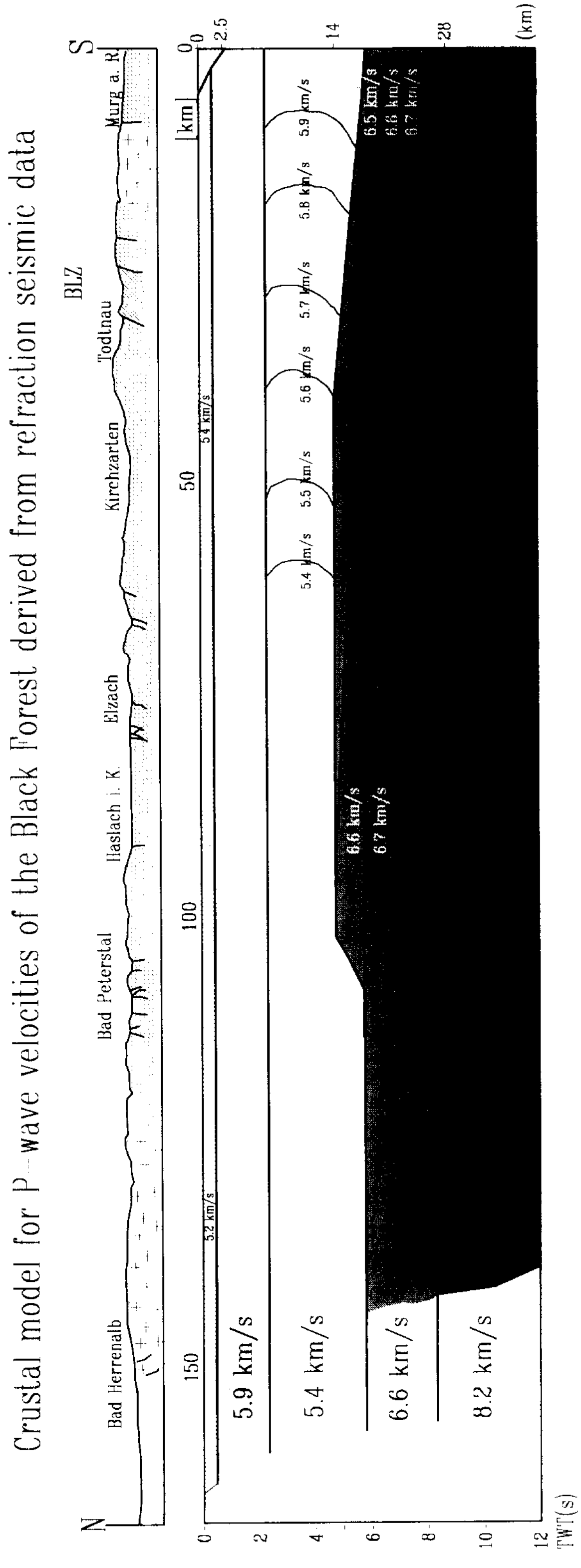
Two-way traveltime as calculated from the model below
Distribution of earthquakes
BLZ Badenweiler-Lenzkirch zone


Crustal model for P-wave velocities of the Black Forest derived from refraction seismic data
N
S
Bad Herrenalb
Bad Peterstal
Haslach i. K.
Elzach
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Todtnau
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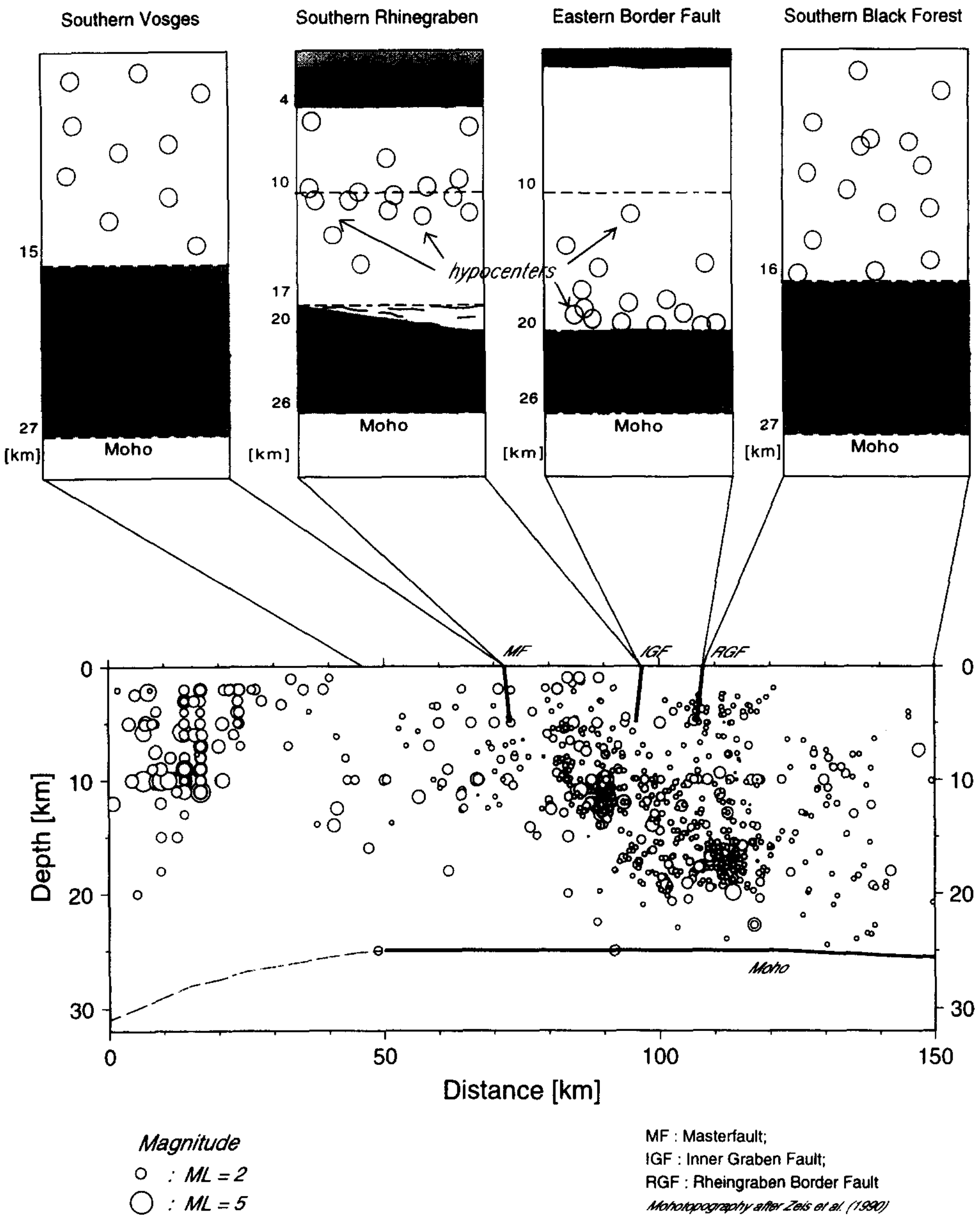

Fig. 7. Conceptual diagram of lower crustal heterogeneities vs. hypocentres (top) and hypocentre cross-section (bottom). The striking features in these diagrams are the concentration of deep crustal earthquakes around the eastern border fault (RGF) and the deepening of hypocentres from west to east, visible in the synthetic as well as in the hypocentre cross-section (for location, see Fig. 5).

Fig. 6. Joint interpretation of coinciding reflection and refraction seismic profiles on the eastern rim including earthquakes with magnitude $M_L > 1$ and a maximum distance of 10 km from the section of the seismic line (Bonjer, 1997). Top: the migrated reflection seismic time-section of the CDP-line KTB-8401 (after Lüschen et al., 1987; Meissner and Bortfeld, 1990). Depth positions (black lines) for the Moho and Conrad discontinuities are calculated with the velocity model (bottom) derived from the coincident refraction seismic studies. Note the coincidence of the crust–mantle boundary defined by the P-wave velocities (refraction seismic) and by the mode of reflectivity (reflection seismic). Filled circles indicate earthquakes. Note that the deepest hypocentres reach to the south increasingly into the reflective lower crust. Bottom: 2-D velocity model modified after Gajewski and Prodehl (1987) of the Black Forest derived from the SFB 108 refraction seismic experiment. Note that the low-velocity zone directly on top of the lower crust is only detected in the refraction experiment.

4.1.1. N–S transect on eastern rift shoulder

Within the observational accuracy, a strong correlation could be established between the lower crust, as defined by the refraction velocities ($V_p > 6.5$ km/s) and the reflective lower crust, as defined by an increase in reflected or back-scattered energy (Fig. 6). However, it must be noted that the reflectivity of the lower crust varies along the transect and decreases perceptibly in its southern part. An outstanding feature of this transect is the presence of earthquake foci within the reflective lower crust beneath the southern Black Forest in the area where its reflectivity weakens (Fig. 6, top). This is taken as a strong indication that the reflective lower crust as seen in seismic experiments, is not an image of the present ductile lower crust. Therefore, the observed laminated heterogeneities may not be related to present deformations in the lower crust. Moreover, this shows that the brittle regime extends well into the reflective laminated lower crust.

The presence of hypocentres embedded in the reflective laminated lower crust suggests, that the deformations responsible for the development of a reflective lower crust predate the present deformation. There is now a consensus that the elongated heterogeneities (lamellae) in the lower crust had been formed during the Permo–Carboniferous re-equilibration under a different thermal regime and under higher strain rates.

At this point we have to note that the deep hypocentres were projected up to 10 km into the KTB reflection section. A more detailed local survey of seismicity and velocity structure will be discussed in Section 5.

4.1.2. E–W transect

An E–W cross-section of hypocentres (Fig. 7, bottom) is derived from a 40-km-wide zone (selected area, see Fig. 5) crossing the Rhine Graben almost E–W from the Vosges Mountains to the Black Forest. Unfortunately, available reflection seismic profiles which traverse the entire rift (DEKORP-ECORS 9N and 9S) are located in regions of weak or shallow seismicity.

Therefore, Fig. 7 (top) shows different observations on crustal structures. For the *Vosges* and the *Southern Rhine Graben* the reflection seismic line DEKORP-ECORS-9S and refraction seismic observations from shotpoint Steinbrunn are used (Dohr, 1967; Edel et al., 1975; Damotte et al., 1987; Brun et al., 1991; Wenzel et al., 1991; Prodehl et al., 1996). For the *eastern Border Fault* and the *Southern Black Forest* the reflection seismic lines KTB-8401 and KTB-8514 and coinciding refraction seismic observations contribute to this image (Edel et al., 1975; Gajewski and Prodehl, 1987; Lüschen et al., 1987).

The striking features in these diagrams are the concentration of deep crustal earthquakes around the eastern border fault (RGF) and the deepening of hypocentres immediately to the east of it. Beneath the Vosges mountains and the Southern Rhine Graben the hypocentres reach maximum depths of less than 15 km and stay therefore clearly above the reflective lower crust. Near the eastern border fault, beneath the western margin of the Black Forest, the maximum depths of the hypocentres reach almost to the crust–mantle boundary and hence enter the reflective lower crust, as already seen on the N–S profile. Summarising the results, we observe regions of thinned lower crust where the hypocentres concentrate right above its top, neighbouring regions of non-modified Variscan crust with hypocentres inside the reflective lower crust.

4.2. Seismicity versus strength and rheology

Seismicity and strength within the lithosphere are connected by the classical rheological model in which the lithosphere is strained independently of depth at a constant rate. The strength of the crust is defined differently in its brittle and ductile part (Fig. 8). In the brittle upper crust the strength is limited by the Brace–Byerlee law. The frictional resistance of a prefractured crust grows with depth under the influence of increasing load (Brace, 1965; Byerlee, 1968, 1978; Brace and Kohlstedt, 1980; Kohlstedt et al., 1995). At the brittle–ductile transition (B/DT) the frictional strength equals the shear stress necessary to produce the constant strain rate required by the model (Meissner and Kuznir, 1987; Rutter and Brodie, 1992). Beneath the B/DT the ductile lower crust will yield by creep following a Weertman law (Weertman, 1968). At a constant strain rate $\dot{\epsilon}$ the shear stress resistance in the ductile crust drops strongly with increasing temperatures

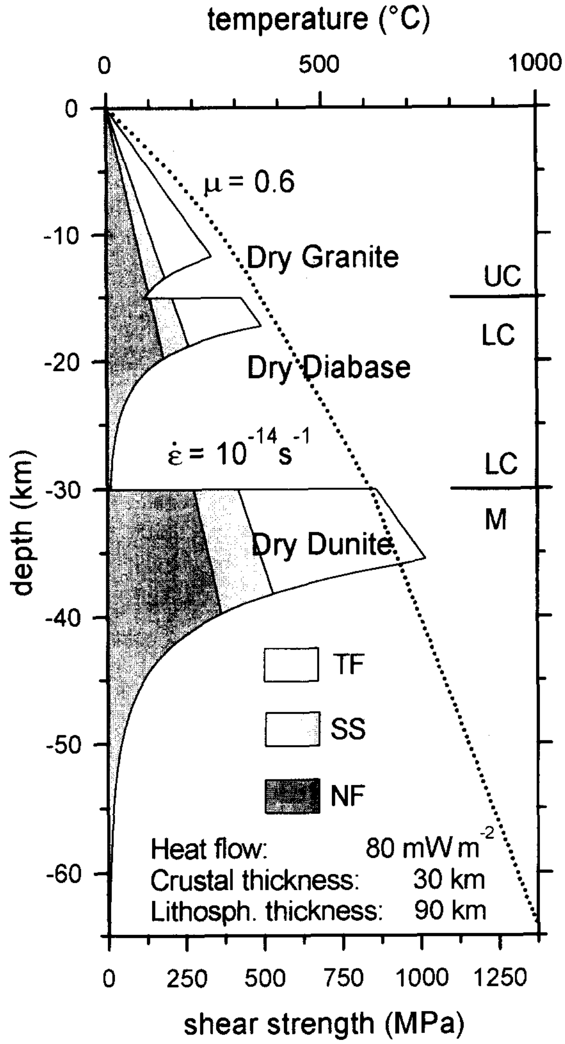

Fig. 8. Strength diagram for western Europe calculated for a constant strain rate of $\dot{\epsilon} = 10^{-14}$ s^{-1} according to the Brace–Byerlee law (Müller et al., 1997). The lithological layering for the crust is 15 km granitoid followed by 15 km gabbroic material. These strength curves are calculated for three different stress regimes (*NF* = normal faulting, *SS* = strike slip, *TF* = transform faulting). The geotherm (dotted) corresponds to a heat flow of 80 mW m^{-2}.

along almost any conceivable geotherm. The change in composition from felsic (quartz-rich) to mafic (feldspar) may cause an intermediate increase in the ductile shear strength dropping again with the increase in depth and temperature until the crust–mantle boundary is reached (e.g., Chen and Molnar, 1983; Cloetingh and Banda, 1992; Downes, 1993). At the crust–mantle boundary, the change from mafic to ultramafic (olivine) composition raises the ductile shear strength by orders of magnitude. However, the effect of increasing temperature lowers this shear strain resistance again with a rheology following the Weertman law (e.g., Brace and Kohlstedt, 1980; Kohlstedt et al., 1995)

In this classical rheological strength model of the lithosphere the earthquakes are considered to be restricted to the brittle upper crust where the deepest hypocentres are supposed to serve as markers of the brittle–ductile transition zone. Within this notion of crustal strength with constantly strained lithosphere and a crustal thickness of 30 km, how can the depth of the brittle–ductile transition be deepened nearly down to the Moho?

The depth level of the B/DT depends on the geotherm, the strain rate, the composition, the grain size, the distribution of pre-existing crustal discontinuities and on the presence or absence of fluids. It is increasing with higher strain rates, with lower temperatures and with larger grain size. The role of fluids in the crust is not so clear. The presence of free fluids on existing faults in the upper part of the crust effectively decreases the load pressure and therefore steepens the Brace–Byerlee curve (frictional sliding). Consequently, the B/DT moves to greater depths. The influence of fluids in the lower part of the crust depends on the petrological composition.

Considering a reasonable geotherm (corresponding to a heatflow of 60–80 mW/m^2, Wilhelm et al., 1989, 1994) and a standard petrological model for the crust (Downes, 1993; Meissner and Strehlau, 1982), we have not only the difficulty to understand the asymmetry in the spatial distribution of earthquakes but also the deep hypocentres at or beneath the eastern shoulder of the Rhine Graben which reach almost to the crust–mantle boundary located at a depth of 25 km (Bonjer, 1997). These deep hypocentres near the crust–mantle boundary beneath the southern Black Forest would either require unusually high strain rates, relative cold temperatures, a very mafic lower crust, or the unknown effect of fluids.

5. Seismic targeting on deep crustal seismicity

In this section we will summarise the results of recent deep seismic sounding experiments which were

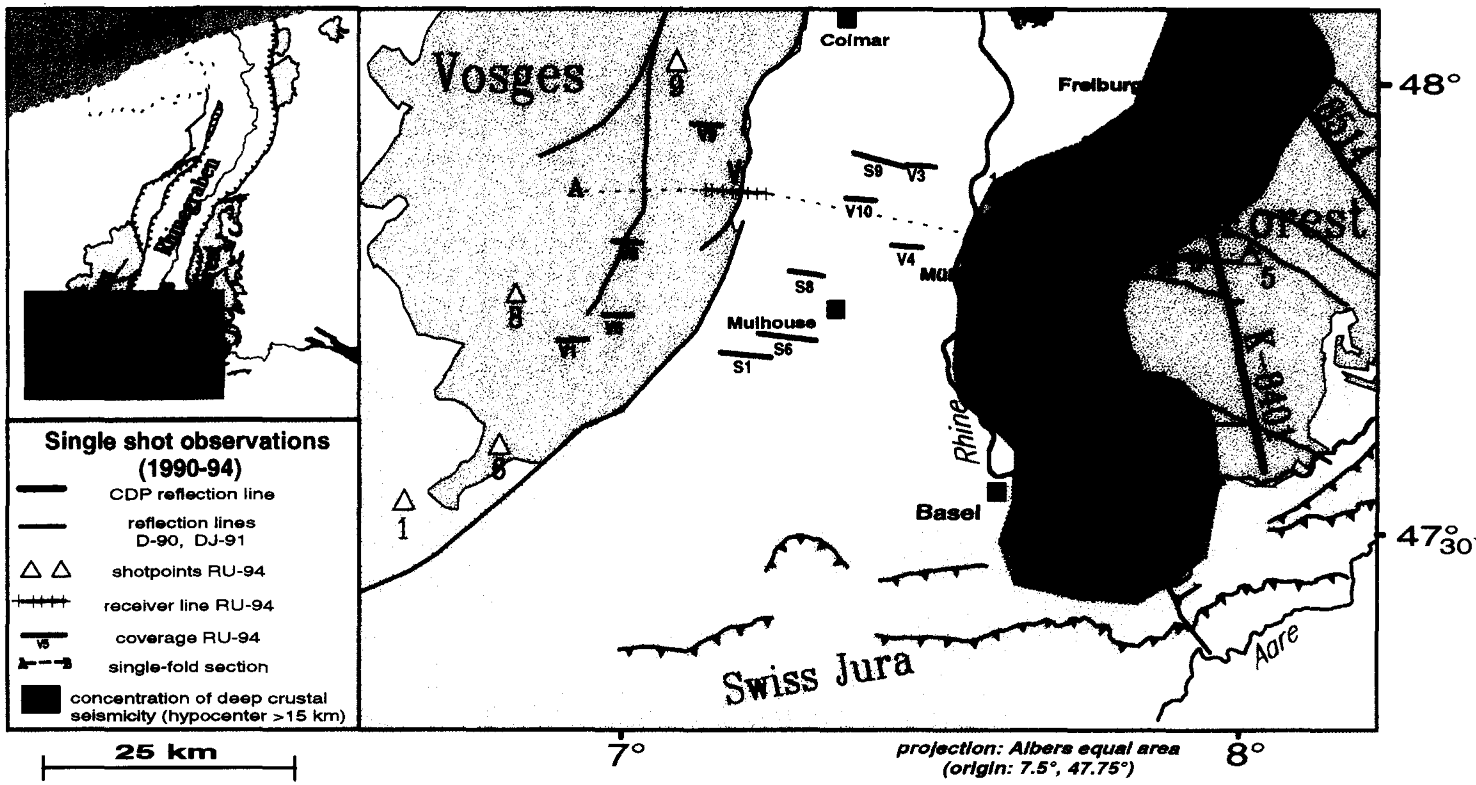

Fig. 9. Location map of reflection seismic investigations in the southernmost part of the Rhinegraben. The multi-fold KTB profiles (K-8401, K-8514) miss the region of the deep crustal earthquakes (green shaded area) while the experiments D-90 and DJ-91 are located on top of this zone. The RU-94 reflection experiment consists of single-shot observations reaching from near-vertical to near-critical reflection distances. The location of the midpoint coverage is given by the labelled line segments (*S1-S10, V1-V10*).

targeted on the region of the deepest seismic activity in the crust beneath the southern Black Forest.

The following seismic investigations started as an attempt to obtain structural velocity information directly in the region of the deep hypocentres, i.e., at the southern tip of the Black Forest in the Dinkelberg area and close to the eastern border fault (Fig. 9). Both location maps (Figs. 2 and 9) show that the southern part of KTB reflection line 8401 is situated east of the region where the deep crustal earthquakes are concentrated. Therefore a new survey was necessary to test whether the apparent presence of hypocentres in the reflective lower crust observed on the southern end of the KTB reflection profile, is not just an effect of projection over a distance of about 10 km or whether the hypocentres are indeed located within the reflective lower crust. The outcome of the series of experiments was a surprise. It led to the discovery of traces of deformations very likely connected with the young rifting process.

The density of digital seismic stations in the network of the southern Black Forest was increased to improve the location capability in depth (1 km) and in epicentre coordinates (1.5 km) (see Bonjer, 1997). Furthermore, two seismic-reflection profiles were recorded in the region of deep seismic activity (Fig. 9, green zone). Profile D-90 extends from the southern end of KTB-8401 across the intermediate Dinkelberg fault block to the eastern border fault of the Rhine Graben. Profile DJ-91, which was operated together with the Swiss National Research Program NFP-20, was recorded east of the Rhine Graben border fault and extends from south of Freiburg, across the Dinkelberg block into the Jura Mountains (J. Ansorge and R. Horstmeyer, pers. note; NFP-20 Atlas, 1996).

Additionally, a near to wide-angle survey (RU-94) was carried out with source and receiver positions on both rift shoulders, traversing the graben at its southern end (Mayer et al., 1995; Bonjer, 1997).

5.1. The Dinkelberg E–W profile

The first reflection-seismic profile in the Dinkelberg is the E–W profile. It was recorded with single-fold coverage by a 3-component receiver array (Echtler et al., 1994; Mayer et al., 1995).

The dominant structure in the P-wave section, given as a linedrawing in Fig. 10 (bottom, right), is a strong reflector at 7 s TWT (two-way traveltime) below the Dinkelberg block, which is interpreted as the top of the reflective lower crust. However, only in the easternmost part of the profile the thickness and the depth position of this reflective zone is comparable to those observed on the KTB-8401 line (Fig. 6 top). The 20-km depth of this high-amplitude event on the western part of the profile, which is only a prominent contrast for P-waves (Fig. 10, top: Z-component), coincides with an increase of the P-wave velocity from 6.3 km/s to 7.0 km/s at the same depth inferred from refraction data near the southern rift axis (Edel et al., 1975). Comparison of the polarity of this event with reflections from the top of the crystalline basement gives evidence for a negative reflection coefficient. The presence of fluids at this depth is suggested by two observations, namely by the negative polarity of this event and by the presence of a strong P-wave reflection in the absence of the corresponding S-wave reflections (Lüschen et al., 1993; Echtler et al., 1994; Lüschen, 1994). The observed apparent thickness reduction of the lower crust from 10–11 km to 5–6 km due to Cenozoic rifting is not compatible with the extension observed in the sediments of about 10% (Illies and Fuchs, 1974; Villemin et al., 1986; Eisbacher et al., 1989). This discrepancy is a common phenomenon in many rifts and had been explained by a destabilization of the Moho by melts, by lower crustal delamination or by a local decoupling of deformation with depth (Ziegler, 1992b,c; Echtler et al., 1994).

Earthquakes beneath the Dinkelberg are concentrated near the top of the laminated lower crust at depths of about 20 km (Faber et al., 1994). This depth is therefore assumed to correspond to the boundary between the brittle and ductile regimes. Unfortunately, this is only half the findings. The subsequent N–S profile revealed a different story.

5.2. N–S reflection survey on eastern rift shoulder

Reflection seismic line DJ-91, which extends from the Badenweiler–Lenzkirch Zone (BLZ) in the north to the Swiss Jura in the south, is given in Fig. 11 as a migrated single-fold time section with colour-coded amplitudes and with projected hypocentres. This line stays within the zone of deep hypocentres (Fig. 9,

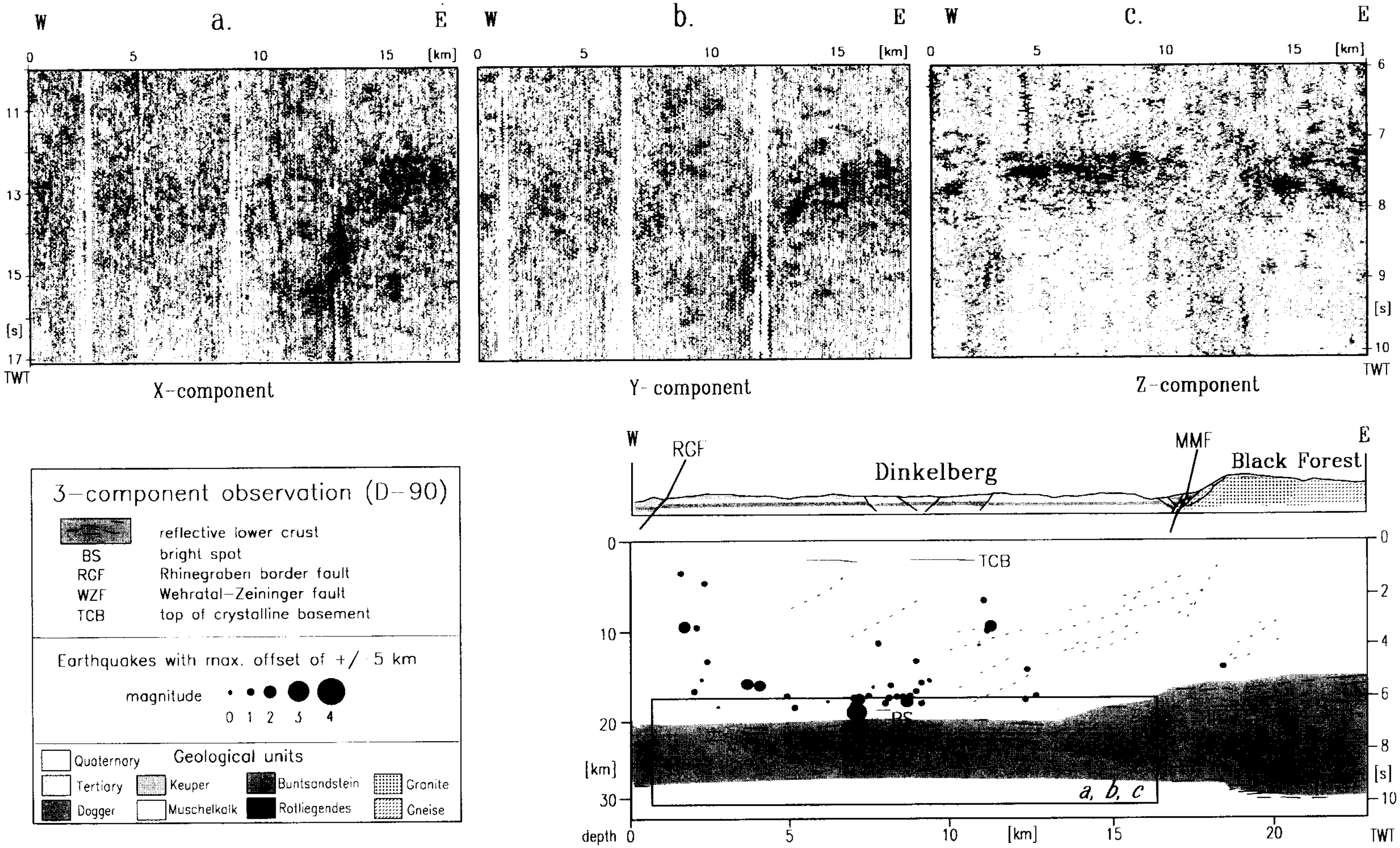

Fig. 10. Three component E–W profile D-90 in the Dinkelberg (see interpretative line drawing, bottom right). It provides additional information on the nature of lower crustal reflectivity especially on the detected bright spot at a depth of about 20 km. A zoom of the S-wave (top left: horizontal component E–W, top centre: horizontal component N–S) and P-wave section (top right: vertical component) in the time range of the prominent reflector shows a negative polarity of the reflections in the vertical component. Time scales of horizontal components (a and b) are stretched by a factor of 1.73 in order to compensate for the mean crustal V_p/V_s ratio. Note that the energy of S-wave reflections is increasing from west to east. While the earthquakes in the Black Forest are evenly distributed over a depth range of 2–22 km, they are concentrated beneath the Dinkelberg in a narrow seismogenic layer right above the top of the reflective lower crust (Echtler et al., 1994; Bonjer, 1997).

green zone) from the BLZ to the Dinkelberg and runs close to the western rim of the Black Forest. The reflectivity of the lower crust and its lateral variations is clearly visible. In the Dinkelberg block, both on the E–W and N–S line, the reflective lower crust is thinned and the deepest hypocentres just touch its top. However, on the N–S line, north of the Dinkelberg block the deepest hypocentres are located within the reflective lower crust.

On closer inspection, the N–S reflection line contains a remarkable modulation of the lower crustal reflection pattern. A band of enhanced reflectivity shifts in two steps from the top of the lower crust in the north to its bottom in the south. Beneath the Swiss Jura (left side of Fig. 11) the dominant reflections are located close to the crust–mantle boundary at a depth of about 28 km. The level of maximum reflectivity rises abruptly by nearly 8 km across the transform zone marking the southern boundary of the Rhine Graben (RSTZ) and remains at this level in the Dinkelberg block. Here the most prominent reflections occurs at the top of the thinned lower crust and only a very weak response from the Moho is obtained. Across the Kandern–Hausen fault, which marks the northern boundary of the Dinkelberg block, further change in the reflectivity and depth position of the lower crust is observed. The high-amplitude reflection bundle of the lower crust disappears and gives way to more uniform distribution of lower crustal reflections in the Black Forest block.

On the entire section the deepest hypocentres occur above the strong band of high reflectivity in the lower crust. Some 10 km to the east on the earlier discussed profile KTB-8401 the band characteristic is not observed.

5.3. Southern Rhine Graben traverse

The question arises whether the thinned or modified lower crust, characterized by narrow bands of enhanced reflectivity is a common feature to rifted areas or whether it is restricted to a narrow zone controlled by faults of crustal dimension. Indications for the presence of a thinned lower crust beneath the graben have already been described by Demnati and Dohr (1965), Dohr (1967), Edel et al. (1975) and Brun et al. (1992) (Fig. 3b and Fig. 4).

Unfortunately, areas covered by dense DSS lines are characterized by weak and only shallow seismicity. Therefore, a special survey was carried out to investigate the zone of prominent and deep seismicity of the Southern Rhine Graben (see Fig. 9 and Fig. 12). Receiver lines and shotpoints were located on both rift flanks (Vosges, Black Forest) in an effort to assume that ray paths of wide-angle reflections from the lower crust would travel only through crystalline crust.

The compiled E–W-striking cross-section, given in Fig. 12, shows the top of the lower crust at a nearly constant depth of 16–17 km increasing only locally near the eastern border fault to 20 km. The thickness of the reflective lower crust increases from about 9 km in the southern Black Forest to 11 km beneath the Vosges and thins under the eastern part of the Graben to as little as 6 km.

The crust–mantle boundary is subhorizontal and increases in depth from 26 km beneath the Graben to 28 km beneath the Vosges. The deep seismicity is limited to a narrow zone along the eastern margin of the Graben with hypocentres located near the top of the lower crust.

In summary, the investigations of structures as depicted by P-wave velocities and local seismicity suggested that the combined occurrence of thinned lower crust and deep crustal hypocentres could only be observed in a narrow, 10–25 km wide strip along the eastern border fault and in the Dinkelberg block.

It is remarkable that in this narrow zone the thinned strongly reflective lower crust is located always immediately beneath the deepest hypocentres. The position where thinned and modified lower crust had been observed is controlled by reactivated Variscan faults of crustal dimension (Echtler and Chauvet, 1992). However, deep hypocentres occur also in the adjacent Black Forest block where the lower crust is not thinned and exhibits with evenly distributed reflections throughout a some 10 km thick layer (see Figs. 11 and 12).

6. Images of recent tectonic activity — future investigations

In this section the hypothesis is advanced that the observed deep crustal seismicity reaching almost down to the Moho and its asymmetric concentration

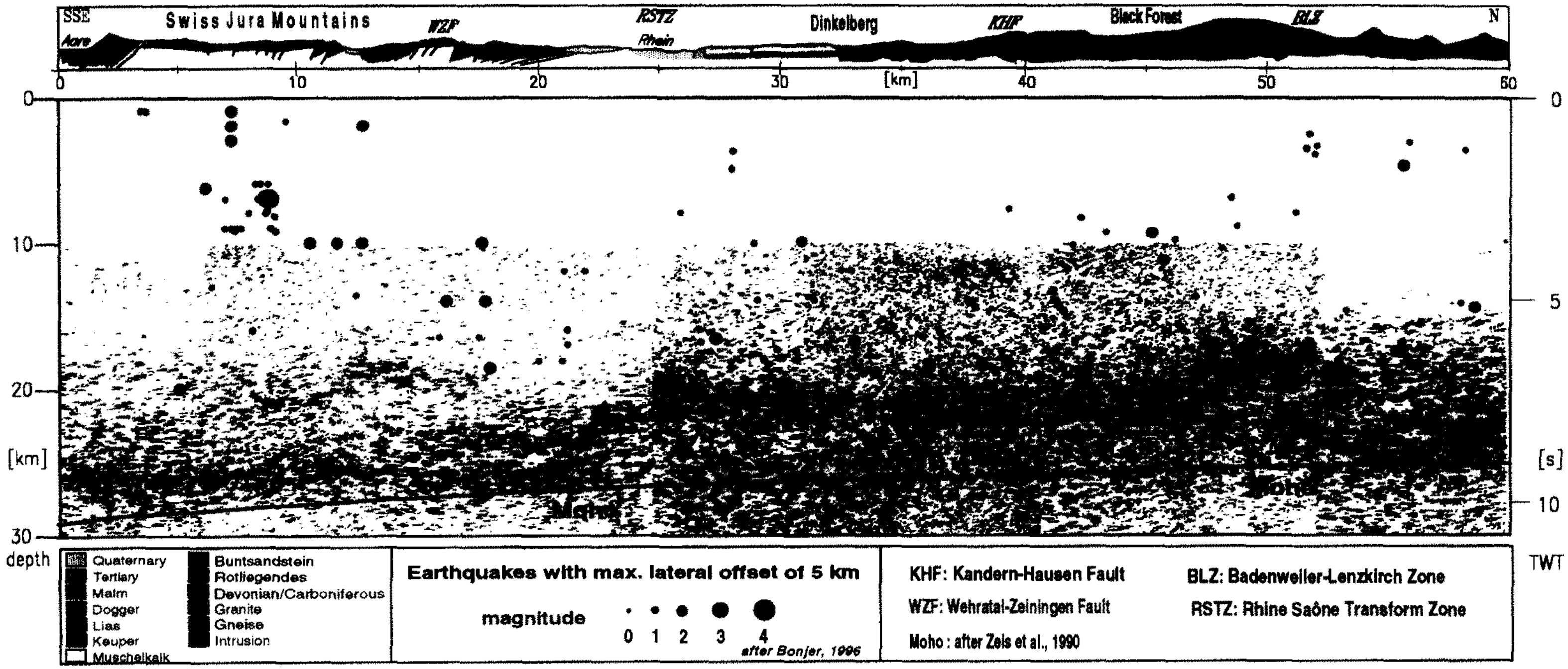

Fig. 11. N–S-striking profile DJ-91 time-migrated, single-fold section with colour-coded amplitudes including hypocentres with maximum lateral offset of 5 km to the seismic line. The character of the reflectivity pattern of the lower crust varies and its top decreases in steps from 15 km in the north to 21 km in the south. Beneath the Swiss Jura the reflections are concentrated in bands with maximum reflectivity near the Moho, whereas the dominant reflections from beneath the Dinkelberg are situated at the top of the lower crust. Beneath the Black Forest the reflectivity pattern is more diffuse and does not show any band characteristics. Considering the hypocentres, the earthquake foci are located well above the top of the lower crust. However, in the Black Forest north of the Dinkelberg, where the reflective band is merging into a homogeneous reflectivity pattern, they even occur within the reflective lower crust (Bonjer, 1997). Note that the strong step in the reflectivity band occurs where the profile cuts the Rhine–Saône Transform Zone (RSTZ).

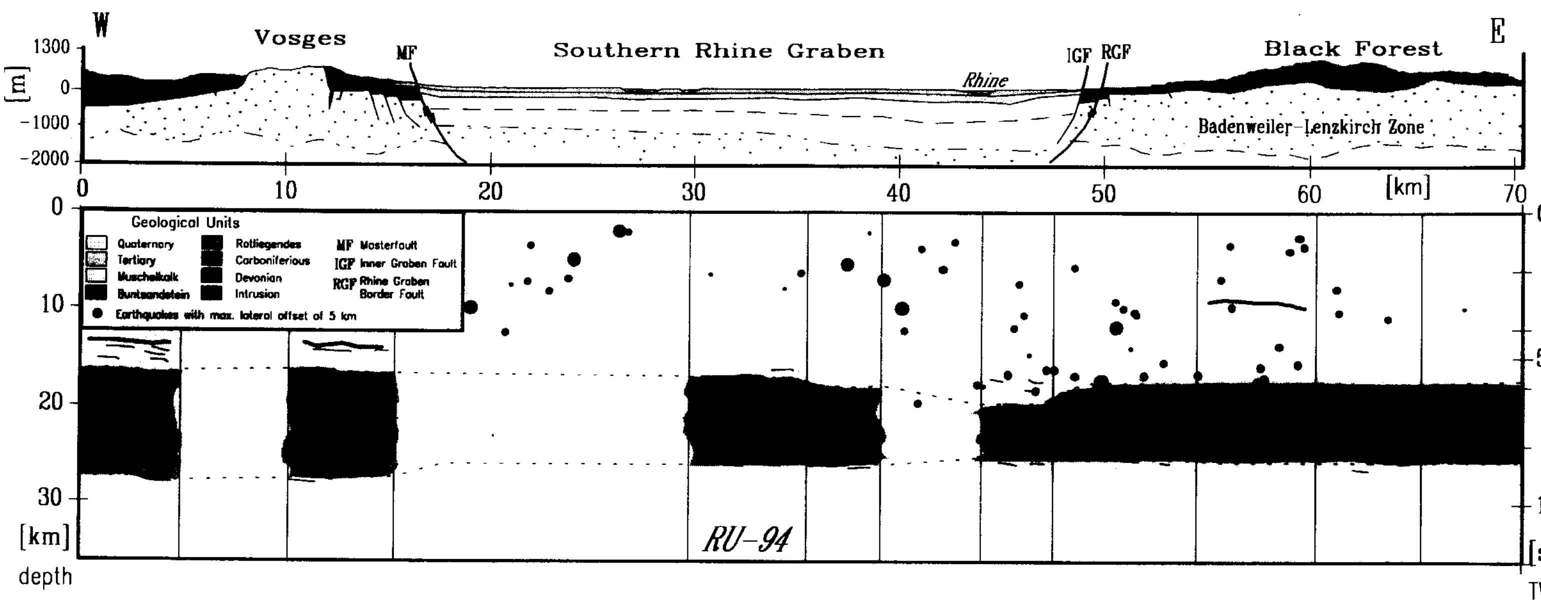

Fig. 12. E–W-striking cross-section RU-94, compiled from single-shot observations (see Fig. 9); the top of the lower crust is observed at an almost constant depth of 17 km (5.7–6.0 s TWT). There is unexpectedly no significant difference in thickness and depth of the reflective lower crust between the Black Forest and the Rhinegraben. Beneath the Inner Graben Fault (*IGF*) a local increase in upper crustal thickness up to 19 km (6.5–6.9 s TWT) is observed. The Moho is found at a depth of 27 km (9 s TWT), slightly dipping towards the west beneath the Vosges. The maximum depth of hypocentres, projected into the linedrawing, increases from west to east. The deepest hypocentres are observed close to the Rhinegraben Border Fault (*RGF*) and are situated occasionally above a thinned lower crust and partially inside a non-modified lower crust.

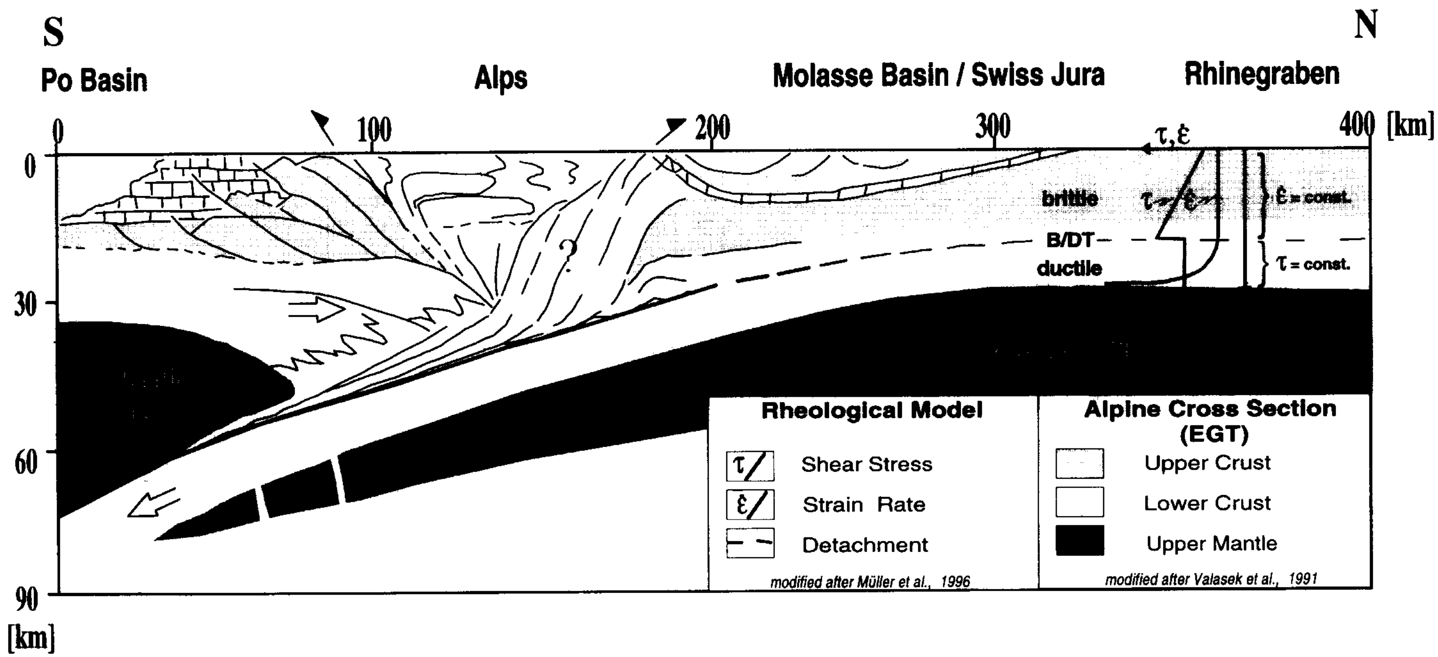

Fig. 13. Schematic lithospheric cross-section of the Alpine collision and subduction based on EGT studies (modified after Valasek et al., 1991). This section is based on an integration of various surface geological information and the combination of NFP-20 and EGT reflection seismic data. Note that a part of the lower crust is subducted together with the uppermost mantle and the detachment occurs somewhere beneath the foreland. The possible rheological behaviour in different stages of the upper lithosphere in case of detachment and depth dependent strain rate is indicated by the conceptional diagram for the horizontal shear stress τ (Müller et al., 1997).

at the southeastern margin of the Rhine Graben is an effect of the position of the latter relative to the Alpine orogen.

First, we will summarise the observations from various fields which encourage us to take certain patterns at the southern end of the Rhine Graben as traces of present tectonic activity.

Thinning of the lower crust along the eastern border fault system of the Southern Rhine Graben is regarded as a first-order structural effect of crustal extension. A connection to 'present-day deformation' is obvious since the deepest hypocentres stay above and reach the top of the thinned lower crust. Even where the hypocentres enter the reflective lower crust, they remain above the band of strongest reflectivity. Although we do not know what is controlling brittle deformation of the crust down to 5 km above the Moho, we realise that the ductile lower crust can have a thickness of only 5 km or less. Tentatively we take the band of strong reflectivity as marker of straining processes in the presently thin, ductile part of the crust.

An other first-order tectonic process that must be considered in the neotectonic evolution of the Southern Rhine Graben is the apparently continued subduction of European continental lithosphere in the Alpine orogen, as imaged by the results of the EGT and the NFP-20 (Fig. 13; Pfiffner et al., 1988, 1990; Schmid et al., 1989; Freeman et al., 1990; Valasek et al., 1991; Blundell et al., 1992; Gutscher, 1995; Ziegler, 1996a). Plate kinematic models, present stress trajectories and the pattern of earthquakes and neotectonic deformations indicate that the Alpine orogen is presently still active. Oligocene and younger post-collisional overthickening of the Alpine orogenic wedge was accompanied by imbrication of the European crust, accounting for some 50 km of crustal shortening since Late Oligocene times, and the subduction of a commensurate amount of lower crustal and mantle-lithospheric material. The geometry of the Aar Massif indicates that its development involved the activation of a mid-crustal detachment horizon (Ziegler, 1996a).

Dynamics of the latest Cretaceous and Cenozoic intra-plate compressional deformations in the European Alpine foreland, involving the inversion of Mesozoic tensional hanging-wall basins and the upthrusting of basement blocks along pre-existing crustal discontinuities, are related to collisional coupling between the Alpine orogen and its foreland. Such deformations can involve either the entire lithosphere or are restricted to crustal levels. In the latter case, a decoupling between the brittle crust and the mantle lithosphere must be invoked (Ziegler, 1995; Ziegler et al., 1996).

In this context it must be noted that at least the late deformation phases of the Jura fold-and-thrust belt may have involved the activation of intra-crustal detachment level (Ziegler, 1996a). This concept is compatible with the distribution of earthquakes and their focal mechanisms in the Swiss Molasse Basin (Deichmann, 1992a). Under such a scenario it cannot be excluded that a lower crustal detachment horizon extends also northwards from the Jura Mountains into the southern part of the Rhine Graben where it currently is activated under the prevailing stress regime.

Continued NW–SE-directed convergence of the Alpine orogen with its foreland probably causes reactivation of the eastern border fault system of the Southern Rhine Graben by sinistral shear, while north of the Badenweiler–Lenzkirch zone the central Black Forest and the Vosges are under NE–SW extension. There is also the possibility that the Vosges block is decoupled along the fault systems of the Rhine Graben from the Black Forest and that it is presently being translated sinistrally along the Burgundy transfer zone linking the southern end of the Rhine Graben with the Bresse Graben (Laubscher, 1970; Illies, 1977; Angelier and Bergerat, 1983; Eisbacher et al., 1989; Echtler and Chauvet, 1992).

The classical notion of the lithosphere under constant strain rates does no longer apply if the upper crust is separated from the mantle-lithosphere by an active lower crustal detachment zone (Kohlstedt et al., 1995) which accommodates differential movements between the upper crust and the mantle (Müller et al., 1997). Strain rates increase towards such a detachment horizon and reach their maximum within it. This implies that the strongest strain energy is released close to such a lower crustal detachment level. Moreover, high strain rates at such a detachment level may not be fully compensated by ductile creep, resulting in episodic brittle failure and thus in earthquake activity.

Although dynamic modelling of intra-plate compressional deformation of a rift zone located in the foreland of an active orogen under the impact of collisional related stresses is only at the beginning, we anticipate that stresses are responsible for the asymmetric distribution of earthquakes and the unusual depths of hypocentres observed in the Southern Rhine Graben. For more realistic modelling we have to take into account not only a decoupling of the crust from the mantle-lithosphere, but also a subdivision of the crust into blocks which are also decoupled from each other along deep crustal faults.

7. Conclusions

In this paper we have jointly interpreted high-density deep seismic sounding and seismicity data observed in the Rhine Graben region, especially at its southern end. These investigations, together with structural geological surveys, allowed us to draw conclusions on deep tectonic processes in the crust and mantle-lithosphere.

The Southern Rhine Graben is a natural laboratory for the study of deep tectonic processes. The Cenozoic Rhine Graben evolved in a crustal segment which was consolidated during the Variscan orogeny. Furthermore the Rhine Graben is located in the northern foreland of the Alpine orogen. At its southern end rifting and subduction tectonics are interfering. The main question in this paper was to what extent we could identify and scrutinise the various signatures of the different tectonic events by an integrated approach of investigations located in the northern foreland of the Alps on top of the subducting lithosphere.

We started by testing two different notions of crustal subdivisions based on the velocity structure and the strength of the crust. Comparing the two notions it was shown that an evolution of the crust has to be considered. The present ductile part of the crust is not necessarily identical with the ductile regime under which the widespread reflectivity of the Variscan crust was generated. The direct observational basis is the occurrence of deep hypocentres which reach far into the lower crust almost reaching the crust–mantle boundary.

Special care was taken to obtain DSS sections directly in the region of deep hypocentres along the Graben margin in the southern Black Forest. Seismicity in the southern part of the Rhine Graben is strongly asymmetrical. The largest number of earthquakes and the deepest hypocentres occur beneath the southern Black Forest. In the Dinkelberg area, an intermediate fault block between the southern Black Forest and the Rhine Graben, DSS data show that the reflective lower crust is thinned to 5 km. Its top is the focus of the deep hypocentres; no hypocentres occur within this thinned lower crust which rests directly on the Moho. In the adjacent southern Black Forest, the reflective lower crust has a thickness of 10 km; it contains a band of increased reflectivity which is located at greater depths in the south than in the north. This band correlates with the top of the reflective lower crust of the Dinkelberg block. In the southern Black Forest, deep hypocentres occur within the reflective lower crust but not below the highly reflecting band.

Since the hypocentres stay above the band of enhanced reflectivity, we interpret this band, as marking the brittle–ductile transition zone. Its thickness varies between 1 and 3 km.

The asymmetry of the deep hypocentre seismicity, the thinning of the lower crust and the deepening of the reflective band are presumably the effect of the activation of an intra-crustal detachment horizon in response to the built-up of intra-plate compressional stresses related to collisional coupling of the Alpine orogen with its foreland. While the Black Forest is mechanically coupled with the Alpine orogen, the Vosges block is decoupled along the eastern border fault zone and the RSTZ transform zone and moves in a southwestern direction.

Acknowledgements

The present investigation was sponsored by the German Research Society (DFG) within the framework of the Collaborative Research Center (SFB-108) 'Stress and Stress Release in the Lithosphere' at the University of Karlsruhe. We thank all participants in the recent reflection seismic field experiments which were carried out in 1990, 1991 and 1994 for their personal engagement and effort. The following institutions participated in the seismic field surveys: Universities of Clausthal–Zellerfeld, Freiberg, Karlsruhe, Kiel, Münster, München and Zürich and the Niedersächsisches Landesamt für Bodenforschung and the GeoForschungs-Zentrum (GFZ) in Potsdam. A cooperation with the Eidgenösische Technische

Hochschule (ETH), Zürich in 1991 made it possible to extend measurements in the framework of the NFP 20 into Switzerland. The very valuable reviews by P. Ziegler, E. Gurria and A. Becker are highly appreciated. Institute Contribution Nr. 720. SFB 108 Contribution Nr. 561.

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