Global patterns of azimuthal anisotropy and deformations in the continental mantle

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SUMMARY

We present a summary of measurements of azimuthal anisotropy in the continental mantle based on the SKS technique and performed mostly with the active participation of the authors. The directions of polarization of the fast quasi-shear wave and the time delays between the quasi-shear waves are obtained at nearly 70 locations in all continents, except Antarctica. These data are interpreted in terms of lattice-preferred orientation of olivine which is caused by deformations in the mantle. The depth interval responsible for anisotropy is unknown but the data suggest that it may reach at least 300 km. The fast directions in SKS do not show clear correlation with the fast directions of the teleseismic P at the same seismograph stations.

In the regions of present-day convergence the fast direction of anisotropy usually aligns with the plate boundary. This correlation implies that the direction of shortening is the same in the crust and the upper mantle. In the regions of rifting, the inferred direction of mantle flow usually aligns with the direction of extension in the crust.

Outside the regions of recent tectonic activity we, most likely, observe a combined effect of frozen anisotropy in the subcrustal lithosphere and of recently formed anisotropy in the asthenosphere. On a global scale, in these regions there is a positive correlation between the absolute plate velocity directions and the fast directions of anisotropy. The correlation is especially strong in central and eastern parts of North America. A clear absence of any evidence of large-scale azimuthal anisotropy in the data of long-range refraction profiling for the upper 100 km of the mantle of that region implies that the effect in SKS is generated mainly at greater depths, in the asthenosphere. Orientation of olivine at these depths reflects recent and present-day flow in the mantle rather than processes of a distant geologic past.

Key words: anisotropy, Earth's mantle, plate tectonics.

1 INTRODUCTION

Movements and deformations of the Earth's crust are often regarded as a reflection of deformations and flow in the mantle. Hence, understanding of processes in the interior of the Earth is one of the main objectives of geophysical studies. Determinations of seismic anisotropy in the mantle are particularly useful because the most abundant mineral constituent of the mantle, olivine, is highly anisotropic and its crystal orientation depends on finite strain. The axis (a) of olivine (100, axis of the fastest P velocity) tends to be parallel to the principal extension axis of the strain ellipsoid or to the direction of flow, if that is in the form of simple shear (Nicolas & Poirier 1976; McKenzie 1979; Nicolas & Christensen 1987; Ribe 1989; Ribe & Yu 1991).

The global patterns of azimuthal anisotropy in the upper mantle were first determined from observations of azimuthal variations of phase velocities of long-period surface waves (e.g. Tanimoto & Anderson 1985). Resolving power of these techniques is in the range of a few thousand kilometres, which is clearly not sufficient for continents where the scales of structures of interest are often in the range of a few hundred kilometres or even less than this.

The studies of anisotropy in the continental upper mantle evolved mainly in two directions. The first is based on observations of azimuthal traveltime variations of the
P-waves (e.g. Bamford 1977). These studies require expensive observational systems and, therefore, were conducted only in a few locations.

The second direction is based on the observations of shear wave splitting which is regarded as the most convincing diagnostic property of azimuthal anisotropy. The first observations of shear wave splitting due to propagation in the upper mantle were made in the high-frequency range where the time delay between the quasi-shear waves (near 1 s) is comparable with the dominant period of oscillation (Ando, Ishikawa & Yamazaki 1983). Unfortunately, the records in this frequency range are often contaminated by the effects of wave scattering, multipathing and source complexity. The other limitation is due to the fact that short-period S-waves with a steep incidence, which is a pre-requisite for these studies, are observed only in the vicinity of deep seismic zones. The method for observing splitting in the seismic phases SKS and SKKS (Vinnik, Kosarev & Makeyeva 1984) is free from these limitations. First, the observations can be made in a frequency range where the dominant period of oscillation is much longer (say, 5–10 times) than the time interval between the split waves. In this range the indirect effect of splitting is strong enough to be observed whereas the effects of lateral heterogeneity are relatively weak. Secondly, the complexity of the source does not affect the measurements of anisotropy since polarization of SKS is completely independent of the source properties. Thirdly, SKS, SKKS and similar phases are recorded free from interference with other seismic phases in an extremely broad distance interval (between, approximately, 86 and 170°), which makes them observable at any reasonably sensitive seismograph station.

This approach was used in several recent studies (e.g. Kind et al. 1985; Silver & Chan 1988; Ansel & Nataf 1989; Vinnik, Farra & Romanowicz 1989a; Makeyeva, Plesinger & Horalek 1990; Savage, Silver & Meyer 1990; Vinnik et al. 1990; Milev & Vinnik 1991; Silver & Chan 1991).

In this paper we present a summary of data on splitting of SKS obtained with the active participation of the authors. Among others these data include results of the analyses of observations of Geoscope and NARS broad-band networks (Vinnik et al. 1989a), of the CIS network of analogue broad-band seismograph stations (Vinnik et al. 1991), of long-period channels of GDSN (Milev & Vinnik 1991), of Czechoslovak (Makeyeva et al. 1990) and former East Germany networks (Bormann et al. 1991). As an exception, we also included in the data set considered, a few measurements made by Silver & Chan (1988) for North America. Their measurements, which are based on a practically similar approach, play an important role in the discussion of the depth distribution of azimuthal anisotropy in tectonically stable continental regions (Section 4.4). As a result, we have a homogeneous data set for almost 70 seismograph stations in all continents, except Antarctica (Fig. 1, Table 1). This number is a few times larger than in any other previously published study.

2 METHOD OF MEASUREMENTS

The method is based on the phenomenon of shear wave splitting in anisotropic media (see e.g. Babuska & Cara 1991). In homogeneous anisotropic media for any direction of propagation there are three waves whose velocities and

Figure 1. Summary of measurements of the parameters of azimuthal anisotropy (Table 1). Arrows show the directions of polarization of the fast split wave. The values of δt are indicated by the size of the circle. The scale indicates that δt is greater than the attached value. Less reliable measurements are marked by open circles.
eigenvectors of the corresponding Christoffel matrix. The
which, at the same time, is broadly consistent with the
polarizations are determined as the eigenvalues and the
properties of upper mantle peridotites (e.g. Estey
Douglas 1986). It follows from the properties of the
symmetry (Vinnik
et al. 1989a, b). This is the simplest form of azimuthal anisotropy
Table 1. List of stations and the parameters of azimuthal ani-
station Lat. Lon. α β M&J Ref.
ARNO 34.96 106.5W 35 1.1
ARH 39.91 32.8S 25 1.3
BCAO 4.46 18.53 60 0.6 23
BDF 15.7S 47.9W 40 1.2
BGR 60.4N 5.3E 105 1.1 122
BCOØ 4.6N 74.0W 20 1.2
CTNO 18.9N 99.0W 100 1.6
CTAO 20.15 146.3E 40 1.0
GAC 45.7N 75.5W 85 0.9 252
GDR 69.3N 53.5W 120 1.2 261
JAS 37.3N 120.4W 80 1.5
KAO 34.5N 69.0W 60 1.8
KEH 69.8N 2.0W 70 1.1 142
KON 59.6N 9.6W 20 0.8 127
MAJ 36.3N 138.3E 130 0.9
McAO 32.9S 117.2E 60 1.5
RSON 50.9N 93.7W 80 1.7 241
SCP 40.8N 77.9W 80 1.2 250
SIMO 25.6N 91.8W 120 1.3
SLR 25.7S 28.3W 90 0.8 37
SNDO 41.3S 174.7E 20 1.8
TAT 25.0N 121.3E 60 1.8
TAU 42.9S 147.3E 80 0.6 220
ZBO 29.0N 68.1E 120 0.8 198
ABR 37.9N 58.3W 25 0.8 Vinnik et al. 1991
SBR 41.7N 43.5E 90 1.7
BRV 53.1N 30.2W 1.2 183
ELT 53.2N 86.3E 80 1.2
FRO 62.8N 74.8E 85 1.0
GAR 39.0N 70.3E 65 1.7
GNS 39.0M 44.3N 100 1.3
ILT 67.9N 178.9W 15 0.9 173
IRK 52.2N 104.3E 75 1.1
KHE 80.6N 58.0E 63 1.2 171
MID 60.1N 150.2W 20 0.8
MRI 69.4N 88.6W 1.0 200
ORN 55.2N 36.6W 60 0.6 155
REM 50.9N 80.3W 170 1.9 190
SRT 62.9N 152.4E 110 1.2
SVE 56.8N 60.6W 0.4 175
TAS 41.3N 49.3W 80 0.8
TIX 71.6N 128.9W 60 1.7 103
ZAK 50.3N 103.3W 130 1.2 Vinnik et al. 1989a
ACD 11.5N 43.8E 45 1.2
CAT 5.0N 52.3W 100 1.2
INO 35.4N 137.0E 170 1.3
KIP 21.0N 156.0E 45 1.5
KRD2 56.8N 9.2W 60 1.2
KRD3 52.1N 5.2W 70 0.5
NEU 30.9N 5.8W 70 0.5
SCZ 36.4N 121.4W 100 1.3
SSB 45.2N 4.3E 140 1.0
TOL 39.9N 4.0E 100 1.3 110
VMN 42.6N 71.5W 80 0.8 254
RKC 54.1N 13.8E 100 1.1 137
KSP 50.8N 16.3E 120 0.9 128
BRG 50.9N 14.0E 100 1.2
CUL 51.3N 13.0E 100 1.0
MOX 70.5N 11.6E 100 0.7
GRF 4.5N 11.2E 90 0.7 133 Vinnik et al. 1989b
STA 48.8N 9.2W 50 0.2
VTS 42.6N 23.3E 120 1.2 190
COL 64.0N 147.7E 77 1.6
LON 46.8N 121.8E 90 0.8
SCP 35.6N 85.4W 46 0.8 125
KNT 62.5N 114.4W 48 1.1 225
OKSD 44.1N 104.0W 54 0.6 230

polarizations are determined as the eigenvalues and the
eigenvectors of the corresponding Christoffel matrix. The
matrix is symmetric and the polarizations of the waves are
mutually orthogonal. The method was formulated for a
medium with hexagonal symmetry and a horizontal axis of
symmetry (Vinnik et al. 1984; Kind et al. 1985; Vinnik et al.
1989a, b). This is the simplest form of azimuthal anisotropy
which, at the same time, is broadly consistent with the
properties of upper mantle peridotites (e.g. Estey &
Douglas 1986). It follows from the properties of the
Christoffel matrix that the vector of polarization of the
S2-wave in this medium is always orthogonal to the axis of
symmetry. In SKS, whose angle of incidence in the upper
mantle is near 10°, the vectors of polarization of S1 and S2
lie in a plane which is nearly horizontal. With the
assumption that the axis of symmetry coincides with the (a)
axis of olivine, the vectors of polarization of S1 and S2 are
oriented in the horizontal plane as shown in Fig. 2. The S1
and S2 waves propagate with different velocities: in the
model considered S1 is fast and S2 is slow, and the velocity
difference is almost independent of the azimuth. Due to
splitting, the linear particle motion in the incoming SV at
long periods becomes elliptic.

The amplitude relationship between the plane quasi-shear
waves and the incoming SV was determined by numerical
modelling of the wavefield generated in an anisotropic layer
by SV coming from an isotropic half-space (Kosarev et al.
1979). It was found (Vinnik, Kosarev & Makeyeva 1988;
Vinnik et al. 1989b) that the amplitudes obtained in
numerical experiments can be simulated by projecting the
amplitude of SV onto the directions of polarization of the
quasi-shear waves (see Fig. 2). The predictions based on
numerical experiments were tested by comparing them with
the observations of SKS and SKKS coming to the GRF
array in many azimuths (Kind et al. 1985). The results of this
test were strongly positive.

Almost all of the results discussed in this paper were
obtained with the algorithm by Vinnik et al. (1989a). For
the sign convention adopted in Fig. 2, the radial (R) and
transverse (T) components of a harmonic component of
SKS can be expressed as:
\[ R(t) = \cos^2 \beta e^{i\omega t} + \sin^2 \beta e^{i(\omega t - \delta t)}, \]
\[ T(t) = -0.5 \sin 2\beta [e^{i\omega t} - e^{i(\omega t - \delta t)}], \]
where \( t \) is time, \( \omega \) is circular frequency, \( \beta \) is the angle shown in
Fig. 2, and \( \delta t \) is traveltime delay between split waves. For

Figure 2. The geometric relation between the \((R, T)\) reference
frame and the orientation of the fast (a) axis of olivine used to
calculate amplitudes of the split waves.
a given record of the $R$ component we can calculate the corresponding theoretical transverse component $T$

$$T = F \ast R.$$ 

In agreement with equation (1) the Fourier transformation of the filter $F$ is given by

$$f(\omega) = -0.5 \sin \beta \frac{1 - e^{-i \omega t}}{\cos \beta + \sin \beta e^{-i \omega t}} = -0.5(\sin \beta) \omega \delta t,$$

i.e. $T$ has the form of the derivative of $R$.

We calculate the penalty function

$$E(\alpha, \delta t) = \left\{ \frac{1}{N \text{ events}} \sum \left| T(t) - \left( T(\alpha, \delta t) \right) \right|^2 dt \right\}^{1/2},$$

where $\alpha$ is a trial azimuth for the axis of fast velocity counted clockwise from North, $\delta t$ is a trial time delay, $T_0$ is the observed $T$ component and $N$ is the number of events. The optimum pair of the parameters $\alpha$ and $\delta t$ corresponds to the minimum of $E$. Accuracy of the results depends on the noise which is present in the seismograms, the noise being not only the ambient noise but also the effects of the complexities in the medium which are not accounted for by the model. The properties of these noise are not well known. Hence, the best way to assess the accuracy of the results is to compare estimates corresponding to different seismic events.

Some unpublished numerical experiments were carried out with a dipping axis of symmetry. If the angle between the axis and the horizontal plane was less than, approximately, $30^\circ$, the corresponding perturbations of the amplitudes of the quasi-shear waves were small. For larger angles, variations of the $T$ component amplitude with the azimuth at low frequencies contained a significant component with the period of $360^\circ$. In real data this component was very weak (Kind et al. 1985; Makeyeva et al. 1990).

We usually assume that the $(100)$ axis of olivine is the axis of symmetry which is true if the mechanism of deformation is shear flow. If the rock is deformed by uniaxial compression, then the axis of symmetry is $(010)$ ($b$, the direction of the lowest $P$ velocity in olivine), which is oriented in the direction of shortening (Christensen & Crosson 1968). In the case $S1$ which is polarized in the direction of the axis of symmetry is slow whereas $S2$ is fast. Using observations of SKS, this symmetry cannot be distinguished from that assumed in Fig. 2.

### 3 DATA AND RESULTS

Most of the results presented in Fig. 1 and Table 1 were previously described in the literature. However, nearly 70 per cent of the data in Fig. 1 came from the CIS seismograph network and long-period channels of the Global Digital Seismograph Network (GDSN). We describe these data since they were previously presented only in a short form, in Russian (Vinnik et al. 1991 and Milev & Vinnik 1991, respectively). The list of events which were used in these studies is presented in Table 2.

### Table 2. List of events used by Milev & Vinnik (1991) and Vinnik et al. (1991).

<table>
<thead>
<tr>
<th>Event Date d m</th>
<th>Time h m s</th>
<th>Lat. deg</th>
<th>Lon. deg</th>
<th>Depth km</th>
<th>Station Distance Back az. m</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 17 Jun 67</td>
<td>5 0 12</td>
<td>58.45</td>
<td>26.86</td>
<td>136 5.9</td>
<td>OBS 124.2 109.7</td>
</tr>
<tr>
<td>2 14 Aug 68</td>
<td>22 14 20</td>
<td>0.18</td>
<td>119.7E</td>
<td>22 6.1</td>
<td>OBS 86.0 95.6</td>
</tr>
<tr>
<td>3 25 Jul 71</td>
<td>2 2 43</td>
<td>0.60</td>
<td>120.4W</td>
<td>80 6.4</td>
<td>OBS 105.3 174.2</td>
</tr>
<tr>
<td>4 25 Jun 82</td>
<td>17 22 38</td>
<td>26.05</td>
<td>64.3E</td>
<td>20 6.1</td>
<td>SET 92.1 249.4</td>
</tr>
<tr>
<td>5 14 Aug 77</td>
<td>27 13 54</td>
<td>7.95</td>
<td>107.6E</td>
<td>60 5.8</td>
<td>IIT 91.2 252.0</td>
</tr>
<tr>
<td>6 7 Sep 78</td>
<td>9 45 58</td>
<td>5.36</td>
<td>137.2E</td>
<td>182 6.5</td>
<td>ASI 113.9 90.7</td>
</tr>
<tr>
<td>7 6 Sep 77</td>
<td>11 45 35</td>
<td>14.35</td>
<td>167.1E</td>
<td>215 5.9</td>
<td>GRS 123.0 89.7</td>
</tr>
<tr>
<td>8 27 Oct 79</td>
<td>14 35 58</td>
<td>13.88</td>
<td>90.9W</td>
<td>62 5.7</td>
<td>TAS 122.2 337.1</td>
</tr>
<tr>
<td>9 22 Nov 79</td>
<td>24 21 15</td>
<td>24.35</td>
<td>67.2W</td>
<td>150 5.7</td>
<td>FBU 134.3 283.2</td>
</tr>
</tbody>
</table>
Most of the data of the CIS seismograph network are obtained from analogue (paper) records of the broad-band Kirnos seismographs SK and SKD. The response of this system to displacement is flat between 1 and 20 s; recording speed is usually 30 mm min⁻¹. The only exceptions are digital data of station BRV and the IRIS station GAR. The analogue data were digitized and low-pass filtered to suppress noise. Only those records were used where the signal in the transverse component was well above the noise.

Fig. 3 shows representative examples of the digitized records of station OBN. Transverse component of SKS in every record is clearly visible and is much stronger than the noise. The signal in the transverse component is always much stronger than the noise and is much stronger than the noise. The signal in the transverse component is always much stronger than the noise and is much stronger than the noise.

In most cases the estimates are based on two or three recordings of a quality comparable with that in Fig. 3. A few determinations were rated less reliable. For example, the result for ASH was rated less reliable because it was obtained from two records of medium quality from the same azimuthal sector. The similarity of the azimuths of the events suggests the possibility of a similar bias in both records. The result for TIK which is rated less reliable is fairly accurate with respect to \( \delta \alpha \), but the estimate for \( \delta \delta \) is rather uncertain. Sometimes the reported amount of recordings for one station is much larger than in our study, but this does not mean much unless the events are well distributed in azimuth. If the events, as often happens, are clustered in two or three back azimuths, the whole cluster is practically equivalent to one well-recorded event.

When the study by Vinnik et al. (1991) was already published, some of its findings were confirmed by independent measurements. The study of shear wave splitting in central Asia by Makeyeva, Vinnik & Roecker (1992) is based on a large amount of high-quality analogue recordings of SKS. The estimates of \( \delta \alpha \) obtained in that study for GAR, TAS and FRU are similar to those reported by Vinnik et al. (1991). The estimates of \( \delta \delta \) for FRU and TAS are also similar, whereas for GAR \( \delta t = 1.0 \) s instead of 1.7 s. The estimates for IRIS station OBN reported by Silver & Chan (1991) (\( \alpha = 5^\circ \), \( \delta t = 0.6 \) s) are similar to those observed in the analogue recordings of the same station and shown in Fig. 1.

The data reported by Milev & Vinnik (1991) were obtained from long-period recordings of GDSN stored at CD-ROM. Nearly 50 per cent of the stations considered were SRO and ASRO which do not record oscillations with periods between, approximately, 2 and 14 s. The theoretical amplitude ratio between the long-period results for BCAO (60° and 0.6 s) are close to the long-period one by Shih, Meyer 1991 for intermediate-depth events in central Asia by Makeyeva, Vinnik & Roecker (1992) is based on a large amount of high-quality analogue recordings of SKS. The estimates of \( \delta \alpha \) obtained in that study for GAR, TAS and FRU are similar to those reported by Vinnik et al. (1991). The estimates of \( \delta \delta \) for FRU and TAS are also similar, whereas for GAR \( \delta t = 1.0 \) s instead of 1.7 s. The estimates for IRIS station OBN reported by Silver & Chan (1991) (\( \alpha = 5^\circ \), \( \delta t = 0.6 \) s) are similar to those observed in the analogue recordings of the same station and shown in Fig. 1.

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Figure 3. Representative examples of the radial (R) and transverse (T) components of SKS at station OBN: (a) event 3 of Table 2; (b) event 14; (c) event 26; (d) event 27; and (e) event 45. Vertical bars indicate the time interval used for the analysis.
Global azimuthal anisotropy

4. DISCUSSION

Due to the relatively large amount of homogeneous measurements at our disposal, we can interpret these data by treating them in a quasi-statistical way. The average value of $\delta t$ in our data is near 1.0 s, the maximum values are near 2.0 s. Depending on the assumed difference between the velocities of the quasi-shear waves, the thickness of the anisotropic layer may lie somewhere between 100 and 1000 km. The observed $\delta t$ are clearly too large to be related to the crust, where their values are usually in the range 0.1–0.3 s (e.g. Kaneshima 1990; Shih & Meyer 1990). On the other hand, the variations in the parameters of anisotropy on a scale of about 100 km that are observed at some locations, would be impossible if the sources were in the lower mantle. Thus, there are good reasons to believe that the observed effects are related mainly to the upper mantle. In this section we address the following questions:

- how are anisotropy and the corresponding deformations in the upper mantle related to the present-day or recent (Cenozoic) deformations in the crust?
- What is the origin of anisotropy observed outside the presently active tectonic regions: (1) frozen anisotropy in the subcrustal lithosphere; (2) present-day flow in the asthenosphere; (3) a combination of them? What inferences about the properties of the upper mantle and, especially, the distribution of anisotropy with depth can be made, if the data in Fig. 1 are combined with the other anisotropy-related seismic observations, such as azimuthal variations of the P-wave traveltimes?

4.1 Shear wave splitting and tectonics

A number of data in Fig. 1 are related to the presently active zones of convergence. A look at the map in Fig. 1 immediately reveals that the fast directions of anisotropy in these regions are often nearly parallel to the plate boundaries and the corresponding mountain ranges. In Eurasia we observe this relationship in the Balkans (VTS), the Caucasus (BKR, GRS), the Tien Shan (TAS, FRU), the Pamirs (GAR), and the Hindu–Kush (KAAO). The direction of fast velocity at SHIO which is located at plateau Shillong in close vicinity of the Himalayas, coincides with

Figure 4. Plots of $E(\alpha, \delta t)$ corresponding to the records in Figs 3(a)–(e) and the results of the combined processing (f). Optimum values of the parameters correspond to the minimum of $E$. 

close to the broad-band estimates reported by Silver & Chan (1991) (81° and 1.1 s). There are some other examples of similarities in the estimates at different periods for the same locations, but there are no examples of serious unexplained disagreements.
the strike of the plate boundary. Seismicity in this region at depth around 50 km suggests compression in the direction orthogonal to the Himalayas (Chen & Molnar 1990).

Another case of similar relationship is the direction of anisotropy at SNZO which is located near the Alpine fault presenting the convergent boundary between the Indian and the Pacific plates. A similar relation between anisotropy and the plate boundary is observed at the west coast of South America, at stations BOCO and ZOBO. It should be mentioned, however, that the situation at ZOBO is complicated by the extensional regime in the crust at that portion of the Andes (Zoback et al. 1989).

Figure 5. Radial (R) and transverse (T) components of SKS recorded at KONO: (a) event 17 of Table 2; (b) event 19; (c) event 58; and (d) event 64. Vertical bars show the time interval used for the analysis.

Figure 6. The same as in Fig. 5 for ANTO: (a) event 35; (b) event 39; and (c) event 64.
The trend of the fast polarization direction in the regions of present-day plate convergence to be aligned with the plate boundary was previously noted by Milev and Vinnik (1991), Silver & Chan (1991) and Vinnik et al. (1991) but the number of documented cases in each of these studies was much smaller than in the present compilation. The simplest explanation of this alignment is provided by the results of deforming ultramafic rocks by uniaxial compression (Christensen & Crosson 1968). In that case the axis of symmetry in olivine is (b) and, assuming horizontal shortening direction, the polarization direction of the slow split wave in SKS nearly coincides with the direction of shortening. This means that the direction of shortening in the upper mantle of the presently active zones of convergence is usually similar to that in the crust.

A clear exception to these patterns is found at station CHTO in Indochina, where the direction of fast velocity is almost orthogonal to the plate boundary and the Indoburman ranges. A complicated structure is also found in Japan (stations INU and MAJO).

The other group of stations is located in regions of rifting. This group includes AGD (the Red Sea rift), ZAK (the Baikal rift), TIK (continental extension of the Hakkel ridge in the Arctic Ocean), ANMO (the Rio Grande rift), NE5 and NE15 (the lower Rhine graben). At AGD, ZAK, TIK, NE5 and NE15, the fast direction of anisotropy is oriented nearly perpendicular to the rifts and parallel to the directions of extension in the crust. At ANMO, the fast direction of anisotropy forms an angle of nearly 60° with the predominantly E–W direction of extension in western North America (Zoback & Zoback 1980). A similar complexity is probably observed at SSB (see Section 4.5). We conclude that in the majority of cases the inferred directions of flow in the mantle are close to the extension directions in the crust. This finding is in agreement with some models of continental rifting (e.g. Werner & Kahle 1980).
There is a coincidence between the fast directions of anisotropy at KONO and KEV and the strike of the Caledonides in Scandinavia. This correlation is less clear in the Appalachian belt in North America where the direction of fast velocity at WFM and SCP forms an angle of 30° with the strike of the belt. The direction of anisotropy at these stations is similar to those at RSON and GAC within the Canadian shield. The direction of anisotropy at SVE in the Urals coincides with the strike of the Urals, but the same directions are found at OBN and NRI, in the Precambrian East European and Siberian platforms, respectively. The situation is further complicated by the observation of a different direction at station ARU, 150 km to the west of SVE (Silver & Chan 1991). The direction of anisotropy at SEM is close to the strike of the Palaeozoic Altay mountains (Belousov 1978), but, again, similar directions are observed at some other stations in Siberia. The predominantly E–W direction of fast velocity in central and western Europe could be related to the Hercynian orogeny, though, as shown by Bormann et al. (1991), this is not a very likely possibility. Silver & Chan (1988, 1991) proposed that anisotropy in central and eastern North America is frozen since the Precambrian. As will be shown later, there are strong doubts about the uniqueness of this interpretation.

4.2 Shear wave splitting and absolute plate motions

The map in Fig. 1 shows several tectonically stable regions with a relatively uniform direction of fast velocity. The directions changing between E–W and NE–SW are observed in eastern and central parts of North America, in Africa and in Australia. E–W direction is dominant in central and western Europe, and N–S direction prevails at many stations in Russia. In some of these regions (Australia and Africa) there is a good correspondence between the directions of fast polarization in SKS and the directions of fast velocity of long-period surface waves, as given by Montagner & Tanimoto (1991). In CIS, among tectonic units with the predominant N–S direction of fast velocity there are Precambrian East European and Siberian platforms (OBN and NRI, respectively), Precambrian Kokchetav massif (BRV), Palaeozoic zones of convergence (SVE and SEM), Mesozoic Okhotsko–Chukotsky volcanic belt (MGD) and Mesozoic tectonic collage of Chukotka (ILT). Heterogeneous structure is also evident in other regions with uniform directions of anisotropy. These observations suggest the presence of a large-scale process in the mantle which is responsible for at least a significant part of the effect. The obvious candidate process is the lithospheric plate motion with respect to the deeper shells. In the central parts of some plates there is a correlation between the direction of maximum horizontal stress and the direction of absolute plate motion (Zoback et al. 1989). This correlation is especially strong in mid-plate North America suggesting that resistive drag at the base of the plate is one of the major sources of stress. Such a drag, if present, could orient the olivine crystals in the direction of absolute plate motion. In fact, the dominant direction of fast velocity of anisotropy in central and eastern portions of North America is in agreement with the direction of absolute plate velocity, as given by Minster & Jordan (1978).

Figure 8. Plots of $E(\alpha, \delta t)$ corresponding to the records in Figs 6(a)–(c) and the results of the combined processing (d).

A few stations are located in regions of strike-slip faulting: SCZ, near the San Andreas fault; ANTO, at a distance of 100 km from the North Anatolian fault, and SEY in the area of Moma rift in eastern Siberia (for a description of this rift and other tectonic features in Russia see e.g. Zonenshein, Kuzmin & Natapov 1990). Only at SEY is the fast direction of anisotropy close to the direction of shear in the crust. A likelihood of causal relation between the fault and the direction of anisotropy at SEY is enhanced by the observation that at 300 km to the south (station MGD) the fast direction of anisotropy differs by 90°. The data of this group suggest that the maximum depth of shear at some faults is too small to produce a measurable effect in the upper mantle and/or this deformation is concentrated in a very narrow zone around the fault. This conclusion is supported by the observation (Makeyeva et al. 1992) that the motion along the Talasso–Fergana strike-slip fault in the Tien Shan does not affect the directions of anisotropy at neighbouring stations.

Proceeding to the zones of convergence of the distant geologic past we sometimes observe, though not so clearly, correlations similar to those in the presently active regions.
The differences between the fast directions of anisotropy and the absolute velocity directions of the plates are displayed in a map (Fig. 9) and also shown in a histogram (Fig. 10). The data used in the histogram have special notations in Table 1 (the corresponding value of the plate velocity direction). The plate velocities were taken from Minster & Jordan (1978). The data from active tectonic regions were excluded because they clearly depended not so much on the absolute plate velocity directions as on the interactions between the plates. The data of lower quality which are shown in Fig. 1 by open circles were excluded as well. The density of measurements in a part of central Europe is very high in comparison with the rest of the world. For a more uniform representation, only two measurements were taken from that region. The resulting histogram in Fig. 10 suggests a strong positive correlation between the fast velocity directions of anisotropy and absolute velocity directions of the plates: nearly 85 per cent of angular differences between the two directions are in the range from $-45^\circ$ to $+45^\circ$. The correlation is especially strong in North America (see Fig. 9). For the rest of the world, the percentage of data in this range is somewhat smaller than for the whole data set (75 per cent).

Surprisingly, in spite of the correlation shown in Fig. 10, there is no clear correlation between the magnitudes of anisotropy ($\delta r$) and the absolute plate velocities. The bulk of data contributing to the peak in the histogram between $-45^\circ$ and $+45^\circ$ comes from North America and northern Eurasia. The average values of $\delta r$ in both regions are nearly the same. At the same time, North America is one of the fastest plates whereas Eurasia is the slowest. A lack of reliable measurements of anisotropy in Australia is explained, at least partly, by small values of $\delta r$ in the south-east of the region. In particular, the effect of anisotropy at station CAN (Canberra) was so small that it could not be detected (Vinnik et al. 1989a). On the other hand, the Indian–Australian plate is among the fastest in the world.

4.3 Shear wave splitting and teleseismic P traveltime residuals

There are data (e.g. Dziewonski & Anderson 1983; Wyliegalla, Bormann & Baumback 1988) which indicate the presence of harmonics with the period of 180° in the azimuthal P traveltime variations. These data were interpreted in terms of azimuthal anisotropy inside the
4.4 Shear wave splitting and long-range refraction

 earth. If azimuthal anisotropy in the upper mantle, as commonly assumed, is due mainly to crystal alignment of olivine, then the direction of polarization of the fast wave in SKS must coincide with the fast direction of the teleseismic P-waves. To check this, we compared both directions for the stations with reliable determinations of the directions of polarization of the fast wave. The histogram thus obtained (Fig. 11) demonstrates absence of a significant correlation. We suspect, that the estimates of fast direction for the teleseismic P-waves are strongly affected by the lateral heterogeneity of the earth and reading errors.

4.4 Shear wave splitting and long-range refraction profiling data

Observations of P-waves on profiles of sufficient length may provide valuable data on azimuthal anisotropy in the upper mantle. Resolution of these observations with respect to lateral variations of azimuthal anisotropy is, clearly, very low in comparison with that of SKS. However, in regions with established near-uniform direction of azimuthal anisotropy, like in the eastern part of North America (see Fig. 1), the problem of lateral resolution is of secondary importance. At the same time, in a region like that, the data on azimuthal traveltimes variations near 10 s. However, the azimuthal variations that were observed in the eastern part of North America in the Early Rise long-range profiling experiment, if present, were smaller by at least an order of magnitude (Iyer et al. 1969). This comparison suggests that the layer responsible for the large-scale component of anisotropy in North American is underneath the lithosphere. The possibility of frozen anisotropy in the subcrustal lithosphere is by no means excluded but it must be a smaller scale phenomenon. Then, for example, the unusually high value of $\delta t$ at RSON (1.7 s) could be due to coincidence of the directions of frozen (lithospheric) anisotropy and present-day asthenospheric anisotropy. In a similar way, at some other stations the observed $\delta t$ can be anomalously small because the directions of lithospheric and asthenospheric anisotropies are at right angles with each other.

Similar conclusions about the depths responsible for continental anisotropy can be drawn from long-range profiling data in Russia. The observations in the East European platform provide some evidence of azimuthal anisotropy in the depth range between 50 and 150 km (Vinnik et al. 1990). This depth range, however, is too small to account for the effect of anisotropy in SKS. The data suggest that anisotropy is concentrated in a few layers rather than distributed continuously. Possible effects of azimuthal
anisotropy in the long-range profiling data for the Siberian platform were described by Vinnik & Yegorkin (1980). At present this interpretation is supported by the observations of SKS. According to the SKS data, the dominant direction of fast velocity in that region is approximately N–S. In the profile Norilsk–Tiksi going E–W the P-waves arrive systematically later than at the profile Dikson–Hilok, going in the direction 140° to the North (Burmakov et al. 1987). The largest differences in the wave velocities are observed at depths exceeding 150 km.

In central and western Europe, where the dominant direction of anisotropy is E–W (Fig. 1), the data of seismic refraction profiling in Germany indicate the azimuth of fast velocity near 20° (Bamford 1977; Fuchs 1983). According to Farra et al. (1991), the bottom of this layer is at 55 km. The bottom depth of the thermally defined lithosphere in the central part of Europe is less than 100 km (Cermak 1982), implying that the layer where anisotropy with the E–W direction can be frozen is less than 50 km thick. This makes the explanation of the E–W trend in the European data in terms of frozen anisotropy even more difficult than in North America.

Finally, Maupin & Cara (1991) inverted observations of long-period surface waves for anisotropy in the crust and mantle of the Iberian platform. In their model which is in agreement with the SKS data for station TOL, the anisotropic layer is located at depths between 100 and 400 km. No anisotropy is found in the lithosphere.

4.5 Observations of special interest

There are a few SKS data deserving special attention. We already mentioned station CHTO where the direction of anisotropy is nearly orthogonal to the convergent plate boundary. An unexpected confirmation of the reality of this anomaly is found in the fault-plane solutions for the earthquakes presumably within the Indian plate being subducted eastward beneath the Indoburman ranges (Chen & Molnar 1990). The maximum compressional strain inferred from these solutions is oriented roughly in the S–N direction which is consistent with the roughly E–W direction of extention inferred from the direction of anisotropy. Chen & Molnar (1990) suggest that the deformation in the Indoburman ranges is decoupled from that in the underlying Indian plate. Unambiguous interpretation of these observations is difficult with the available data but we think that they could be explained as the mantle flow out of the zone of convergence underneath the Himalayas. This direction seems to be reasonably close to the direction of extrusion of Indochina to the south–east due to the penetration of India into Eurasia (Tapponier et al. 1982). However, apparently, in a similar situation at ANTO, where central Turkey is moving to the west (Sengor 1979), this motion does not show up in the direction of anisotropy.

The other observation deserving special attention is at station SSB where the direction of anisotropy is in a sharp disagreement with the dominant E–W direction at many other stations in the central and western parts of Europe. It is possible that the anomaly is related to the rifting in the French Massif Central. The rift is oriented roughly in S–N direction. The direction of anisotropy at SSB, if related to the rifting, suggests that the mantle flow is nearly parallel to the rift. There are some other indications of this direction of flow (Ramananantoandro 1988). Similar flow could probably explain the anomalous direction of anisotropy in the Rio Grande rift area (station ANMO).

Station KIP is located within the Hawaii hotspot. Vinnik et al. (1989a) noted that the direction of anisotropy at KIP (45°) is strongly different from the present-day Pacific plate motion direction (120°) but is reasonably close to the fossil spreading direction (70°) with the age around 100 Ma. This observation implies that the effect is due to frozen anisotropy in the subcrustal lithosphere. However, the magnitude of the observed anisotropy is large (near 1.5 s) requiring an enormously thick subcrustal lithosphere, whereas the depth of the threshold of 900°C in the hotspot is strongly reduced. Therefore, we propose that the direction of anisotropy at KIP could be due to flow in the mantle plume which is associated with the hotspot. The flow is vertical in a thin conduit connecting the plume with the source, but it becomes near-horizontal in the inflating part of the diapir.

Unfortunately, every anomaly is sampled by only one station. Without dense seismograph networks we can not determine the dimensions of the anomalies which are of crucial importance for interpreting the data.

5 CONCLUSIONS

The data presented in this paper indicate persistent presence of azimuthal anisotropy in the continental mantle. Published seismic data indicate that the magnitude of the P-wave azimuthal anisotropy in the upper mantle is between 0 and 30 per cent of that expected in the model with perfectly aligned crystals. The same ratio can be assumed for shear wave anisotropy. In the pyrolite model of the upper mantle with perfectly aligned crystals the difference between the velocities of the quasi-shear waves is close to 7 per cent. With the assumption of δνp = 2 per cent which corresponds to the upper limit of the ratio, the thickness of the layer responsible for anisotropy with δv = 1 s is near 250 km. That is, probably, the lower limit for the thickness of the layer which, hence, includes the whole asthenosphere.

In the regions of present-day tectonic activity we observe a strong positive correlation between the deformations in the crust and the upper mantle, though there are some remarkable exceptions.

In stable regions there is a positive correlation between the absolute plate velocity directions and the fast directions of anisotropy. This finding is compatible with the conclusion that a significant contribution to the observed shear wave splitting comes from the asthenosphere. Apparently, the anisotropic ‘roots’ formed during the active tectonic episodes are partially destroyed by subsequent processes.

The idea of strong large-scale azimuthal anisotropy being frozen in the subcrustal lithosphere of the old continental platforms since the Precambrian is in strong contradiction with the data of long-range refraction profiling. A possibility of frozen anisotropy in the upper mantle of the old and stable continental regions is by no means excluded. However, if the directions of anisotropy in the lithosphere and the asthenosphere are different, the direction of the resulting anisotropy in SKS is intermediate between the two. Then a similarity between the directions of anisotropy and
of crustal fabric at a given station can no longer be regarded as a strong evidence that the observed anisotropy is frozen.

In the future, we need more data everywhere but, especially, in the southern hemisphere. A dense coverage of the continents by the measurements of shear wave splitting could help to separate contributions from the present-day flow in the asthenosphere and frozen anisotropy in the subcrustal lithosphere, because the corresponding scales are probably strongly different. Very dense observational networks are needed in the regions of special interest, such as hotspots, zones of zoning and plate convergence.

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