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Seismic anisotropy of the subcrustal lithosphere as evidence for dynamical processes in the upper mantle

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Summary. Anisotropy of seismic waves in the uppermost mantle has not only been observed in the oceanic but recently also in the continental lithosphere. Laboratory experiments on the formation of preferred orientation of olivine crystals suggest plastic flow as the most likely mechanism for the genesis of anisotropy in the upper mantle. Since the direction of maximum velocity correlates in the ocean and on the continent with a number of tectonic features, a causal connection between anisotropy and dynamical processes related to plate motions must be suspected.

1 Introduction

Until recently, anisotropy was absent in most seismic models of the Earth's interior. There was a reluctance by seismologists to include anisotropy in their models since theory of wave propagation in anisotropic media and the analysis of observations becomes immensely more complex than in isotropic media. Also the inclusion of anisotropy was not required by the observations since it did not improve the fit of observed and computed data. Although, of course, it was known that single crystals possess anisotropic elastic properties, the absence of anisotropy on a macroscopic scale could easily be explained by the assumption of random orientation of the single crystals during their formation resulting in an isotropic medium. This view could be taken as long as no processes inducing a preferred orientation were known and no seismic data required anisotropy.

Dynamical processes within the Earth's interior as claimed in the framework of the new global tectonics are possible sources for such a preferred orientation of anisotropic crystals on a large regional scale. If this would lead to an observable anisotropy and if the mechanism of its formation was known, the detection of seismic anisotropy could become an important key to unlock the dynamical processes of the Earth's interior.

It is the purpose of this paper to review briefly the explosion seismic experiments in the oceans and on the continents which show evidence of anisotropy in the upper mantle. Laboratory work on the formation of preferred orientation of anisotropic minerals, as well as correlation of the direction of anisotropy with other tectonic features suggest that anisotropy is induced by flow and stress fields in the upper mantle through preferred orientation of minerals.

2 Observations of seismic anisotropy in the upper mantle

2.1 ANISOTROPY IN THE OCEANIC UPPER MANTLE

Only a few years after formulating his sea-floor spreading hypothesis, Hess (1964) pointed out that the velocity of seismic waves in the uppermost mantle of the Pacific Ocean showed a clear dependence on the azimuth of propagation (Fig. 1). He visualized a connection between this kind of anisotropy and the spreading of the sea-floor since the direction of smallest velocity was perpendicular to the large fracture zones, i.e. parallel to the ridge axis. He based his assumption on the observation that olivine crystals tend to a preferred parallel orientation of their crystallographic major axis and supposed that such an alignment could take place by large-scale flow induced by shearing stresses. In particular, the *b*-axis, i.e. the direction of slowest velocity tends to align perpendicular to the shear planes since the (010)-plane is also the plane of best cleavage (Fig. 2).

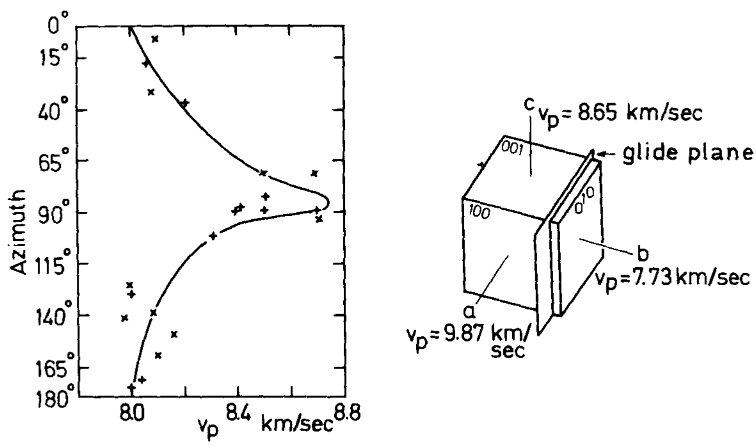

Figure 1. Anisotropy of *P*-wave velocities in the upper mantle of the Pacific Ocean and its explanation by a preferred orientation of olivine (after Hess 1964).

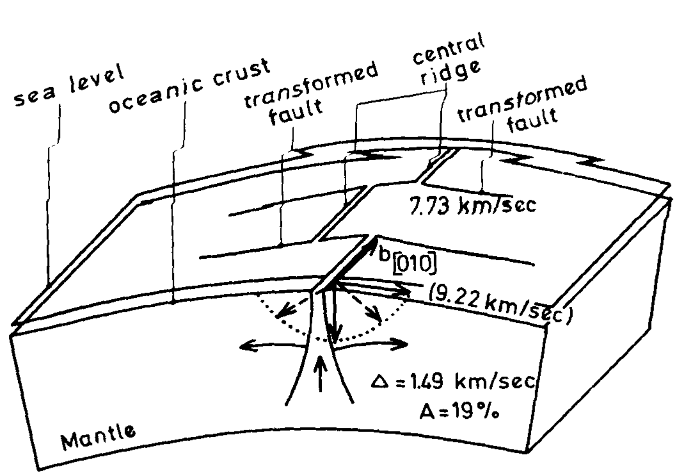

Figure 2. Orientation of olivine axis in relation to the ridge axis and to the transformed faults (after Hess 1964).

Systematic explosion–seismic anisotropy experiments were conducted subsequentially in the Pacific Ocean to exclude the possibility that Hess' observation of azimuthal dependence of velocities was erroneously deduced from lateral variations of velocities only (Raitt *et al.* 1969, 1971; Morris, Raitt & Shor 1969; Keen & Barrett 1971). Backus (1965) has shown that for a weak anisotropy (<10 per cent) the azimuthal dependence of the seismic *P*-velocity V_p may be described by

$$V_p(\phi) = \bar{v} + a\cos 2\phi + b\sin 2\phi + c\cos 4\phi + d\sin 4\phi.$$

$\bar{v}$ is the average velocity. c and d are in general much smaller than a and b. Fig. 3 is an example from an area near Hawaii (Morris *et al.* 1969) where this azimuthal dependence can

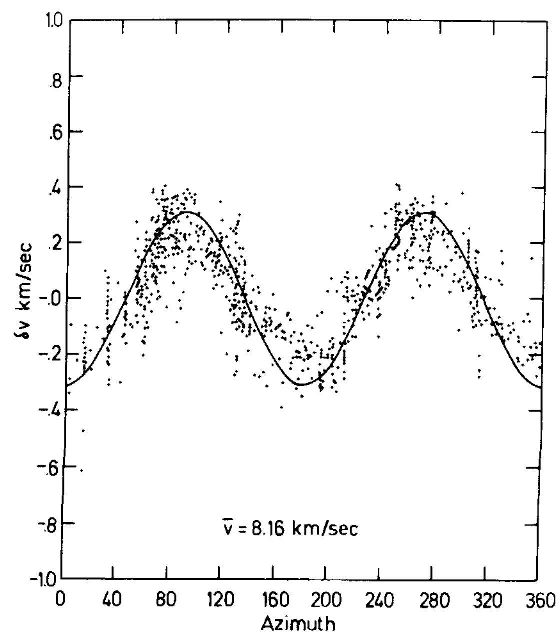

Figure 3. Azimuth dependence of *P*-wave velocities near Hawaii (after Morris *et al.* 1969).

clearly be recognized. The anisotropy amounts here to nearly 8 per cent. In Fig. 4 all available anisotropy data in the Pacific Ocean are summarized as to direction and magnitude. It is quite obvious that the directions of maximum velocity are nearly perpendicular to the ridge axis and parallel to the large faults. In the Atlantic Ocean, only one anisotropy experiment has been published so far (Keen & Tramontini 1970) which resulted in 8 per cent anisotropy.

These explosion seismic experiments tend to sound only the properties of the uppermost mantle immediately below the crust–mantle boundary. For an understanding of the formation of anisotropy a knowledge of the properties of the deeper parts of subcrustal lithosphere is of extreme importance. From an analysis of surface-waves dispersion, Forsyth (1973) derived a value of 2 per cent on the Nasca plate. The maximum anisotropy of 2 per cent occurs at periods between 50 and 90 s parallel to the spreading direction. From this, the upper mantle should be anisotropic down to a depth of 125 km, i.e. the whole subcrustal lithosphere and possibly also parts of the asthenosphere.

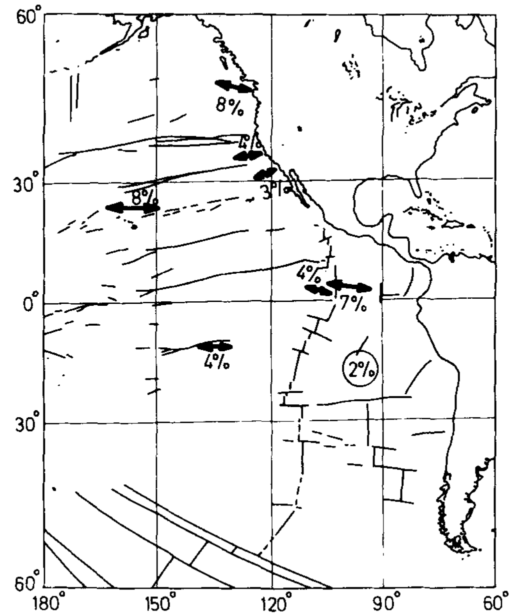

Figure 4. Distribution of anisotropy – amount and direction of maximum velocity – in the Pacific Ocean.

2.2 ANISOTROPY IN THE CONTINENTAL UPPER MANTLE

Until recently the seismic anisotropy of the uppermost mantle was reckoned to be a typical and exclusive property of the oceanic subcrustal lithosphere. The anisotropy was supposed in some way or other to mirror dynamic processes during the birth and early lifetime of the oceanic lithosphere below or in the immediate neighbourhood of oceanic ridges. Since the azimuthal dependence of the seismic velocity is induced most likely by glide-processes or non-hydrostatic stress, it cannot be excluded that such anisotropy is also generated by dynamical processes in the upper mantle below continents during motions of continental plates. Temperatures of 800–1000°C and pressures of more than 10 kbar at the crust–mantle boundary of continents are much higher than those under oceans and, therefore, the generation of a preferred orientation of olivine axes is more likely to be achieved (see Section 3.2, Fig. 8). In fact, recent explosion-seismic experiments have brought forth evidence that the seismic anisotropy may neither be restricted to the oceanic subcrustal lithosphere nor to its uppermost part below the crust–mantle boundary.

With one exception, however, no azimuth-dependent velocity anisotropy has been reported so far for the continents. Special anisotropy experiments like those in the oceans are more difficult to arrange on the continents and, therefore, have not been conducted so far. Most of the available explosion seismic data on the continents are not suited for an anisotropy analysis.

Although not designed for this purpose, the network of seismic refraction profiles in the Federal Republic of Germany (Fig. 5) offers a possible basis for such an analysis. Especially in southern Germany observations of the P_n phase propagating immediately below the crust–mantle boundary are available with an almost equal azimuthal distribution. Bamford (1973, 1976) subjected all available P_n data in the FRG to several kinds of time-term analysis. A very striking and unexpected result of his investigation was the large amount of anisotropy of 7–8 per cent. Even more surprising was the direction of the maximum velocity of 8.3–8.4 km/s with an azimuth of N (15–20)° E.

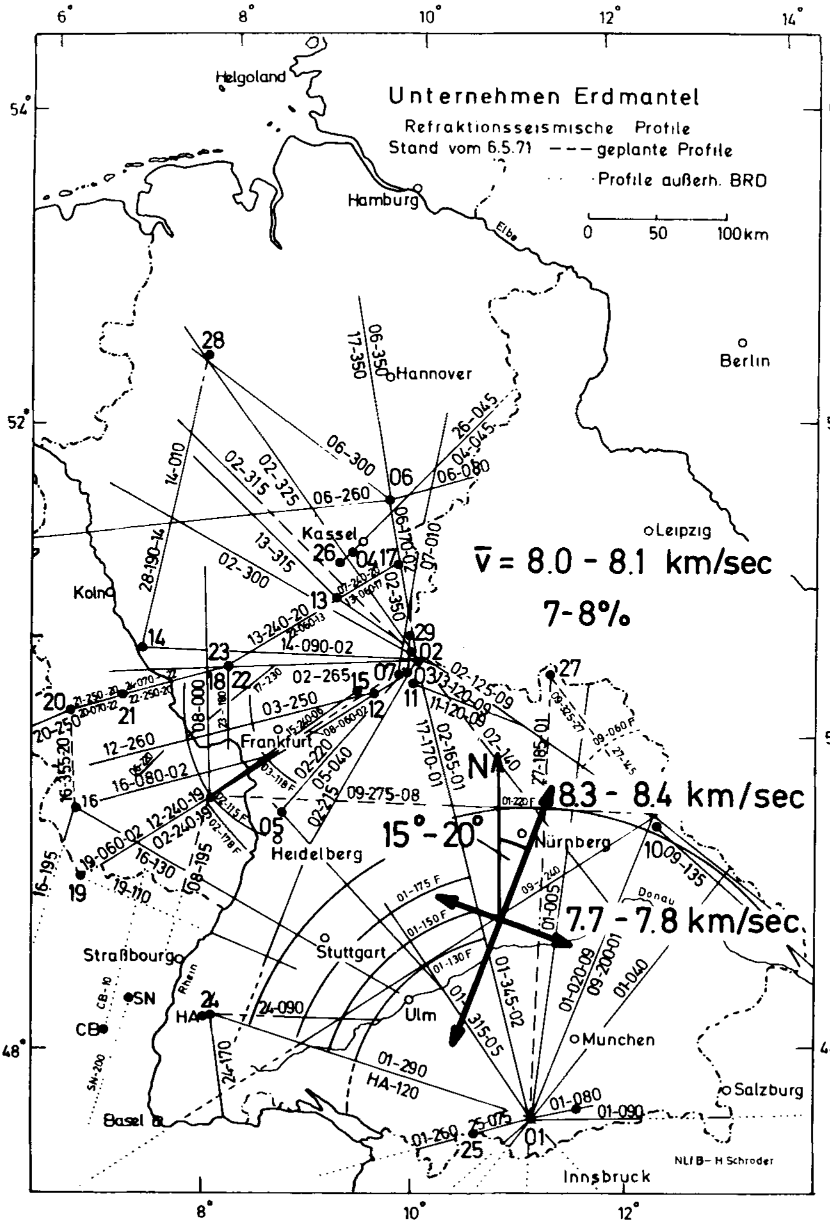

Figure 5. Anisotropy of seismic velocities in the upper mantle in southern Germany deduced from time-term analysis of all available P_n observations on explosion seismic profiles in the Federal Republic of Germany by Bamford (1973, 1976).

3 Explanation of the observed anisotropy and its genesis

3.1 MODELS INVOLVING ISOTROPIC LAYERS

The observed azimuthal dependence of seismic *P* velocities in the Pacific Ocean raised immediately the discussion on the origin of the anisotropy. Backus (1965) and Morris *et al.* (1969) pointed out that even in an isotropic medium an anisotropic stress field may induce an anisotropic velocity distribution. Dahlen (1972), however, could show from theoretical considerations that with 100-bar deviatoric stress the effect would be smaller by several orders of magnitude than the observed anisotropy of 4–8 per cent. Azimuth-dependent anisotropy may also be caused by a sequence of thin vertical layers with alternating high and low velocities (Fig. 6, e.g. dyke-injections). To explain the observed anisotropy of 4–8 per cent. Unlikely low velocities have to be assumed in the intruded layers. Furthermore the dykes would have to be oriented perpendicular to the ridge axis and parallel to the fracture zones – in opposition to our general understanding of spreading centre processes and structure.

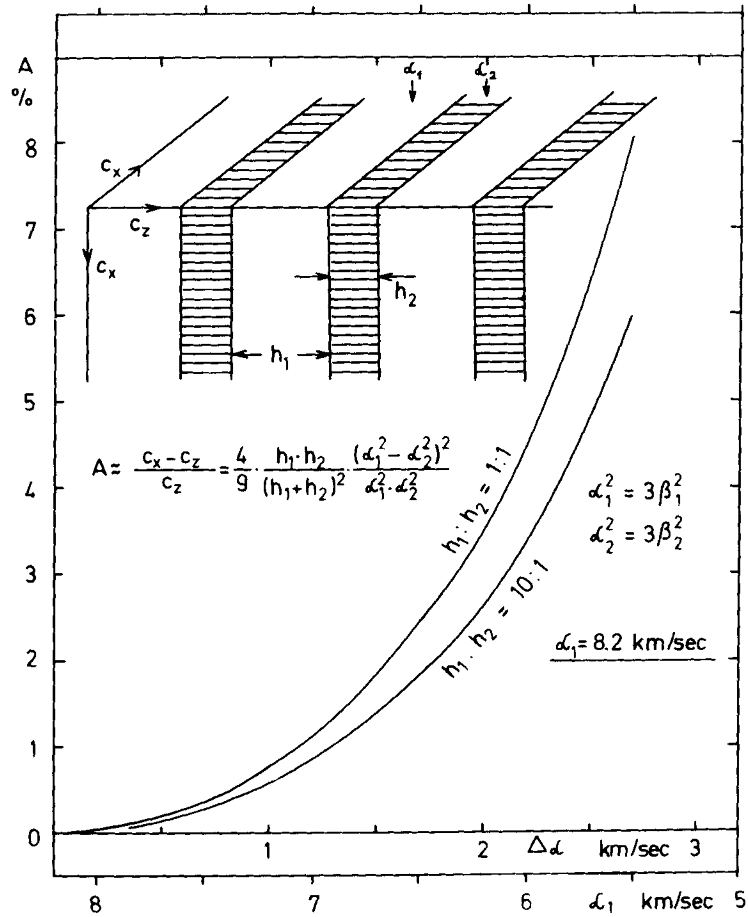

Figure 6. Model of a layered medium with azimuth dependence of seismic velocities caused by sandwiching of alternating isotropic high- and low-velocity layers (Fuchs 1975).

3.2 MODELS WITH INTRINSIC ANISOTROPY

Most authors, therefore, agree that the upper mantle possesses an intrinsic anisotropy produced by a preferred orientation of anisotropic crystals. Olivine in peridotite and dunite has enough velocity anisotropy and is present in the upper mantle in sufficient quantities. The relatively weak anisotropy of eclogite is a strong argument against an eclogitic composition of the oceanic uppermost mantle.

What are the causes for the preferred orientation of olivine in the lower lithosphere? The answer to this question may give some clues to the driving mechanism and to the stress distributions in the lithospheric plates.

The Hess model of olivine orientation had some shortcomings. Francis (1969) pointed out that the anisotropy observed during experiments in the Pacific Ocean is in the mean only 4 per cent and, therefore, much less than predicted by the Hess model. Francis' main concern, however, was the place where the orientation should take place. He excluded the possibility of creep and gliding in the temperature–pressure range (200° C, 2 kbar) at the oceanic crust–mantle boundary; rather brittle deformation should here be much more likely. Therefore, Francis places the genesis of the preferred orientation at greater depth directly beneath the midoceanic ridges; the largest shearing stresses should occur where the upwelling mantle material bends from the vertical into the horizontal direction. From laboratory experiments Raleigh (1968) finds that in the temperature range between 400 and 800° C gliding of the olivine occurs in the [100]-direction of the *a*-axis on all planes {0 k l} (pencil glide). Thus, the *a* axis is oriented dominantly into the direction of the flow lines. After the bending of the stream of mantle material, the *a* axis is pointing into a horizontal direction perpendicular to the ridge axis (Fig. 7). Francis' model solves the problem of low pressure and low temperature at the oceanic crust–mantle boundary; it leads, however, to larger anisotropy (20 per cent) than in the original Hess model.

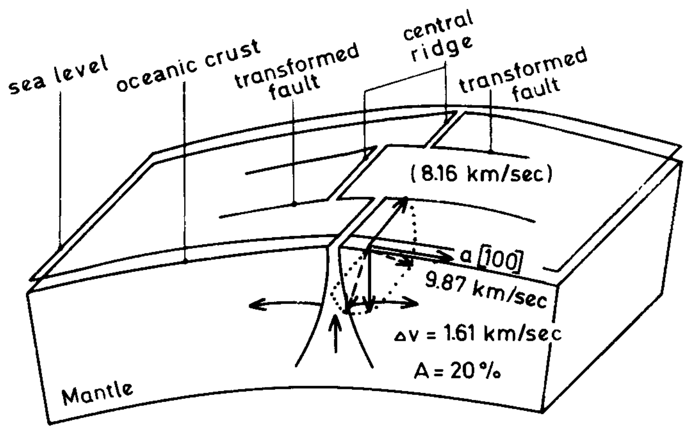

Figure 7. Orientation of olivine-axis in relation to the ridge axis (after Francis 1969).

The mode of deformation of olivine does not only depend on pressure and temperature but also on the rate of deformation. In a number of laboratory experiments Ave'Lallemant & Carter (1970) and Carter & Ave'Lallemant (1970) have studied the deformation of olivine at temperatures *T* between 300 and 1400° C, at pressures *P* between 5 and 20 kbar, and at deformation rates $\dot{\epsilon}$ between 10^{-3} and $10^{-8}\,s^{-1}$ (Fig. 8). By extrapolation from $\dot{\epsilon}=10^{-8}$ to geological deformation rates of $\dot{\epsilon}\sim 10^{-14}\,s^{-1}$ (over orders of magnitude!) they conclude that pencil glide [0 k l] [100] may start even at temperatures as low as 400° C. The softening effect of the low geological deformation rates and of H_2O (Griggs 1967) permits plastic flow to occur at temperatures as low as 200° C, however, in another slip system {110} [001] (see Fig. 8). Therefore, plastic flow and preferred orientation of olivine may take place also outside the midoceanic ridges.

In laboratory experiments Ave'Lallemant & Carter (1970) and Carter & Ave'Lallemant (1970) have studied syntectonic recrystallization of olivine. It can be observed for $\dot{\epsilon} = 10^{-3}\,s^{-1}$

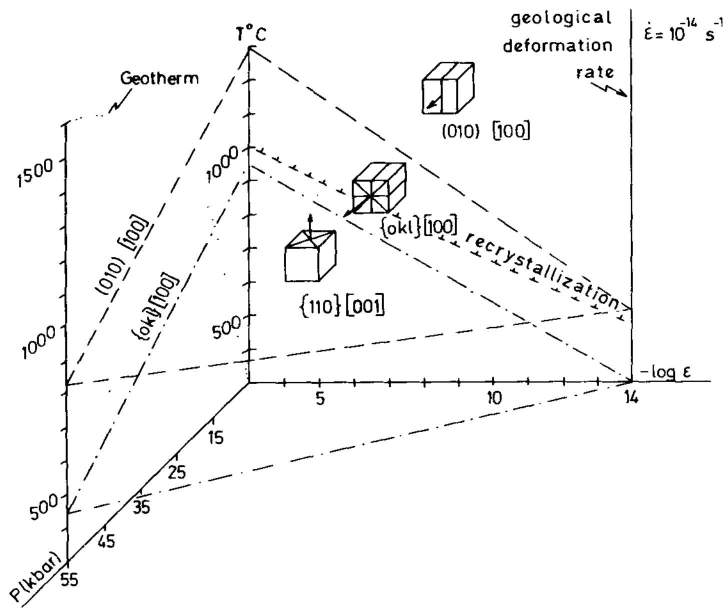

Figure 8. Laboratory experiments on the deformation of olivine at various P and T conditions as a function of the deformation rate $\dot{\varepsilon}$ (after Ave'Lallemant & Carter 1970).

and $P = 3$ kbar at temperatures exceeding 1050°C. Extrapolating to $\dot{\varepsilon} = 10^{-14}\,s^{-1}$ recrystallization may start as low as 500°C. Above the recrystallization temperatures a deformed body begins to develop stress-free grains. Both authors find that the b-axes [010] of the new crystals point into the direction of the maximum compressional stress σ_1. They, therefore, consider the process of recrystallization as the dominating mechanism for the generation of a preferred orientation in the oceanic lithosphere. This process would be restricted to the upwelling flow below oceanic ridges. Here below the 500°C isotherm the b-axis of the olivine is oriented into the direction of maximum compression in the shear stress field. Two objections must be raised against the predominance of this process: the azimuthal dependent anisotropy would be restricted to the immediate vicinity of oceanic ridges and descending plates; the b-axis of the olivine would not be oriented perpendicularly to the plane of maximum shear as observed in the field but would be inclined at 45°. Although recrystallization may occur, it is, therefore, probably not responsible for the anisotropy which is observed in the oceanic uppermost mantle.

The anisotropy of the zone of Ivrea in Northern Italy has been investigated both in the field and in the laboratory by Nicolas, Boudier & Boullier (1973) and Peselnick, Nicolas & Stevenson (1974). In these studies the direction of minimum velocity (olivine b axis, [010]) is perpendicular to the planes of foliation. The direction of maximum velocity (olivine a axis, [100]) is in the flow plane parallel to the foliation plane and points into the direction of flow lines. The authors also present evidence that foliation and flow planes were nearly horizontal when the crust was formed and conclude that anisotropy developed at least partly after the turnover of the uprising mantle flow into the horizontal direction and would not necessarily be a frozen-in feature as required by Francis (1969) and Ave'Lallemant & Carter (1970). The observations of Peselnick *et al.* (1974) indicate that pencil glide {0 k l} [100] and (010) [100]-gliding are responsible for the preferred orientation of olivine crystals and the observed anisotropy.

4 Further evidence for anisotropy in the lower lithosphere

The field observations of anisotropy in the oceanic and continental upper mantle were not concerned with wave propagation in the topmost upper mantle other than immediately below the crust–mantle boundary. Until now these experiments are much too scarce to establish with certainty a relationship between anisotropy and dynamical processes in the lower lithosphere. Laboratory experiments, however, make it quite likely that a preferred orientation of anisotropic minerals can be impressed by the conditions of pressure, temperature and strain rate of the upper mantle both in the oceanic and the continental lithosphere through plastic flow and non-hydrostatic stress fields. However, there are a number of possible models of genesis of anisotropy which can be distinguished only if further observational evidence becomes available.

In this paragraph further direct or indirect evidence on dynamic processes in the lower lithosphere is compiled. Recently, reports on high seismic P velocities in the oceanic upper mantle (e.g. Hales, Hesley & Nation 1970; Zverev 1970) and continental upper mantle (e.g. Ryaboi 1966; Kosminskaya, Puzyrev & Alexeyev 1972; Hirn *et al.* 1973; Kind 1974) have accumulated since long-range observations with depth of penetration to about 100 km became available. For a review of anisotropy in the oceanic lower lithosphere see Bottinga, Steinmetz & Allègre (1976). In both regions *in-situ* velocities well above 8.2 km/s and up to 8.8–9.0 km/s have been measured. If these velocities are corrected to surface temperature and pressure the STP velocities range between 8.6 and 9.2–9.4 km/s. Minerals with such high velocities and sufficient abundance in the upper mantle are not too numerous. Only garnet and olivine are possible candidates. Especially the higher velocity can only be explained by the anisotropic velocities of olivine (Bottinga *et al.* 1976).

Another indication that the presence of seismic anisotropy in the upper mantle is connected with the dynamics of plate motion comes from the correlation of the direction of maximum velocity with the direction of other features indicative of plate dynamics.

Hess (1964) was the first to point out the coincidence of spreading direction with the direction of maximum velocity and that, conversely, the slowest velocity was parallel to the ridge and perpendicular to the large shear fractures.

On the European continent the direction N 20° E of maximum P_n velocity determined in southern Germany (Bamford 1973, 1976) correlates with a number of trends related to plate dynamics.

In Fig. 9 the following features are compiled: (i) the motion of the European plate as calculated by Minster *et al.* (1974) has this same direction; (ii) it coincides also with the direction into which subcrustal diapirs are supposed to have been sheared in Norway (Aki, Christofferson & Husebye 1976); (iii) it is parallel to the strike of the Rhinegraben and, therefore, also parallel to the axis of compression during the opening of the graben; (iv) new earthquake mechanisms from fault plane solutions in southern Germany indicate that in the lower crust between a depth of 12 and 25 km the axis of maximum compressions is horizontal with an azimuth of about N 10° W (Bonjer 1977).

It is not yet fully understood how the stress field induced by plastic flow in the asthenosphere is transmitted through the subcrustal lithosphere into the brittle crust. A shearing stress applied by horizontal flow in the mantle to the base of the crust which is constrained in its motion by neighbouring plates can produce a strong horizontal compressional stress component in the lower crust in the direction of flow in the upper mantle. The N 10° W direction deviates only by 15–20° from the direction of maximum velocity. In general the three directions – maximum compressive stress in the lower crust, direction of plate motion, shearing in subcrustal lithosphere – need not coincide since the motion of the plate is constrained by its neighbours; but if they do and, in addition, match the direction of maximum

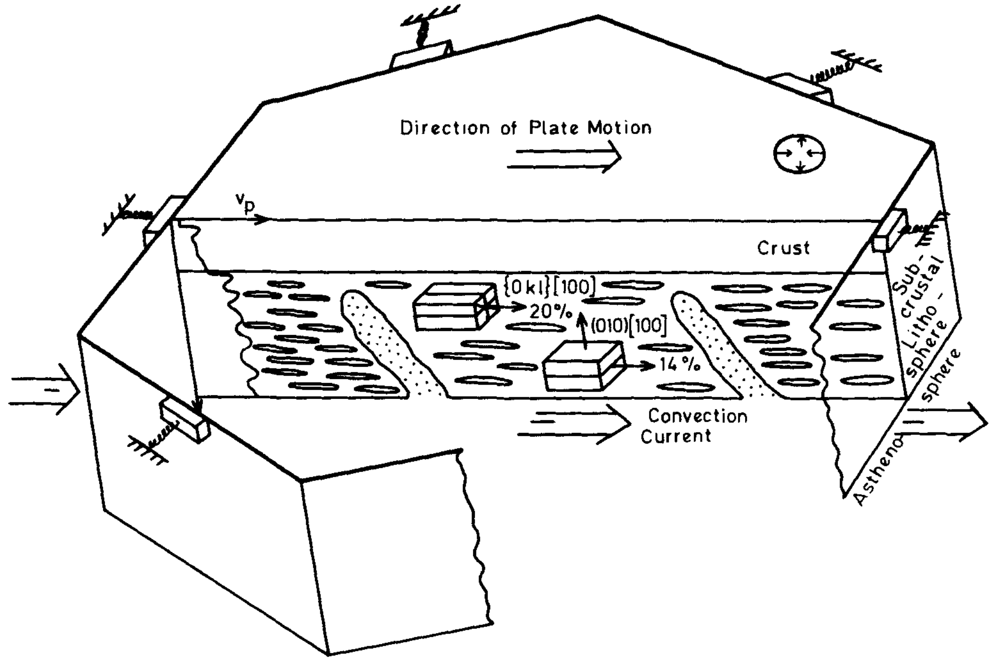

Figure 9. Schematic compilation of tectonic features on the European plate related to the dynamics of plate motion. The direction of maximum velocity (Bamford 1973, 1976), probably related to the direction of the *a* axis of preferred olivine orientation, correlates with the direction of plate motion (Minster *et al.* 1974) and with the azimuth of inclined subcrustal diapirs deduced by Aki *et al.* (1976) in Norway. The presence of thin laminas in the subcrustal lithosphere with a nearly parallel orientation is deduced from the occurrence of high *P* velocities V_p in this region (Hirn *et al.* 1973; Kind 1974), low-frequency tunnel waves and teleseismic transmission of high-frequency P_n and S_n waves (Fuchs & Schulz 1976). It is suspected that the cause for the parallel orientation of the laminas, for the preferred orientation of olivine and for the deformation of subcrustal diapirs is the same plastic deformation of the lower lithosphere induced by convection currents in the asthenosphere. The major compressional axis in the lower crust of southern Germany tends also in the same direction (Bonjer 1977). The direction of plate motion is constrained by forces from neighbouring plates and has not to coincide with the other directions mentioned above (see text).

P_n velocity a causal relation must strongly be suspected. Schematically Fig. 9 displays also the orientation of olivine (compare Fig. 7) in the field of horizontal plastic flow induced by convection currents in the asthenosphere. Immediately below the crust–mantle boundary pencil gliding $\{0\,k\,l\}$ [100] is supposed to be dominant: the *a* axis points into the direction of flow while the *b* and *c* axes form girdles perpendicular to the *a* axis. This would lead to a horizontal anisotropy of 20 per cent for pure olivine. With increasing pressure and temperature the plastic flow changes to a (010) [100] glide with a vertical *b* axis. The horizontal anisotropy decreases to 14 per cent for pure olivine. Thus a variation of the amount of anisotropy with depth can be expected even if the composition of the lower lithosphere does not change.

If anisotropy in the subcrustal lithosphere is caused by coupling of dynamical processes in the asthenosphere to the plates this anisotropy should not be confined to a few kilometres immediately below the crust–mantle boundary. Is there evidence that the whole subcrustal lithosphere is affected by the dynamical processes of plate motion?

It has been suggested that the high velocities in the lower lithosphere occur in thin laminas with more or less parallel horizontal orientation (Fuchs & Schulz 1976). This structure is supported by the observation of low-frequency tunnel waves reflected and transmitted through the lower lithosphere into the mantle and by high-frequency transmission of P_n and S_n waves to teleseismic distances. In the latter case the thin high-velocity laminas form a wave guide only if the laminas are sufficiently parallel. It is suggested that the parallel orientation is generated by the same mechanism which produces the preferred orientation (Fig. 9). For a detailed discussion of this subject the reader is referred to Fuchs & Schulz (1976).

P-wave velocities of phases from the subcrustal lithosphere have been determined on various long-range profiles in a number of azimuths. However, nowhere are long-range observations available which sound the subcrustal lithosphere at the same location in a sufficient number of azimuths ($\geqslant 3$). In southern Germany and eastern France three profiles, two of which have reversed observations, sound the lower lithosphere to a depth of about 45 km where the following *P* velocities have been derived: 8.2 km/s (N10°W), 8.6 (N25°E), 8.45 (N60°E) (Ansorge, Bonjer & Emter 1976). If a certain lateral homogeneity both in mean velocity and in direction of anisotropy is assumed these observations are compatible with an observed maximum velocity of 8.6 km/s in the direction N30°E and a computed minimum velocity of 7.7 km/s N60°W, corresponding to an anisotropy of 11 per cent. The amount and direction of this anisotropy are not very reliable, but from the results of these observations it must be concluded that either anisotropy extends into the lower lithosphere or strong lateral variations of seismic velocities are present.

From an analysis of *P* and *S* waves on long-range profiles in western Europe Hirn (1977) derives evidence for anisotropic properties of the subcrustal lithosphere of which birefringence of *S* waves is the most convincing observation.

The many long-range observations in the USSR (Ryaboi 1966) are too widely separated and usually observed only in a single azimuth to permit an estimate of anisotropy. The same is true for the *Early Rise* experiment in the USA – for a recent review see Ansorge (1975): although the azimuthal coverage is quite dense the various lines traverse rather different provinces of the upper mantle.

Conclusions

From the present knowledge of subcrustal velocities in the upper mantle, a general property of the lower lithosphere seems to be a velocity anisotropy which depends on the azimuth of propagation both below the oceans and the continents. This anisotropy has been derived most reliably for the P_n phase immediately below the crust–mantle boundary. Evidence for a similar anisotropy in deeper parts of the lower lithosphere is accumulating. Laboratory experiments suggest that this anisotropy is the result of a preferred orientation of olivine crystals induced by plastic flow. The correlation of the directions of maximum velocity with other tectonic trends – direction of seafloor spreading, direction of plate motion, direction of shearing of subcrustal diapirs, direction of maximum compression from fault plane solutions and other tectonic elements – suggest a causal connection between the forces driving the lithospheric plates and the generation of anisotropy.

However, it must be kept in mind that there is no simple connection which permits determination of, for example, the direction of convection currents from the direction of anisotropy. At present it can only be stated that any realistic dynamic model of plate motion has to take into account the occurrence of anisotropy. Furthermore, the observation of anisotropy places new constraints on petrological models of the lower lithosphere.

To obtain a better insight into the dynamical processes of the subcrustal lithosphere a combined experiment is proposed between a specially designed study of depth-dependent anisotropy and, at the same locality, a survey of lateral heterogeneities of the upper mantle by the analysis of travel time anomalies from teleseisms. The determination of the depth dependence of anisotropy could possibly provide new bounds on the viscosity of the lower lithosphere and on the differential motion between the 'rigid' plates and convection currents in the asthenosphere which would have to be taken into account in numerical models.

Acknowledgments

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