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Seismic lamination and anisotropy of the Lower Continental Crust

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Abstract

Seismic lamination in the lower crust associated with marked anisotropy has been observed at various locations. Three of these locations were investigated by specially designed experiments in the near vertical and in the wide-angle range, that is the Urach and the Black Forrest area, both belonging to the Moldanubian, a collapsed Variscan terrane in southern Germany, and in the Donbas Basin, a rift inside the East European (Ukrainian) craton. In these three cases, a firm relationship between lower crust seismic lamination and anisotropy is found. There are more cases of lower-crustal lamination and anisotropy, e.g. from the Basin and Range province (western US) and from central Tibet, not revealed by seismic wide-angle measurements, but by teleseismic receiver function studies with a P–S conversion at the Moho. Other cases of lamination and anisotropy are from exhumed lower crustal rocks in Calabria (southern Italy), and Val Sesia and Val Strona (Ivrea area, Northern Italy). We demonstrate that rocks in the lower continental crust, apart from differing in composition, differ from the upper mantle both in terms of seismic lamination (observed in the near-vertical range) and in the type of anisotropy. Compared to upper mantle rocks exhibiting mainly orthorhombic symmetry, the symmetry of the rocks constituting the lower crust is either axial or orthorhombic and basically a result of preferred crystallographic orientation of major minerals (biotite, muscovite, hornblende). We argue that the generation of seismic lamination and anisotropy in the lower crust is a consequence of the same tectonic process, that is, ductile deformation in a warm and low-viscosity lower crust. This process takes place preferably in areas of extension. Heterogeneous rock units are formed that are generally felsic in composition, but that contain intercalations of mafic intrusions. The latter have acted as heat sources and provide the necessary seismic impedance contrasts. The observed seismic anisotropy is attributed to lattice preferred orientation (LPO) of major minerals, in particular of mica and hornblende, but also of olivine. A transversely isotropic symmetry system, such as expected for sub-horizontal layering, is found in only half of the field studies. Azimuthal anisotropy is encountered in the rest of the cases. This indicates differences in the horizontal components of tectonic strain, which finally give rise to differences in the evolution of the rock fabric.

Keywords: Reflection seismics; Seismic lamination and anisotropy; Lower continental crust; Crust–mantle boundary

1. Introduction

One of the most spectacular appearances of seismic reflections is the densely laminated reflectivity in the lower crust that is widely observed in the shallow crust of Phanerozoic extensional areas (Sadowiak and Wefer, 1990; Mooney and Meissner, 1992; Rey, 1995; Meissner and Rabbel, 1999). This reflectivity has been detected along many marine profiles in western Europe around the British Isles during the reflection programs BIRPS, starting in 1981, (British Institutions Reflection Profiling Syndicate) (Matthews and the BIRPS Group, 1990; Klemperer and Hobbs, 1991 and references

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therein), DEKORP in the German Variscides, (Meissner and Bortfeld, 1990 and references therein) and ECORS in the Paris Basin (Bois et al., 1986, 1987; Pinet et al., 1987). Strong lower-crustal reflectivity was later also found in parts of North America, for instance in the Basin and Range province, (Allmendinger et al., 1986; McCarthy and Thompson, 1988; Cook et al., 1997; Peng and Humphreys, 1997), and even in the thick compressed belts of Tibet (Ross et al., 2004; Huang et al., 2000; Sherrington et al., 2004), which show some extensional components of stress.

Among the many explanations, including multiple scatterers or fluids, a general tectono-thermal origin under stress and strain in a warm, ductile environment seems to be the most appropriate explanation (Mooney and Meissner, 1992; Meissner and Rabbel, 1999). Stretching and flow processes of heterogeneities, including mafic intrusions, are supposed to form long, sub-horizontal and therefore reflecting thin layers ("lamellae"), producing marked contrasts in seismic impedance. Scatterers and/or the presence of fractal inhomogeneities may also form some near-horizontal images (Emmerich, 1992; Hollinger et al., 1994).

However, the observed (ordered) reflectivity with continuous reflector lengths of more than 10 km or the missing "diffraction tails" are hard to explain. The presence of large amounts of fluids has also been suggested (Matthews and the BIRPS Group, 1990), but this explanation seems doubtful when considering several near-vertical experiments with P- and S-waves that show a comparable reflection strength, at least in the Black Forrest (Lueschen, 1999; Holbrook et al., 1992).

We think that stretching and ductile flow are the most important prerequisites for the generation of lamination and LPO-related-anisotropy in which the fast axes of anisotropic minerals (e.g. hornblende, muscovite, biotite) are aligned in the flow direction (Ribe, 1989; Savage, 1999; Park and Levin, 2002). We speculate that both processes – the formation of seismic lamination and anisotropy – are linked together, as suggested already by Pohl et al. (1999) and Meissner and Rabbel (1999). Both phenomena – lamination and anisotropy – should be frozen-in together during cooling (and modified or destroyed by new tectonic stresses). We will concentrate on three locations where this hypothesis can be tested, that is, where seismic lamination was observed in the lower crust and where complementing wide-angle investigations of seismic anisotropy were performed: (1) the Urach area and (2) the Black Forrest in the Variscan internides, and (3) the Donets Basin within the East-European craton.

Furthermore, we will briefly discuss two more locations showing both lamination and anisotropy, but it does not exclusively originate in the lower crust. These are the Basin and Range province in the western US, (Allmendinger et al., 1986, 1987; McNamara and Owens, 1993; Howie et al., 1991), and Central Tibet where tectonic escape is observed and independent investigation of lamination and anisotropy were carried out (Ross et al., 2004; Huang et al., 2000; Sherrington et al., 2004), and where crustal anisotropy was revealed using receiver function methods (Sherrington et al., 2004). In three more cases lamination and anisotropy of a former lower crust is inferred from petrological studies of exhumed lower crustal rocks now exposed at the surface (Weiss et al., 1999; Pohl et al., 1999).

Finally, we will briefly mention the eastern Alps where lamination and anisotropy are observed but here an additional anisotropy of the upper crust, caused by fault zones or foliation, may also contribute to the crustal anisotropy.

2. Main characteristics of lower-crust lamination

The terms "lamination" or "lamellae" have been used since the beginning of the 1980s, when the first land-based, near-vertical reflection experiments in Europe were conducted across the geothermal anomaly near Urach in the Moldanubian terrane in southern Germany (Bartelsen et al., 1982). The reflection patterns from these experiments always show strong reflectivity of the lower crust. These patterns were also observed in large parts of western Europe. Fig. 1 presents a typical profile SW of England across an extensional Mesozoic basin showing pronounced lower-crustal lamination (Warner, 1991; Cheadle et al., 1987). This reflection pattern is completely different from those recorded earlier in the USA during the COCORP program, which had already started in 1975 (Oliver, 1986, 1990; Brown et al., 1986).

Lower-crustal lamination, compared to the significantly lower reflectivity in the crystalline upper crust and the upper mantle, is considered a special type of continental reflectivity that contrasts with other reflectivity types like diffuse, fault- or convergence-related reflectivity, or special highly reflective zones at the base of the crust (McGeary, 1987; Sadowiak and Wefer, 1990; Meissner and Brown, 1991). Lower-crustal lamination is certainly based on the tectonic-rheological development of special areas (Mooney and Meissner, 1992). It seems to be concentrated in extensional areas, specifically in areas where the most recent tectonic event involved extension. This explains the lack of

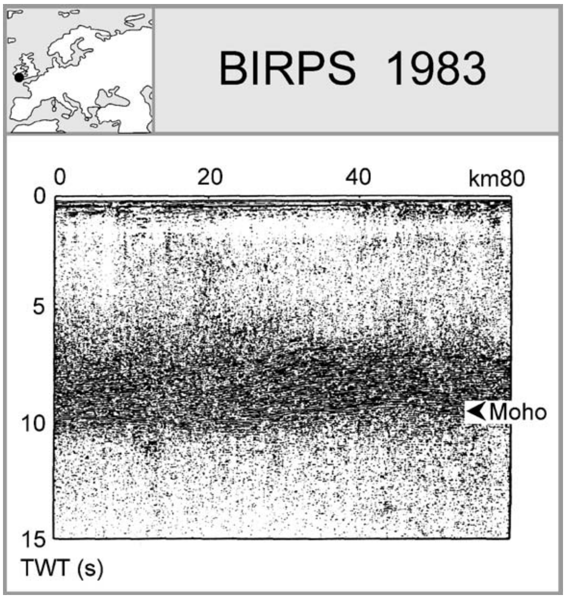

Fig. 1. Early example of lower crust lamination from the marine BIRPS survey, 1983. Mesozoic extensional basin southwest of England; (Warner, 1991; BIRPS and ECORS, 1989); TWT = Two-way-travel time in seconds.

lamination observed in the early COCORP data in the US which were obtained mostly in Proterozoic platform areas (Brown et al., 1986; Oliver, 1990). Later surveys in young extensional areas also showed the laminated reflectivity pattern of the lower crust (McCarthy and Thompson, 1988).

The lower-crustal lamination is characterized by a dense pattern of planar, mostly horizontal, reflectivity, terminating at the Moho, while the upper crust often, but not always, shows a "transparent" reflectivity pattern (McGeary, 1987). Lamination so far has not been observed in the mantle, because of its limited variation in mineral composition. Neither is it observed in the crystalline upper crust (Meissner and Rabbel, 1999). Although the dimension might differ from centimeters to more than 50 km — (from oriented minerals up to seismic layers within the lamination), the lengths of the seismic packages in which lamination and anisotropy are found are generally the same, i.e. 5 to 20 km. Single "lamellae" may have thicknesses of less than 300 m and some even 50 m, but still have a correlating length of up to 10 km and more (Bartelsen et al., 1982; Gajewski, 1987; Wenzel et al., 1987).

A note of caution seems to be appropriate regarding the term "lamination". Here, and in all near-vertical reflection surveys, lamination stands for the (sub-horizontal) appearance of lower crust reflectivity. However, some researchers apply it also to multiple wide-angle events of the P_MP phase (Jensen et al., 2002; Bayer et al., 2002). We have to consider that the near-vertical reflection coefficients represent simple or combined reflections of a real layering, the individual layers having possibly a thickness much lower than the (high frequency) seismic reflection wavelength. The near-horizontal and lower frequency waves of wide-angle reflections, on the other hand, may contain near-vertical gradient zones, and not necessarily real reflections. Another fundamental aspect of seismic lamination is the possible contribution of scatterers that may under certain conditions generate coherent energy visible over a few kilometers (Emmerich, 1992), but diffraction tails associated with these scatterers have not been observed. Various efforts have been made to model alternating velocity layers in the lower crust in order to match the observed lamination (Lueschen et al., 1987; Singh et al., 1998). Generally, modelling is based on the reflectivity method (Fuchs and Müller, 1971) and in one case (in the strongly laminated Moldanubian Black Forrest crust), anisotropic layering in the lower crust was modelled (Lueschen et al., 1990) in order to explain S-wave lamination from a near-vertical recording of signals from a shot. In the wide-angle area, such differences were observed in our Urach and Donbas experiments and even in the Black Forrest (Gajewski, 1987), but in the near-vertical area no large differences are expected from variations in mineral composition (see next paragraph and Discussion section).

Fig. 2 gives a summary of lower-crustal lamination that shows the crustal thickness values for which lamination typically occurs and the velocity associated with lamination, as obtained from additional wide angle studies of (near vertical) lamination. We see (left) that most lamination patterns are from rather thin (extensional) crusts with the thickness generally less than 30 km, and typically "felsic" velocities, i.e. velocities between 5.9 and 6.9 km/s (e.g. Kern, 1982, 1993). Even the anomalously thick crust of central Tibet (and the Alps) (right) shows these low velocities. A few areas have velocities slightly above 7.0 km/s, among them our study area in the Donbas Basin. All these areas with lower crust velocities higher than 7 km/s are from old (Proterozoic) shield and platform areas (middle).

In the following we summarize the most important observations on lower-crustal lamination:

1. The lamination observed world-wide is restricted to continental lower-crust, where there is nearly always a viscosity minimum between the rigid upper crust

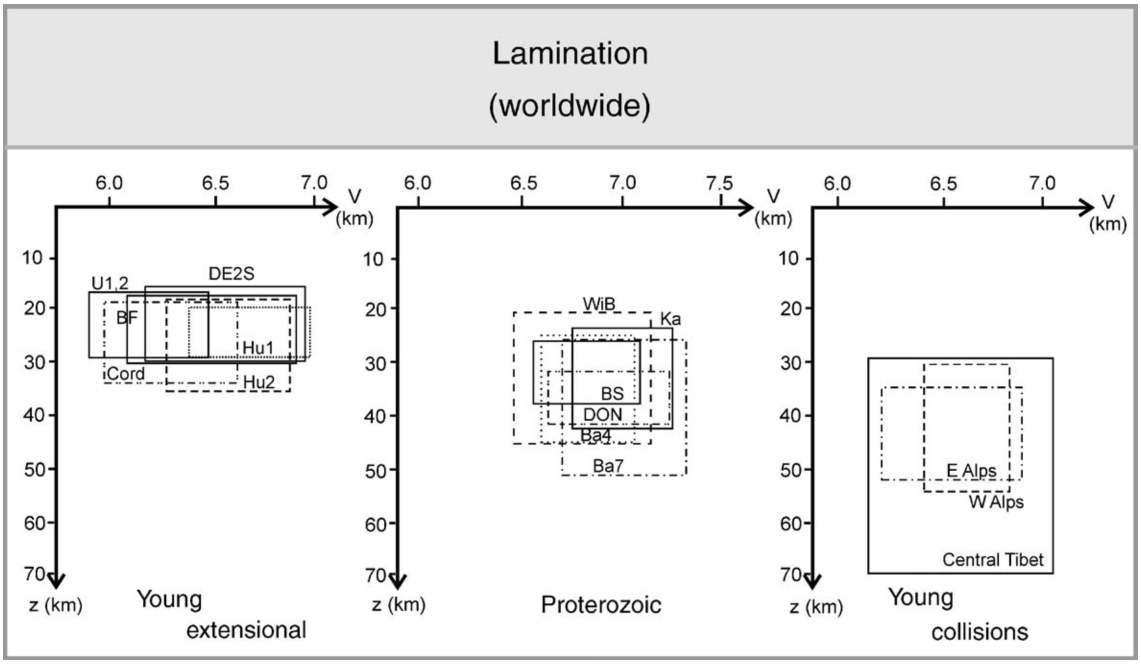

Fig. 2. Velocity — depth diagram and ranges of V_p in the laminated lower crust from areas where near-vertical and wide-angle observations were available; The three diagrams show young extensional, Proterozoic and young orogenic areas. Cord = Am. Cordilliera (Prussen, 1991)DE 2S = DEKORP 2 South, Saxothuringian and Moldanubian (Meissner and Bortfeld, 1990), Hu 1, Hu 2 = Hunsrück area (Glocke and Meissner, 1976; Meissner and Bortfeld, 1990), BF = Black Forrest area (Lueschen et al. 1987), U1, 2 = Urach 1 and 2 — lines (Bartelsen et al., 1982), WiB = Williston Basin (Prussen, 1991), Ka = Kapuskasing (Wu and Mereu, 1991), BS = Baltic Shield (EUROBRIDGE, 1999), DON = Mesozoic Donbas Rift inside craton (Maystrenko et al., 2003), BA 4, BA7 = various BABEL lines in the Fennoscandian Shield (BABEL, Working Group, 1993) WAlps = Western Alps (Heitzmann et al., 1991), EAlps = Eastern Alps (Lueschen and the TRANSALP Working Group, 2003) Central Tibet = Changtang Block (Ross et al., 2004; Meissner et al., 2004).

and the uppermost mantle (Sibson, 1982; Ranalli, 2000). Some contribution from scatterers cannot be ruled out (Emmerich, 1992).

2. Lamination is observed in many extensional areas throughout the world. We relate this to the initiation of creep, which needs 3 to 4 times lower stresses in extension than in compression at comparable temperatures (Meissner and Strehlau, 1982). Recently, seismic lamination in the lower crust has also been observed in the thick and warm crusts of compressional (possibly collapsing?) belts (Ross et al., 2004; Transalp Working Group, 2002).
3. Often, laminated lower crust is associated with a “transparent” upper crust. This might be an effect of near-vertical reflection methods, which, for geometrical reasons, cannot image (extensional) steep-angle faults (but see Meissner and Rabbel, 1999 for refined processing techniques).
4. Observed lamination appears very planar and is often sub-horizontal. Reflecting layers must be very planar and probably created in a ductile, creeping environment (Mooney and Meissner, 1992).
5. In the majority of cases, high-velocity mafic intrusions inside felsic material are suspected to generate considerable impedance contrasts.

Based on these observations we conclude the following:

The (thin) mafic intrusions that contributed heat during their emplacement, are presently the main contributors to the strong reflectivity. In addition, LPO-related seismic anisotropy can increase reflectivity (Kern and Wenk, 1991). The large amount of felsic material causes the low average velocities (Fig. 2). The few cases in which lamination with velocities greater than 7 km/s occur all are in Proterozoic areas. This is because these old shields and platforms – in contrast to young and extensional areas – generally have a thick, 3-layer crust with a high-velocity lower crust (Meissner, 1986; Guterch et al., 1999; Jensen et al., 2002).

Mafic intrusions into the lowermost crust and a gabbro–eclogite transition in thick crust might be responsible for a missing “strong” impedance contrast at the Moho (Mengel and Kern, 1992). Often the reflections are terminated by the "refraction" Moho (Meissner and Brown, 1991).

3. Seismic anisotropy in the lower crust and its relationship to the symmetry of the anisotropic rock fabric

Various kinds of anisotropy occur in the continental crust (Babuska and Cara, 1991; Vinnik et al., 1992; Savage, 1999). In the upper crust periodic layering (Helbig, 1984; Crampin, 1989) and aligned micro-cracks contribute to the observed anisotropy (Rabbel, 1992, 1994). Shear wave splitting observations (Crampin, 1981, 1984) give evidence that upper-crustal anisotropy is mainly caused by oriented cracks and micro-fractures. At pressures of 200 to 300 MPa, corresponding to about 10–15 km depth, much of the anisotropy disappears, primarily due to the closure of cracks. Therefore, any crustal anisotropy below these depths is probably caused by other phenomena, for example, by minerals aligned during ductile flow. This means that (crystallographic) lattice preferred orientation (LPO) is the main source for anisotropy in the lower crust and in the sub-crustal lithosphere, at least down to the Lehman Discontinuity at about 200 km depth (See Jones et al., 1999; Table 1 in Meissner et al., 2002, and related references therein).

It should be mentioned here that some papers, which concentrate on the elastic constants of the (anisotropic) stress- or strain tensor, assume axial symmetry with a slow or fast axis of symmetry: i.e. a vertical slow axis as in our crustal approach, and a fast axis mostly in the horizontal, that produces azimuthal variations (Peng and Humphreys, 1997; Levin and Park, 1998; Sherrington et al., 2004). Such an approach reduces the number of elastic parameters significantly and makes calculations simpler, but it does not agree in most cases with the fabric symmetry of the rocks constituting the lower crust and upper mantle (see Fig. 3). " Axial " in our sense always refers to the slow (near vertical) axis of lower crustal rocks.

The LPO patterns are generally interpreted as resulting from dislocation glide in the crystal lattices, sometimes accompanied by dynamic re-crystallization (Savage, 1999). The LPO of rock-forming minerals is generated in ductile layers within the lithosphere under appropriate stress systems, e.g. in the continental lower crust and in the mantle below a rigid mantle lid. Although some dynamic re-crystallization and annealing will occur during cooling, a significant part of the anisotropy may remain in the rock unit (so-called "fossil" anisotropy). Generally, the latest thermo-tectonic event leaves the strongest anisotropy effect in the lower crust or upper mantle.

A marked contribution from fabric-related seismic anisotropy has been documented in thin sections of the upper crust along the KTB deep borehole in southern Germany (Rabbel, 1994; Popp and Kern, 1994) and along the Kola superdeep borehole in northern Russia (Digranes et al., 1996; Kern et al., 2001). Large parts of the lower crust seem to be tectonically and rheologically similar, as indicated by the large zones of uniform reflection patterns, like for instance, the laminated types of the lower crust.

In ductilely deformed (foliated) lower crustal and upper mantle rocks, P- and S-wave propagation and S-wave polarization are strongly controlled by the crystallographic fabric (LPO) of the constituent minerals and the overall symmetry of the resulting strain-induced rock fabric (e. g. Kern, 1993). Fig. 3 schematically presents the LPO-based 3D-velocity distributions of relevant lower crustal and upper mantle rocks (mica gneiss, amphibolite, peridotite) and illustrates the directional dependence of P- and S-wave propagation and S-wave polarization. It is clear from the diagrams that the direction of wave propagation in relation to the preferred orientation of the major minerals, and to the symmetry of their lattice fabric, is of great importance. In the transversely isotropic mica gneiss and in the orthorhombic amphibolite, the directions of maximum shear-wave splitting coincide with the direction(s) of maximum P-wave velocity. This is in accordance with the maximum hornblende [001]-axes and (inferred) biotite [100]+[010]-axes, respectively, which are the fast directions in both minerals. This is not the case in the upper mantle peridotite exhibiting almost an overall orthorhombic symmetry. Parallel to the maximum concentration of the olivine [100]-axis, which is the fast direction in the olivine single crystal, there is no significant shear-wave splitting observed, (Kern, 1982, 1993). Here, the maximum shear-wave splitting is found to be subnormal to lineation (sub-parallel to [001]axes) within the foliation plane, and moderate but significant shear wave-splitting is observed subnormal to foliation (sub-parallel to [001]). In contrast, in the mica gneiss and the amphibolite, a second shear wave normal to foliation will practically not be generated. Typical values of LPO-related intrinsic P-wave anisotropies for different crustal rocks (quartz–micaschists; felsic gneisses, granulite-facies meta-pelites; amphibolites) range from 5.4% to 10.7% (Holbrook et al., 1992).

Potential upper mantle rocks in massifs and from xenoliths (peridotite; dunite) show P-wave anisotropies

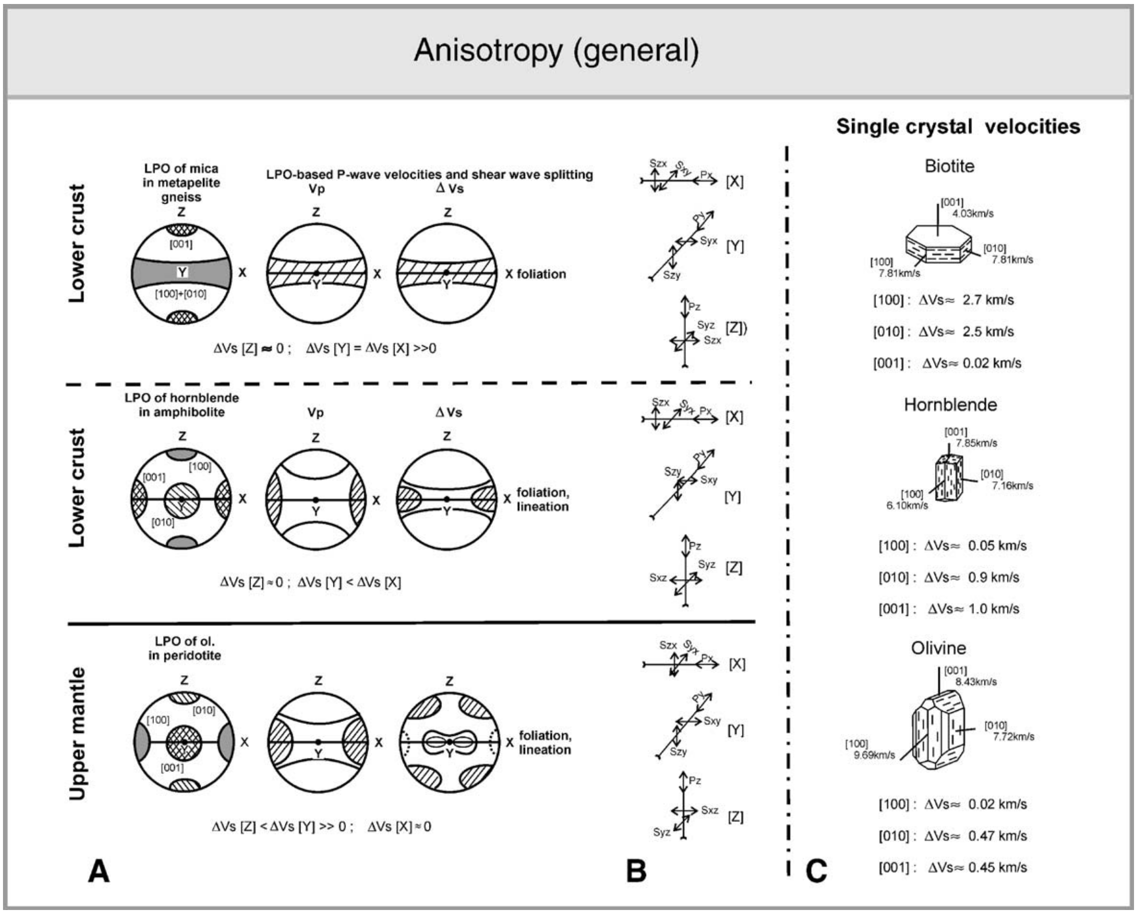

Fig. 3. Conceptual model for P- and S-wave propagation in foliated lower crust and upper mantle rocks, dominated by crystallographic preferred orientation (LPO) of biotite, hornblende and olivine, respectively. A) LPO and models of velocity surfaces (V_P and ΔV_S) in foliated lower-crustal and upper mantle rocks. B) Propagation and polarization of P- and S-waves with respect to the structural reference frame X, Y, Z. C) V_P and V_S of dominating single crystals.

varying between 7.7% to 8.3% (Babuska and Cara, 1991). Elastic anisotropy is mainly caused by the LPO of olivine. The presence of pyroxene (enstatite) in mantle peridotites produces a dilution of the seismic anisotropy (Mainprice and Silver, 1993). Importantly, in crustal rocks, the variability of the mean P-wave velocities due to variation in composition is generally higher than the variation caused by velocity anisotropy (except for schists). However, the reverse is true in the olivine-rich upper mantle rocks (Babuska and Cara, 1991). This probably implies that on a regional scale, fabric-related anisotropy is more important in the oceanic and continental upper mantle than in the continental crust.

Major contributions reporting crustal anisotropy (Rabbel and Lueschen, 1996; Gajewski, 1987;Weiss et al., 1999; Pohl et al., 1999; Godfrey et al., 2002) confirm an axial symmetry with the SH velocity faster then the SV velocity. In contrast, the Donbas Basin, the Basin and Range area and Central Tibet show azimuthal components (Next section). It is important to note that foliated lower crustal rocks exhibiting marked lineation are generally orthorhombic in symmetry. This holds, in particular, for hornblende-rich rocks such as biotite-honblende gneisses and amphibolites (Kern et al., 2001; Barruol and Kern, 1996; Siegesmund et al., 1989). On the other hand, the teleseismic receiver function method with a P–S conversion at the Moho shows two near-vertical S components in an anisotropic crust. Directional observations might reveal azimuthal anisotropy, if present, but from the near-vertical rays very little real axial symmetry with $V_{SH} > V_{SV}$ can be observed.

Horizontally to sub-horizontally layered crustal rocks exhibiting axial symmetry do not contribute much to the shear wave splitting observed by teleseismic SKS studies. However, as shown by Barruol and Mainprice (1993), specific geological structures, such as (near vertical) strike-slip faults that cut the whole crust may give a delay time of 0.3–0.5 s between the perpendicular polarized shear waves for a 10 km thick layer and thus contribute up to 30% of the anisotropy measured by SKS studies.

With regard to our new observations (next section), it should be mentioned here that at least these two kinds of anisotropy – axial and orthorhombic – are observed in areas of lower-crustal lamination.

In the following we summarize the most important observation on anisotropy:

1) Most (teleseismic) shear-wave splitting observations with near-vertical ray paths do not monitor crustal anisotropy and generally cannot distinguish between crustal and mantle anisotropy (Park and Levin, 2002).
2) Special wide-angle experiments with controlled sources and selected source-receiver patterns or the receiver function method with an P–S conversion at the Moho boundary (Sherrington et al., 2004), are needed to reveal anisotropy in the lower crust (Rabbel et al., 1998).
3) Lower-crustal anisotropy often seems to have an axial symmetry (Levin and Park, 1998), but azimuthal patterns are also found (Babuska and Cara, 1991; Kern et al., 2001). The receiver function method with near vertical ray path mainly reveals azimuthal crustal anisotropy.

Most researchers agree that an alignment of anisotropic minerals with the flow direction takes place under pure shear or progressive simple shear. Hence, we argue that the fast axes of anisotropic minerals (e.g. biotite, muscovite), and in some cases also hornblende minerals, will align within the foliation plane such that a transverse isotropy for a horizontal orientation of the foliation is produced (Barruol and Kern, 1996; Savage, 1999). Differential stress and strain causing an overall orthorhombic rock fabric (including near-vertical structural elements) might cause an azimuthal component in the lower crust.

A warm lower crust reflects a viscosity minimum (Sibson, 1982; Ranalli, 2000) and offers the possibility for creep and a uniform alignment of anisotropic minerals. Warm and ductile lower crust associated with intrusions of mafic magmas occurs in extensional areas (Klemperer and Hobbs, 1991; Meissner et al., 2002) and even in the overthickened crust of young mountain belts (sometimes with tectonic escape).

Some lamination and anisotropy might survive cooling, annealing and re-crystallization (Vinnik et al., 1992).

4. Observations of lower-crustal lamination and anisotropy

4.1. Near-vertical and wide-angle reflection studies

Areas of laminated and anisotropic lower crust revealed by extensive near-vertical and wide angle seismic experiments include the Variscan Internides, the Urach geothermal anomaly and the Black Forest, as well as in the Donbas Foldbelt.

4.1.1. The Urach geothermal anomaly and the Black Forrest area

The seismic investigations in 1979 across the Urach geothermal anomaly, southern Germany, are one of the earliest near-vertical reflection studies in Europe. They show a lower crust dominated by reflecting lamellae with a low (felsic) average velocity below a rather transparent upper crust. The whole area belongs to the Moldanubian terrane in the Variscan internides which is considered to have suffered from the collapse of the Variscan mountain belts and shows a laminated lower crust everywhere. It is observed on large N–S profiles, for instance DEKORP 2S, and along all N–S and E–W profiles in the west (i.e. in the Black Forrest area: Lueschen et al., 1987; Gajewski, 1987) and in the south below the Alps where Moldanubian crust is transported down into the Alpine root zone (Lueschen and the TRANSALP Working Group, 2003).

During the collapse of the Variscan belts, a delamination of the original overthickened lower crust and the upper mantle must have taken place (Franke and Oncken, 1990; Downes, 1993; Onken, 1998). One consequence of this process is the production of large syn- and post-orogenic granitic bodies in the Variscan internides and total reworking of the crust (Onken, 1998). It is important to note that the original 50 to 70 km thick crust of the Variscan mountain belts was transformed into a thin crust of less than 30 km. During and after the collapse, the original compressive stress systems changed into transtension and extension. The lower crust became thin, felsic, mobile and laminated (see also Fig. 2, U1). Since the Variscan period of collapse and extension around 300 Ma ago, no strong tectonic phase has effected the area, except for some uplift in connection with the Alpine orogeny (with its maximum around 60 Myr ago, or the uplift of the flanks of the Rhinegraben in Tertiary times (Ziegler, 1986).

Fig. 4A shows line drawings of line Urach 1 and Fig. 4B the Black Forrest line 8403, a western DEKORP extension of the U1 line.

A special study of the laminated lower crust focusing on anisotropy was carried out in the Urach area in 1994 (Rabbel and Lueschen, 1996; Rabbel et al., 1998). In addition to the reflection studies from 1979, special wide-angle experiments were performed along two perpendicular profiles. Along both lines, expanding spread profiling was performed, the common midpoints of which centered on an area of pronounced lamination. Signals along various ray paths, especially S_MS-waves were recorded by three-component seismometers. These waves travel a considerable part of their ray path in the lower crust. Along both profiles, S-waves of the SH type traveled several hundreds of milliseconds faster at distances of 30 to 90 km. No clear azimuthal anisotropy could be detected. The coincidence of SH-velocities in both directions and the differences in SH and SV velocity indicate an overall axial symmetry (transversal isot-

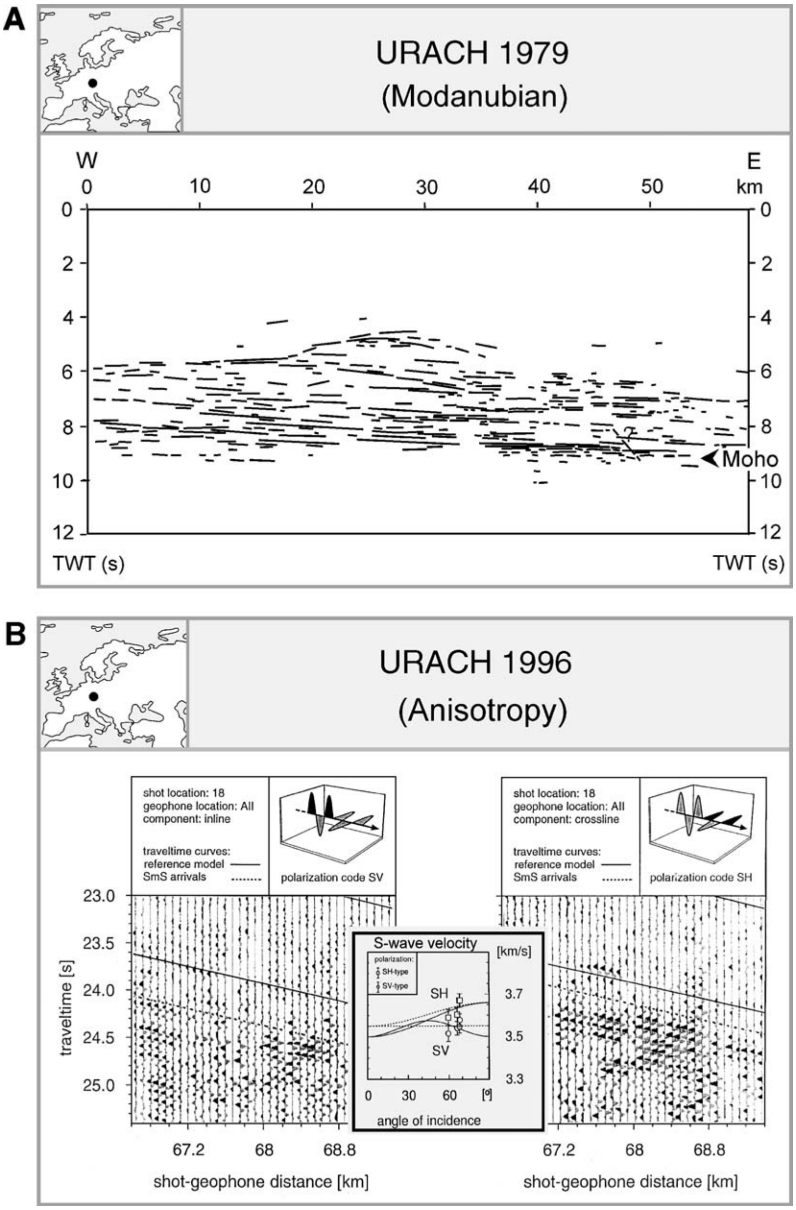

Fig. 4. (A) Manual line drawings of Urach 1979 (Bartelsen et al., 1982) and its western extension (Black Forrest line 8403, 1984) (Lueschen et al., 1987). Insert shows location of the study area. TWT = two-way travel time in seconds; M = Moho. (B) S-wave anisotropy in Urach: SH — waves arriving 250 ms earlier than SV waves. Solid line: reference model; stippled line: S_MS arrivals; left picture SV, right picture SH.

ropy), an observation which agrees with many laboratory studies on foliated lower crustal rocks like metapelites and gneisses (Kern, 1982; Barruol and Kern, 1996; Weiss et al., 1999). Fig. 4B shows the different travel times of the recordings along profile U1. SV waves travel considerably slower than SH, which is in accordance with the theory for hexagonal structures (Fig. 3). They arrive nearly 250 ms later than SH waves.

Several types of anisotropy were investigated, especially periodic layering of thin (isotropic) layers with contrasting velocities that is most commonly found in sediments (Crampin, 1989; Meissner et al., 2002). This kind of anisotropy was also initially assumed for the lower crust because of the dense and strong seismic lamination (Rabbel and Lueschen, 1996). The calculations with maximum contrasting velocities, however, could not explain the observed large differences between the SV and SH velocities.

The tectonic development and situation of the Black Forest area is similar to that of the nearby Urach area: The Black Forrest is also part of the Moldanubian terrane where extension dominated during the collapse of the Variscan mountain belt. During the DEKORP programme, the Black Forest was the focus of much attention because it was one of the selected deep drilling sites. Hence, many near-vertical reflection (Fig. 4B) and some wide-angle lines were observed, one of them even connecting to the Urach area (Lueschen et al., 1987, 1990; Meissner and Bortfeld, 1990). The lower crust in the whole Black Forrest area shows a classical lamination with (rather low) P-wave velocities of 6.2 to 6.8 km/s (Fig. 2). In the wide-angle area a marked anisotropy was observed (Gajewski, 1987) with travel-time differences between radial (SV) and tangential (SH) components of more than 200 ms at distances of 60 to 120 km. An axial anisotropy with no azimuthal contributions, similar to the Urach observations, is observed. Also in the Black Forrest, a strong correlation between seismic lamination and anisotropy of the lower crust is present.

4.1.2. The Donbas Foldbelt

A combined seismic reflection program, called DOBRE, was carried out in 2000 and 2001 by an international group of researchers across the Donbas Foldbelt in the Ukrainian craton, an uplifted and mildly deformed segment of a Late Devonian–Carboniferous intracratonic rift. The southern margin of the Donbas Foldbelt was uplifted in Early Permian in a trans-tensional regime. Inverted fault systems were created up to the Late Cretaceous (Maystrenko et al., 2003; Stephenson, 2003). The seismic studies in the Donbas Basin were much more detailed than those of the Urach and Black Forest investigations. As shown in the map of Fig. 5A, near-vertical reflection seismic experiments were carried out along the lines A and B and wide angle seismic profiles were arranged so that seismic rays from shots along line A traveled to three-component seismometers along line B, thereby providing a wide range of azimuths with a nearly common midpoint in the lower crust below the Donbas Basin. Hence, in addition to the near-vertical investigations, these extensive wide angle surveys provided different azimuths. These observations formed the backbone of anisotropy studies. Details about these studies are in discussed Rabbel et al. (2003).

From the near-vertical studies, 20 km thick sediments in the centre of the rift and a high-velocity lower crust (7.1 to 7.2 km/s) with seismic lamination are observed (Fig. 5B). As is evident from Fig. 2, the Donbas foldbelt is one of the few areas where seismic lamination is observed together with high velocities. The high velocities in the lower crust are not unusual for shield areas, but lamination in an old shield area is not commonly observed (Fig. 2). It is likely that the high-velocity lower crust was already generated during the Devonian rifting period by a massive emplacement of mafic magmas at the base of the crust. This emplacement was a consequence of extension and isostatic processes similar to those found in various basins all over the world (Matthews and the BIRPS Group, 1990). Extensional creep at that time most probably also created the seismic lamination connected to the extensional rift process of the (intracratonic) basin which later cooled and solidified.

Based on 11 wide-angle seismic recordings, an anisotropy for the S_MS wave is derived from the different travel times recorded at the individual horizontal (and vertical) seismometers. These recordings show a complex azimuthal anisotropy where the maxima of shear-wave splitting are arranged obliquely, but symmetrically, to the axis of the Donbas Basin (see Fig. 5C). This shows that the tectonic structure apparently determines the system of anisotropy. The special azimuthal function of shear-wave splitting and structural modelling favour the presence of a system of horizontal and vertical structures (Rabbel et al., 2003). According to our preferred model, ultramafic magma has intruded the felsic or intermediate lower crust along some near-vertical feeder dikes (Fig. 5D) and then spread sub-horizontally in a manner similar to the processes that generate horizontal sills. An additional contribution from massive, sub-horizontally aligned mafic-ultramafic minerals is possible. In summary, the complex structure of the lower crust and the high (mafic–ultramafic) velocities, which are probably due to (aligned?) orthopyroxene or olivine minerals, are suggested to be responsible for the strong, azimuthal anisotropy observed in the lower crust of the Donbas Basin.

4.2. Seismic reflection combined with receiver function studies

In this section we briefly summarize studies in the Basin and Range province and in Central Tibet, followed by a short description of the TRANSALP transect (eastern Alps).

4.3. The basin and range province

The main extension in the Basin and Range province (Western US) is in the NW–SE direction. Many seismic investigations have been performed, among them some COCORP lines in the north (Allmendinger et al., 1987; Howie et al., 1991)

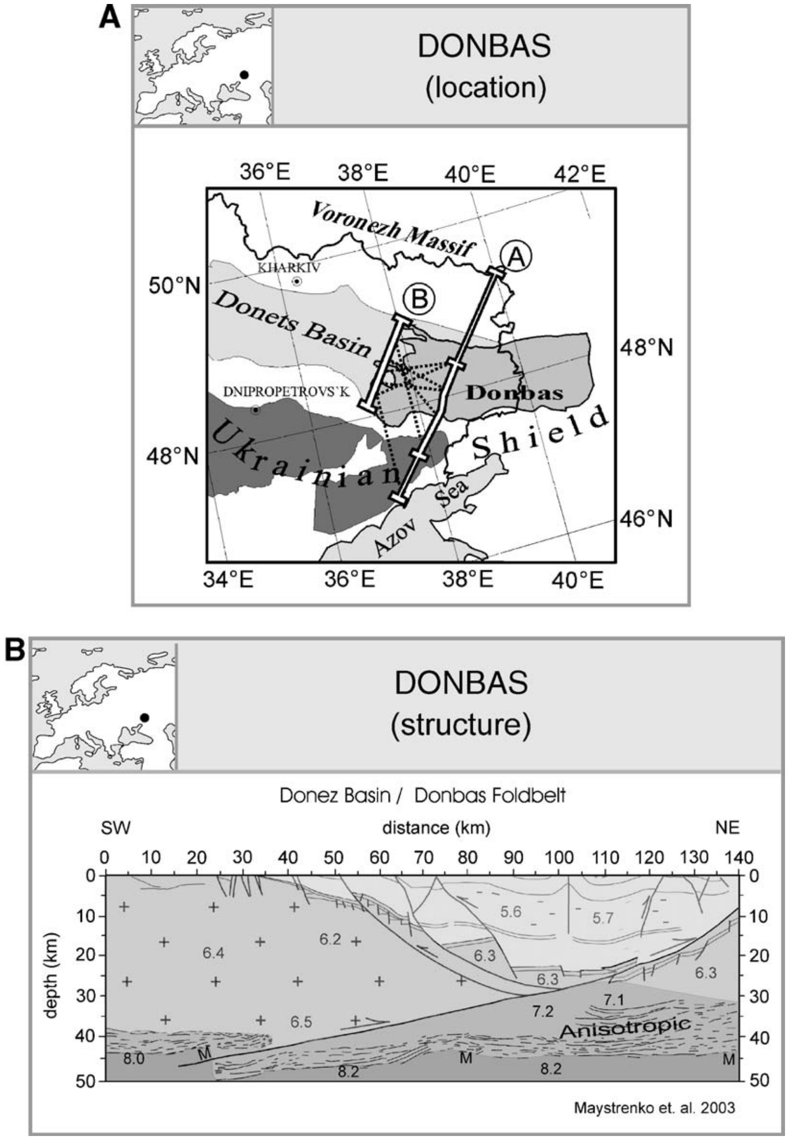

Fig. 5. (A) Situation map and acquisition schedule of wide angle measurements in the Donbas area, signals generated on line A and recorded on line B. Simplified geology. (B) DOBReflection and velocity structure of the Donbas area with seismic lamination in the lower crust. (C) Time difference Δt between split S-waves as a function of azimuth. Δt values are symmetric to the basin axis and show a complex azimuthal anisotropy (see text). (D) Conceptual (feeder dyke) models of mafic/ultramafic intrusions (black lines), explaining the observed (two-period) S-wave anisotropy. Our data fit into the left part (0–180° of the lowermost picture.

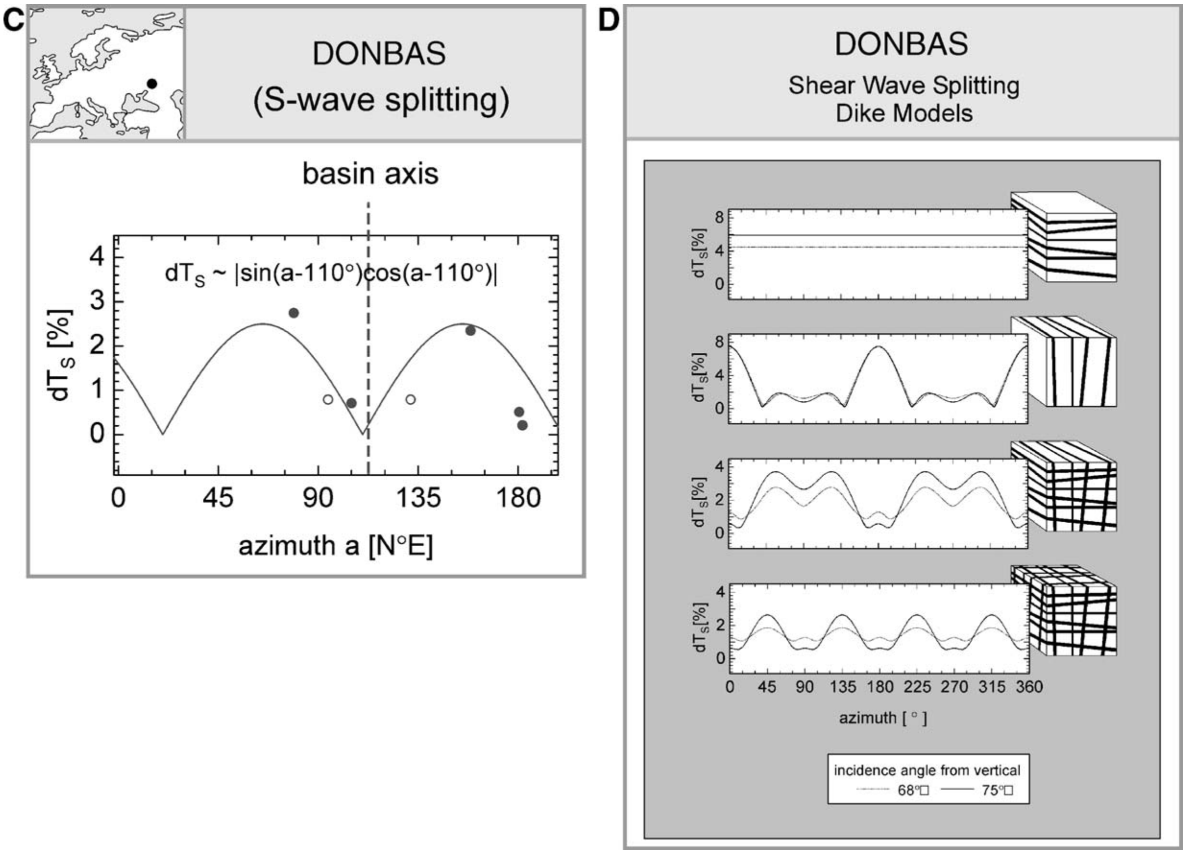

Fig. 5 (*continued*).

and some nearby PASSCAL teleseismic receiver function studies (McNamara and Owens, 1993; Peng and Humphreys, 1997). While Allmendinger et al. (1987) and McCarthy and Thompson (1988) concentrate on the seismic lamination and its evolution and temperature, Carbonell and Smithson (1991a,b) suspect already a connection between crustal reflectivity and P-anisotropy. McNamara and Owens (1993 and Peng and Humphreys (1997) apply the receiver function method using more than 200 teleseismic events and analyze Ps phases converted at the Moho. The steep converted S phases reveal an azimuthal anisotropy with a delay of more than 200 ms and the fast axes oriented NW–SE, i.e. in the direction of maximum Cenozoic extension.

An axial component in addition to the azimuthal anisotropy is possibly also present because SH waves are generally a little faster than SV-waves, but proof of this anisotropy is impossible using the receiver function method. Aligned low-velocity minerals like quartz and mica have to be supposed for the lower crust because of the rather low P-wave velocities. In addition to the asymmetry of stress and strain, some vertical faults or feeder dykes, like in the Donbas Basin, cannot be ruled out. We consider the Basin and Range province to be a prominent example for the presence of seismic lamination and anisotropy.

4.4. Central Tibet

During the long period of convergence between India and Asia, the Tibetan plateau has developed the highest elevation and the thickest crust on Earth (Tapponier et al., 1986; Bird, 1991). Many N–S directed extensional faults, W–E directed strike-slip faults and recent GPS surveys (Beghoul et al., 1993; Van der Voerd et al., 2002; Wang et al., 2001) prove the tectonic escape of central Tibet, especially that of the central part north of the Bangong–Nuijang Suture (BNS). We believe that the eastward directed (extensional) escape movement takes place on top of the warm and ductile lower crust, which based on travel time interpretation of shots along the INDEPTH III profile and some nearby earthquakes (Zhao et al., 2001; Meissner et al., 2004), has rater low P-wave velocities (6.0 to 6.6 km/s).

Five near-vertical reflection tests were carried out near the BNS, adjacent to the INDEPTH III profile (Ross et al., 2004). In contrast to the reflection experiments in southern Tibet, these new shots showed a good penetration and reveal a lamination between 10 and 20 s TWT (Two-way travel time), equivalent to 30 to 70 km depth, i.e. the whole lower crust in Tibet. The upper crust and the upper mantle are transparent. Never before has such a deep and strong lamination been observed (see Fig. 2). The presence of lamination agrees with our postulation that hot, ductile and mobile lower crust is escaping to the east. Teleseismic studies indicate pronounced anisotropy along the INDEPTH III line (Huang et al., 2000). The sudden onset of anisotropy slightly south of the BNS clearly shows that the crust makes a strong contribution to anisotropy, while the large amplitudes argue for a major contribution from the mantle.

In the eastern part of central Tibet, special receiver function studies with a P–S conversion were carried out (Sherrington et al., 2004). Apparently, there are several systems of anisotropy inside the (thick) crust. The lower crust in the central part shows anisotropy with a clear W–E trend of the fast axes. It is consistent with the general escape direction, and the possibility that lower crust and upper mantle are deforming coherently (Sherrington et al., 2004). We refer again to the different horizontal stress and strain, possibly supported by vertical intrusions in the form of feeder dikes, as an explanation for the azimuthal anisotropy of the thick laminated lower crust of central Tibet.

4.5. Eastern Alps

Indications for the connection of lamination and anisotropy are also provided from TRANSALP, an international and interdisciplinary study along a large transect from Munich (southern Germany) to Venice (Italy) through the eastern Alps (TRANSALP Working Group, 2002; Lueschen and the TRANSALP Working Group, 2003).There is a strongly laminated lower crust, 5–6 km thick, similar to the well-known Moldanubian crust in southern Germany, as mentioned before.

Many teleseismic and wide-angle shear-wave splitting measurements show some crustal anisotropy, but this anisotropy is not restricted to the laminated lower crust. Fast axes are oriented W–E in the crust and mantle, consistent with the easterly tectonic escape of the eastern Alps (Ratschbacher et al., 1989).

4.6. Studies of exhumed lower crustal rocks in Calabria and Ivrea

Exposed lower crustal rock units with a periodic, quasi laminated appearance in Calabria (Southern Italy), Val Sesia and Val Strona (Ivrea body, northern Italy) have been investigated in situ and in various laboratories. These studies give information on the mineral content and the anisotropy of rocks constituting the lower crust. From the various samples of exposed lower crustal sections, Weiss et al. (1999) investigate 8 different crustal rock types including gneisses, granulites and metapelites. For Calabria and Val Strona (Ivrea), the average symmetry of the 3D velocity distributions is axial (transversal isotropy) with the fast axes parallel to the foliation. In contrast, in Val Sesia (Ivrea) the mafic rocks (metagabbro) exhibit almost orthorhombic symmetry (azimuthal anisotropy). Here, the symmetry is orthorhombic in P but more complex in S. In order to investigate how rock heterogeneity and anisotropy transform to a seismic response, Pohl et al. (1999) applied the anisotropic reflectivity method to the data from Calabria and Ivrea provided by Weiss et al. (1999). The resulting synthetic seismograms verified that both the Calabria and Ivrea sequences can be regarded as laminated lower crust in terms of seismic reflectivity. In terms of bulk anisotropy, they indicated a transversely isotropic behaviour for Calabria and Val Strona, while for Val Sesia the rock pattern is found to be more complex and only weakly anisotropic. Based on their data, the authors created some spectacular models both for crustal anisotropy and seismic lamination and speculate already about connections. These studies were important forerunners for our ideas and experiments. A synthesis of our field and lab studies is found in the Appendix and its two tables.

5. Discussion

Observations of seismic lamination in extensional or collapsed areas and theoretical arguments pointing to a lower-crustal viscosity minimum, give evidence that ductile flow in the lower continental crust is responsible for the planar or even sub-horizontal layering (Meissner and Brown, 1991; Smithson et al., 1987; McKenzie and Jackson, 2002). Mafic magmas from the mantle that intruded into the lower crust are responsible for heating and layering, and ductile deformation is considered to be responsible for the alignment of anisotropic, rock-forming minerals causing transverse isotropy or azimuthal anisotropy.

The main reason for the occurrence of lamination in extensional or collapsed areas is assumed to lie in the asymmetry of the relevant stress fields. The early experiments of Byerlee (1970) have shown that extensional stresses 3–4 times lower than compressive stresses can initiate frictional shear and creep. Hence, under a limited (?) tectonic stress, there will be much more extensional than compressive deformation (Fig. 2).

When lamination was first documented by characteristic reflectivity patterns in western Europe in the beginning of the 80 s, it was thought that this kind of reflectivity would be evident globally when using appropriate techniques. Today, we estimate that globally only about 15% to 20% of seismic lines show this reflectivity pattern, but we can distinguish special tectonic areas where lamination dominates. First, there is the (extensional) collapsed Moldanubian terrane, especially the Variscan Internides in central Europe, where nearly all profiles (close to 100%) show lamination (Meissner and Bortfeld, 1990). Heat and ductility from delamination might have supported mobility in this area. In contrast, the northern Varisciscan externides, in the Rhenish Massif, only about 200 km north of the Moldanubian, show a completely different reflection pattern (Glocke and Meissner, 1976; Meissner and Rabbel, 1999). Also, further north in the North German Basin, there is a different type of reflectivity.

Lamination is found in about 50% of the BIRPS profiles around the British isles that cross many Mesozoic and Cenozoic basins with Moho uplift and strong lamination, especially in the west and southwest (McGeary, 1987; Matthews and the BIRPS Group, 1990). Only about 10–15% lamination is found in Proterozoic areas in Baltica (The BABEL Project reports, 1992, 1996; EUROBRIDGE Working Group, 1999) or in the US and Canada (Wu and Mereu, 1991; Cook et al., 1987, 1997; Vadesuvan and Cook, 1998).

The voluminous data from the Canadian Lithoprobe project, running mainly through Archean and Proterozoic cratons, show generally a highly reflective crust (like some of the BABEL data from Scandinavia, BABEL Working Group, 1993), but only sporadically crustal lamination with reflector-length of more than 5 km. Also, the lower crust of Australia does not show wide-spread lamination (Wright et al., 1987). This lack of lamination in old shield areas agrees with observations of Miller (2003). Old shields and cratons generally have a thick three-layered crust, the lower crust showing velocities higher than 7 km/s (Fowler, 1990; Guterch et al., 1999; EUROBRIDGE Seismic Working Group, 1999). They generally do not show lamination except where mobile belts or rift system, like the Donbas basin (see Section 3) are present.

Another problem with seismic lamination is the fact that near-vertical ray paths can only see sub-horizontal or slightly dipping reflectors, not steep angle faults or intrusions. Such near-vertical feeder dykes must be present in order to transport magmas upward (invisible in the reflection picture) to allow later formation of (thin) sub-horizontal lamellae (Klemperer and Hobbs, 1991; Rey, 1995; McKenzie and Jackson, 2002). Inside the laminated lower crust, velocity differences should exceed a value of 0.1 for the reflection coefficient (Warner, 1991; Mooney and Meissner, 1992). This is easily achieved in (extensional) felsic and even in mafic lower crusts (e.g. Kern, 1982, 1993; Holbrook et al., 1992). We realize that there are many examples of lower-crustal lamination worldwide, mostly in extensional or collapsed areas, but fewer examples for crustal anisotropy are available. We know that lower crust anisotropy is difficult to observe, and most reflection surveys do not incorporate additional and costly wide angle or receiver function measurements. For example, the early COCORP land profiles in the US and the early marine BIRPS profiles around the British Isles did not contain any additional measurements.

Regarding anisotropy we find a clear transverse symmetry with SH waves faster than SV-waves in the laminated lower crusts of the Urach and the Black Forrest area, in accordance with the assumption of a horizontal foliation and low velocities of crustal rocks that are dominated by quartz, biotite or muscovite. We find strong evidence for azimuthal anisotropy in the high-velocity (mafic) lower crust in the Donbas Basin. Here, the most extensive investigation of crustal anisotropy has been carried out. We also find azimuthal components in the Basin and Range area, in Central Tibet and in the eastern Alps where the laminated lower crust always has a low velocity.

We offer three explanations for the azimuthal anisotropy of the lower crust. The first is connected with vertical structures either from vertical shear zones or vertical intrusion (Barruol and Mainprice, 1993). In these near-vertical zones, the foliation is near-vertical (not sub-horizontal as in the visible lamination). This means that we may have vertically aligned anisotropic minerals with hexagonal symmetry that produce (horizontal) azimuthal components for near-vertical ray paths (See again Fig. 3). The horizontal lamination has to be fed by some near-vertical intrusions (feeder dikes). We recall that for geometric reasons no vertical structure can be observed by (near vertical) reflection seismic studies.

The second explanation is based on the fact that we know that thin mafic intrusions are necessary to explain the high impedance contrasts of seismic lamination. If there are some aligned mafic–ultramafic minerals, like hornblende or olivine, in the near-horizontal thin layers, they produce an orthorhombic symmetry and thus cause azimuthal anisotropy.

Our last explanation is based on a more general tectonic reasoning and may dominate and explain the other two possibilities. This is the presence of differences in (horizontal) tectonic stress and strain that is evident from many tectonic structures. These differences lead not only to directional creep with oriented mineral fast axes but also to the opening of cracks, feeder dykes and intrusions. See Tables in Appendix A.

We have shown that in our investigated areas seismic lamination and anisotropy coincide, their joint presence providing an additional argument for our concept of lamination (true layering) as opposed to hypotheses incorporating scatterers or fluids. We postulate that creep will produce both lamination and anisotropy. Tectonics and its structural geometry, plus mineral composition and mineral alignment, will determine whether lamination occurs and the type of anisotropy that results. Some important questions remain. Will azimuthal and hexagonal effects anywhere appear together and is there somewhere anisotropy without lamination or lamination without anisotropy?

6. Conclusion

To date, lamination in the lower crust has been observed worldwide and much more frequently than anisotropy. About 15% to 20% of near-vertical seismic surveys show lower-crustal lamination. Seismic S-wave wide-angle studies and/or teleseismic receiver function studies in areas of (near vertical) seismic lamination seem to be promising tools for observing crustal anisotropy. Seismic lamination and anisotropy in the continental lower crust are certainly generated under very similar tectonic-rheological conditions, i.e. ordering and alignment processes in a high temperature, ductile and mobile environment, preferably under tensional stress. Hence, lamination and anisotropy

Table 1
Seismic anisotropy of reflective lower crust (general schedule)

Rock type	**Vp [km/s]**	**Strain as a proxy of tectonic stress**				**Symmetry system of seismic anisotropy**			
		Principal strains	Maximum compressive strain	Rock fabric		Isotropic	Hexagonal		Ortho-rhombic
				Foliation	Lineation		Transvers. isotropic	Azimuthal	
Felsic ⇔ biotite gneiss	6.0-6.4	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	⊗			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		⊗		
			Horizontal	+	-			⊗	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+		⊗		
			Horizontal	+	+			⊗	
Inter-mediate ⇔ hornblende gneiss	6.4-6.9	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	⊗			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		⊗		
			Horizontal	+	-			⊗	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+				⊗
			Horizontal	+	+				⊗
Mafic and ultramafic ⇔ gabbroic + peridotitic gneiss	6.9-7.5	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	⊗			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		⊗		
			Horizontal	+	-			⊗	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+				⊗
			Horizontal	+	+				⊗

White background: verified in laboratory experiments; grey background: not verified in laboratory experiments. "⊗" shows most probable symmetry systems of seismic anisotropy related to rock type, P-wave velocity and strain regime.

should both be generated and observed together. At all sites discussed here, seismic lamination and anisotropy coincide, thus supporting our hypothesis of a joint generation, although differences are observed in average P-wave velocity and in the type of anisotropy. Whereas in the Variscan Internides we observe (horizontally) transverse isotropy, there is a dominance of azimuthal anisotropy in the Donbas Basin, in the Basin and Range province and in Central Tibet. Studies in the Donbas area, where the most extensive investigations of lower-crustal anisotropy have been carried out, suggest the presence of vertical pathways comprising mafic minerals and an anisotropy dependence of tectonic structures. Such vertical pathways that are not visible in laminated lower crust, together with a complex stress-strain system, seem to be the most probable explanations for the observed azimuthal anisotropy. The main mineral composition is perhaps of secondary importance. Our results show that joint investigations of seismic reflectivity and anisotropy

Table 2
Field studies of anisotropic reflective lower crust

Rock type	**Vp [km/s]**	**Strain as a proxy of tectonic stress**				**Symmetry system of seismic anisotropy**			
		Principal strains	Maximum compressive strain	Rock fabric		Isotropic	Hexagonal		Ortho-rhombic
				Foliation	Lineation		Transvers. isotropic	Azimuthal	
Felsic ⇔ biotite gneiss	6.0-6.4	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	⊗			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		Urach, Black Forest Calabria Val Strona		
			Horizontal	+	-			⊗	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+		Urach, Black Forest, Calabria, Val Strona		
			Horizontal	+	+			⊗	
Inter-mediate ⇔ hornblende gneiss	6.4-6.9	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	⊗			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		⊗		
			Horizontal	+	-			Tibet (?), Basin and Range (?)	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+				⊗
			Horizontal	+	+				⊗
Mafic and ultramafic ⇔ gabbroic + peridotitic gneiss	6.9-7.5	$\varepsilon_1=\varepsilon_2=\varepsilon_3$	-	-	-	Val Sesia I			
		$\varepsilon_1>\varepsilon_2=\varepsilon_3$	Vertical	+	-		⊗		
			Horizontal	+	-			⊗	
		$\varepsilon_1>\varepsilon_2>\varepsilon_3$	Vertical	+	+				Donbas, Val Sesia II
			Horizontal	+	+				

White background: verified in seismic field experiments; grey background: not verified in seismic field experiments. "⊗" shows most probable symmetry systems of seismic anisotropy related to rock type, P-wave velocity and strain regime (cf. Table 1).

have a considerable potential for identifying deep crustal processes.

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Appendix A

We summarize the seismic, petrological and petrophysical laboratory measurements and define a schedule for reflective (laminated) lower crust relating the seismic observables (P-velocity and anisotropy) to the rock candidates and the strain. The state of strain, considered to be a proxy of stress, can be described by magnitude and spatial orientation of the principle components in the geological units (ε_1, ε_2, ε_3). See Table 1.

It is petrologically expressed by the presence or absence of foliation and lineation, marked by "+" or "–". Also, the symmetry system of anisotropy for the special combination of rock types and strains are indicated. As shown by the laboratory measurements, felsic and intermediate rocks are expected to show hexagonal symmetry, its orientation depending on the strain orientation. Orthorhombic symmetry is expected if the lower crust has significant mafic or ultramafic component or if different tectonic strain was active (last two lines of Table 1). The table may also be used to attribute rock composition and strain conditions in situ from seismic field measurements.

As an example, we applied the interpretation schedule of Table 1 to the field data from the different areas discussed in the main text by putting the names of the investigation areas in the respective boxes combining observed P-wave velocity and anisotropy system (Table 2). Table 2 shows that Urach, Black Forest, Calabria and Val Strona fall into the same category of lower crust characterized by felsic material. Because of its transverse isotropy, maximum compressive strain is oriented vertically, and differences in horizontal strains were small at the time when the reflective lower crust formed. Tibet and the Basin and Range form another group to which an intermediate composition has to be attributed and where, in contrast to the previous group, differential horizontal strains have played a crucial role. This is indicated by the azimuthal component of anisotropy. The presence of mafic and/or ultramafic material is required to explain the coincidence of orthorhombic behaviour and high P-wave velocity in a third group represented by the Donbas and Val Sesia II data. Val Sesia I is the only example of reflective lower crust showing no anisotropy.

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