Interpretation of SKS-waves using samples from the subcontinental lithosphere

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ABSTRACT

The seismic properties of five South African kimberlite nodules have been calculated from petrofabric measurements. These garnet lherzolite and harzburgite composition nodules (56–80% olivine) are known from previous studies to have originated at depths of 120–170 km in the subcontinental lithosphere. Their in situ seismic properties have been calculated by extrapolating the elastic constants to the appropriate temperature (900–1050°C) and pressure (3.0–3.5 GPa) conditions. The average of the five samples has a maximum S-wave anisotropy (δVp) of 3.7% which is sufficiently high to explain previously reported teleseismic SKS delay times (δt) in continental shield areas of up to 1.7 s. The biggest delay time is observed when the foliation is vertical, the lineation horizontal, and the fastest S-wave polarized parallel to the lineation, i.e. an orientation for transcurrent motion. Delay times of over 1 s can only be generated by this orientation. A strong mantle–crust mechanical coupling is suggested for such situations. The potential use of seismic anisotropy as an indicator of strain in the lithosphere has been investigated using experimental and simulated fabric data for olivine, the most abundant phase (70%) in the upper mantle. Recrystallization tends to reduce fabric strength and seismic anisotropy, resulting in saturation values for experimental and simulated fabrics which correspond to approximately 50–60% strain or 8–9% δVp for pure olivine aggregates. A survey of 25 naturally deformed peridotites of oceanic origin suggests an average maximum S-wave anisotropy for the olivine component of 9%, or about 8% for a lherzolite or harzburgite rock composition when the orthopyroxene component is taken into account. The ophiolite samples are twice as anisotropic as the kimberlite nodules. If an average anisotropy value is representative of the lithosphere, then the SKS delay times represent variations in anisotropic layer thickness for delay times over 1 s as the orientation is constrained to be foliation vertical and lineation horizontal. The magnitude of S-wave anisotropy is less sensitive than Vp to variations in olivine volume fraction in range 50–100%, and to deformation or fabric intensity, further suggesting that S-wave anisotropy is particularly apt for the determination of the anisotropic lithosphere thickness. For smaller delay times there is some trade off between structural orientation and thickness.

1. Introduction

Recent measurements of shear-wave splitting in SKS teleseismic phases at continental stations show that the azimuth of the fast polarization direction is most often parallel to the trend of mountain belts (Vinnik et al., 1984, 1989; Silver and Chan 1988, 1991; Savage et al., 1990). To clarify the interpretation of these measurements and their geodynamic implications we present an examination of the possible sources of seismic anisotropy and their orientation and magnitude. The delay time between arrivals of the two shear waves gives an empirical measure of the seismic anisotropy accumulated along the propagation path of the SKS phases and is typically between 0.5 and 2.0 s, being on average about 1 s (Silver and Chan, 1991). SKS phases are created by the P-wave to S-wave conversion at the core–mantle boundary (CMB) and have an essentially vertical propagation path from the CMB, through the mantle and crust, to the surface receiver station.
The steep arrival angle of such teleseismic waves beneath the receiver provides a good lateral resolution of about 50 km. Although SKS phases only provide a measure of the anisotropy along a one-dimensional near-vertical direction, the azimuth of fast polarization and delay time provides an unambiguous constraint on the orientation and magnitude of anisotropy. The SKS measurements are made directly from the three-component broad-band seismograms recording and do not require extensive data sets and computations (e.g. surface-wave anisotropic tomography; Nataf et al., 1984; Montagner and Tanimoto, 1990, 1991) or station-dependent normalization procedures (e.g. P-wave travel time residuals; Babuška et al., 1984; Poupinet et al., 1992) needed in other techniques.

2. Seismic anisotropy along the SKS propagation path

The source of the seismic anisotropy, along the propagation path from the CMB to the surface cannot be determined from SKS observations. Some constraints on the depth of the anisotropy are imposed by the station spacing, typically the less than 1000 km from the surface for permanent stations. The SKS-waves traverse the lower mantle (ca. 2100 km of path length), the transition zone (270 km), the upper mantle (370 km) and the continental crust (ca. 30 km) on their way to the surface. All of these layers could contribute to the measured delay time and polarization. The anisotropy of the lower mantle appears to be very small as shown by SKS recordings at the same station using earthquake sources of different back-azimuths. Two different lower-mantle propagation paths show nearly identical delay times and polarizations (Kaneshima and Silver, 1992).

From the petrological point of view the lower mantle can be assumed to be essentially a mixture of perovskite (80% by volume) and wüsite (20%) (e.g. Ringwood, 1991). We have calculated the maximum S-wave anisotropies ($\delta V_s$) for perovskite and iron wüsite using the single elastic constant data of Yeganeh-haeriet al. (1989) and Sumino et al. (1980) which give 9% and 13%, respectively. Unfortunately, very little is known about the petrofabrics of these materials, a preliminary report of silicate perovskite deformed in a diamond anvil cell by Meade and Jeanloz (1990) suggests that no petrofabric developed. Karato and Li (1992) have conducted deformation experiments on synthetic aggregates of an oxide ($\text{CaTiO}_3$) structurally analogue to the silicate perovskite. The tests show that the fine-grained analogue deforms by a grain boundary sliding mechanism which does not produce a fabric. However, most fine-grained materials exhibit such behaviour at high temperature, hence one cannot consider this characteristic to be especially intrinsic to perovskite. If these experiments do indeed reproduce the deformation behaviour of mantle perovskite then no seismic anisotropy is to be expected in the lower mantle despite the 2100 km path length.

The transition zone, so-called because there are several phase transitions in this depth interval, has a more complex mineralogy than the lower mantle. The volumetrically important minerals are $\beta$-(Mg, Fe)$_2$SiO$_4$, sometimes incorrectly called modified $\beta$-spinel although it is not a true spinel structure (L. Finger, personal communication, 1992), and $\gamma$-(Mg, Fe)$_2$SiO$_4$ ($\gamma$-spinel) which occupy about 60% of the volume and majorite garnet (30%). These minerals have maximum S-wave anisotropies ($\delta V_s$) of 14% for $\beta$-(Mg, Fe)$_2$SiO$_4$, 2% for $\gamma$-(Mg, Fe)$_2$SiO$_4$, and 1% for majorite garnet which have been calculated from the single crystal elastic constants given by Sawamoto et al. (1984), Weidner et al. (1984) and Bonczar et al. (1977), respectively. The constants given by Bonczar et al. (1977) are for garnet; no single crystal values are currently available for majorite garnet. $\beta$-(Mg, Fe)$_2$SiO$_4$ is the dominant phase in the upper 70 km of the transition zone, whereas $\gamma$-spinel is the dominant phase in the bottom 130 km according to the phase diagrams by Ringwood (1991). Hence, only the 70 km thick layer dominated by $\beta$-(Mg, Fe)$_2$SiO$_4$ could significantly contribute to the SKS delay time. There are no petrofabric measurements of $\beta$-(Mg, Fe)$_2$SiO$_4$. From our own observations on olivine aggregates we find that the seismic anisotropy is at most about 25% for $V_s$ of the
single crystal value, hence we expect a maximum anisotropy for $\beta-(\text{Mg}, \text{Fe})_2\text{SiO}_4$ aggregate of about 3%. A 3% anisotropy would give 0.2 s maximum delay time for a 70 km layer at an average $V_p = 5.5$ km s$^{-1}$ and assuming the $\beta-(\text{Mg}, \text{Fe})_2\text{SiO}_4$ to be about 60% by volume. Clearly the delay time contribution from the transition zone is well below the 0.5–2.0 s observed for SKS.

The upper mantle is known to be seismically anisotropic from $P_n$ velocity measurements in the ocean basins (e.g. Raitt et al., 1969; Shearer and Orcutt, 1986) and azimuthal and polarization anisotropy in long-period surface waves (e.g. Anderson, 1961; Aki and Kaminuma, 1963; Nataf et al., 1984; Montagner and Tanimoto, 1990, 1991). In particular, the top 200 km of the subcontinental mantle has been shown to be anisotropic. A 5% $S$-wave anisotropy is sufficient to give a delay time of 2 s with a 200 km layer of upper mantle composed of 70% olivine and 30% pyroxene. In this paper we demonstrate, using kimberlite nodule samples from the subcontinental lithosphere, that such a material in the appropriate orientation is capable of explaining the observed SKS delay times for a lithospheric thickness of 150–250 km, and the tectonic significance of the azimuth of the fast polarization.

Lastly we should consider the possible contribution of the continental crust to SKS delay time. Shear-wave splitting measurements for the entire crust (Kaneshima et al., 1987; Kaneshima et al., 1989; McNamara et al., 1989) yield delay times in the range 0.1–0.3 s. From considerations of typical crustal compositions and measured petrofabrics Barruol and Mainprice (this issue) have estimated $S$-wave delay times to be 0.1 s per 10 km which is consistent with the seismic observations. Again such delay times cannot explain the SKS observations.

We estimate the maximum possible delay time contributions of the transition zone at 0.2 s and the crust at 0.1–0.3 s, both of which are small compared with the SKS values observed. In summary, the upper mantle seems to be the main source for the observed delay times of between 1 and 2 s. The observational seismology constrains the anisotropy to be within 1000 km of the surface from permanent station studies (e.g. Silver and Chan, 1991) or even closer to the surface from more closely spaced portable seismometer studies (e.g. Silver et al., 1989). Seismic records of the lower mantle also show no evidence for anisotropy. From petrophysical arguments, we also conclude that the transition zone and crust are unlikely to produce the observed delay times. So, despite the limited control on the location of anisotropy, SKS shear-wave splitting does provide a unique and unambiguous tool for exploring the magnitude and orientation of the anisotropy of the subcontinental lithosphere.

3. South African kimberlite nodules

3.1. Modal composition and microstructure

Measurements have been made on garnet lherzolite and harzburgite peridotite nodules from South African kimberlite pipes. The South African kimberlite nodules constitute the most extensively studied and numerically important collection of subcontinental mantle samples presently available (Nixon, 1987). All our samples come from kimberlite pipes in the Kaapvaal craton. The pipes have intruded along fracture zones that transect basement structures and intrude earlier Jurassic sediments and lavas. All the nodules have been dated at 120–80 Myear except specimens from the Premier mine which have a Precambrian age of about 1200 Myear. Approximately 90% of the recorded nodules have garnet lherzolite or harzburgite compositions with the remaining 10% being eclogites. It is now well established from petrological studies (e.g. Boyd, 1973) that these nodules come from depths of 120–205 km, and hence they represent a unique direct sampling of the subcontinental lithosphere. Chemical analysis of heavy minerals by Schulze (1989) of upper-mantle nodules led to the conclusion that the peridotite nodules were representative of 99% by volume of the upper 200 km of the subcontinental mantle.

In this study we focused our attention on the granular nodules which are typical of the subcontinental lithosphere. Granular nodules (KBBF20, M57 and KBBF4) are thought to have had a
two-stage mechanical history with large strain ductile deformation and subsequent annealing in the upper mantle (Gueguen, 1977; Mercier, 1979) prior to a rapid eruption. The porphyroclastic nodules (KBJ17 and LMA 2.5) may be less representative of the lithosphere, but have similar petrofabrics. The sheared nodules with mylonitic (mosaic) microstructures have been excluded as their origin has remained controversial over the years, representing either the sheared base of the lithosphere (Boyd, 1973), a kimberlite diapir source region (Green and Gueguen, 1974, 1983), deformation minutes before the eruptive process (Goetze, 1975; Mercier, 1979), or sheared subduction products (Kesson and Ringwood, 1989). In any event they do not represent the bulk of the subcontinental mantle likely to be sampled by teleseismic SKS-waves. The specimens studied here have been previously described by Boullier and Nicolas (1975) and their general characteristics are given in Table 1.

Specimen KBBF20 is a spinel harzburgite 55.7% olivine, 43.7% enstatite, 0.6% symplectite of clinopyroxene and red-brown spinel around the enstatite. The grain size is between 5 and 10 mm. The foliation is defined by the olivine grain shape and the lineation by the elongate enstatite grains. The olivine grains are cut by pull-apart fractures normal to the lineation. The olivine crystals have straight (100) subgrain boundaries and dislocation densities of about $10^6$ cm$^{-2}$ (Gueguen, 1977).

Specimen M57 is a garnet harzburgite 66% olivine, 29.8% enstatite, 4% garnet, 0.2% phlogopite. The grains have a tabular shape with a grain size of 2.5–5.0 mm. Euhedral tabular or ‘tablet’-shaped grains are characteristic of fluid-assisted grain boundary migration (Drury and Van Roermund, 1989). The lineation is clearly defined by the long axes of the grains. No optical strain features were observed and the dislocation densities in the olivine grains are less than $10^7$ cm$^{-2}$ (Gueguen, 1979).

Specimen KBBF4 is a garnet harzburgite 68% olivine, 27% enstatite, 2.4% garnet, 2% clinopyroxene, 0.5% phlogopite and 0.1% spinel. The grains are equant in shape with grain size of 5–20 mm with olivine being much larger than enstatite. The foliation and lineation are defined by the enstatite grain shape. Both olivine and enstatite have undulose extinction.

Specimen KBJ17 is garnet lherzolite 70.1% olivine, 22% enstatite, 7.75% garnet, 0.15% clinopyroxene. A strong foliation is defined by the enstatite grain shape and the lineation by elongate garnet grains. Three types of olivine crystals can be distinguished. First, strongly deformed porphyroclasts with closely spaced sub-

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**TABLE 1**

The characteristics of South African kimberlite xenoliths

<table>
<thead>
<tr>
<th>Reference number</th>
<th>Location</th>
<th>Microstructure</th>
<th>Composition (%)</th>
<th>Pressure (GPa)</th>
<th>Temperature (°C)</th>
<th>Depth * (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>KBBF20</td>
<td>Bulfontein (Kimberly)</td>
<td>Large grains (granular)</td>
<td>56.0 olivine, 44.0 enstatite</td>
<td>3.5</td>
<td>900</td>
<td>120</td>
</tr>
<tr>
<td>M57</td>
<td>Monastery</td>
<td>Tabular</td>
<td>66.0 olivine, 30.0 enstatite, 4.0 garnet</td>
<td>3.5</td>
<td>900</td>
<td>120</td>
</tr>
<tr>
<td>KBBF4</td>
<td>Bulfontein (Kimberly)</td>
<td>Tabular</td>
<td>70.0 olivine, 27.0 enstatite, 3.0 garnet</td>
<td>3.5</td>
<td>900</td>
<td>120</td>
</tr>
<tr>
<td>KBJ17</td>
<td>Jagersfontein</td>
<td>Porphyroclastic</td>
<td>70.2 olivine, 22.0 enstatite, 7.8 garnet</td>
<td>3.0</td>
<td>1000</td>
<td>140</td>
</tr>
<tr>
<td>LMA 2.5</td>
<td>Matsoku (Lesotho)</td>
<td>Secondary large grains</td>
<td>62.0 olivine, 33.0 enstatite, 5.0 garnet</td>
<td>3.5</td>
<td>1050</td>
<td>170</td>
</tr>
</tbody>
</table>

* According to the pyroxene geotherm of Boyd (1973).
grain boundaries and high dislocation densities
$10^7 - 10^9$ cm$^{-2}$ (Green, 1976; Gueguen, 1977). Secondly, polygonal recrystallized grains with a
grain size similar to that of the subgrain size in
the porphyroclasts with a dislocation density of
$5 \times 10^7$ cm$^{-2}$ (Boullier and Gueguen, 1975).
Thirdly, some euhedral tablet grains growing at
the expense of the porphyroclasts. The garnet
grains are cut by pull-apart fractures which are
perpendicular to the lineation.

Specimen LMA2.5 is garnet lherzolite 61.2%
olivine, 33.2% enstatite, 5% garnet, and 0.6%
clinopyroxene with a strong foliation defined by
the grain shapes of the enstatite and garnet. Olivine
forms bands of large grains (2–3 mm)
with no deformation features and 120° triple
points alternating with bands of finer grained (0.5
mm) enstatite and garnet. Some olivine grains
have inclusion trails of enstatite and garnet which
are parallel to the foliation. Enstatite occurs as
deformed porphyroclasts and undeformed small
grains. Both garnet and clinopyroxene are de-
formed.

3.2. Petrofabrics

All the petrofabrics reported in this paper
have been measured using an optical petrological
microscope equipped with a 5 axis universal stage.
The measurements on the kimberlite nodules
were undertaken by Boullier (1975). The original
orientation data have been converted to the three
Euler angles ($\varphi_1, \phi, \varphi_2$) of Bunge (1982) which
define the orientation matrix $g$ describing the
rotation of the crystal from some standard orienta-
tion ($\varphi_1 = 0^\circ$, $\phi = 0^\circ$, $\varphi_2 = 0^\circ$) to its measured
orientation. In a rock with many grains of various
orientations, the ensemble can be described by an
Orientation Distribution Function (ODF), which
is a function $f(g)$ that quantitatively describes
the volume fraction of the sample with an orient-
tation $g$. The function $f(g)$ is given in terms of
symmetrical generalized spherical harmonics
(Bunge, 1982) as

$$f(g) = \sum_{l=0}^{L_{\text{max}}-1} \sum_{m=-l}^{l} \sum_{n=-1}^{1} C_{l,m}^{m n} T_{l,m}^{m n}(g)$$

where $C_{l,m}^{m n}$ are the coefficients of the series de-
development of the texture function $f(g)$, and

$T_{l,m}^{m n}(g)$ are the generalized spherical harmonic
functions, $M(l)$ is the number of linearly inde-
hendent harmonics and $L_{\text{max}}$ is the maximum
degree used in the expansion.

The harmonic method has been applied to all
petrofabric or texture measurements presented
here. In the series development method the index
$L$ may be evaluated to infinity. In practice it has
to be truncated at some value of $L$ called $L_{\text{max}}$
(see Wagner et al., 1981); in the present case
$L_{\text{max}} = 22$. Once the $C_{l,m}^{m n}$ coefficients have been
calculated, they can be used to calculate the pole
figures and many other types of representation.
In the present application, we required some
standard method to define fabric strength. In
quantitative texture analysis (Bunge, 1982), the
strength or power of a fabric can be defined as

$$J = \int f(g)^2 dg$$

where $J$ is the $J$ index and $dg = 1/8\pi^2 \sin\phi \ d\phi_1$
$\ d\varphi_1 \ d\varphi_2$. In general the $J$ index, a dimensionless
number, should be used with caution as the ODF
may be a complex distribution with various ele-
ments of different densities. However, in the case
of naturally deformed olivine, the ODF are very
similar (Pouthier, 1989) and its use in the present
application is justified.

In the following description we present the
fabric (texture) data as pole figures as this repre-
sentation is the most familiar to Earth Scientists.
The right-handed orthogonal reference frame is
the lineation ($X$), the normal to the foliation
plane ($Z$) and the normal to these two directions
($Y$). All the olivine fabrics have the same general
characteristics (Fig. 1). The [100] pole figure has a
point maximum which is adjacent to the grain
shape lineation ($X$). The [010] pole figures has a
single point maximum perpendicular to the folia-
tion ($Z$). Similarly, the [001] pole figures has a
single point maximum perpendicular to the folia-
tion ($Z$). Similarly, the [001] pole figures have
single point maxima in the foliation plane and
normal to the lineation ($Y$). The exception is
M57 with a maximum of [001] poles being almost
parallel to the lineation ($X$). The overall simplic-
ity of these pole figure distributions results in
strong fabric patterns.

The enstatite fabrics also are simple (Fig. 2),
being composed of point maxima. All samples
have [100] pole figures with the strong point maximum normal to the foliation. Samples M57 and KBJ17 have some [100] normal to the lineation and in the foliation (Y). All the samples have [010] pole figures with a point maximum normal to lineation and within, or close to, the foliation plane. All the samples have [001] pole figures with a point maximum parallel to the lineation (X), except KBBF20 where the maximum is in the foliation and normal to the lin-

Fig. 1. Pole figures of olivine for [100] [010] and [001]. Equal area projections with the lineation (X) east–west and the foliation plane (XY) perpendicular to the projection plane. Contours at intervals of 1 times uniform distribution; the lowest contour (×1) is a dashed line, higher values are solid lines. The location of the highest value is represented by a black square.
The typical fabric pattern for enstatite is [100] normal to the foliation (Z), [010] in the foliation plane and normal to the lineation (Y) and [001] parallel to the lineation (X).

3.3. Seismic properties

The seismic properties of rock can be calculated from the modal analysis, the single crystal

ORTHOPYROXENE POLE FIGURES

[100] [010] [001]

KBBF20

M57

KBBF4

KB17

LMA2.5

Fig. 2. Pole figures of orthopyroxene (enstatite) for [100], [010] and [001]. Conventions as in Fig. 1.
elastic constants, density and crystallographic fabric of each component. In the case of kimberlite nodules, which are typically hand specimens and often friable, this technique is better adapted than direct experimental measurement. We have simplified the modal analysis of the nodules to the three principal components: olivine, enstatite and garnet (Table 1). The garnet fabric was not measured; however, as it is elastically isotropic, the elastic and seismic properties do not depend on orientation. The garnet contribution to the elastic properties of the rock were taken into account by using the single crystal elastic constants and its volume fraction. The single crystal constants of olivine (Kumazawa and Anderson, 1969), enstatite (Frisillo and Barsch, 1972) and garnet (Bonczar et al., 1977) were extrapolated to upper-mantle pressures and temperatures using the following relationship

\[ C_{ijkl}(P, T) = C_{ijkl}(P_0, T_0) + (\frac{dC_{ijkl}}{dP})(P - P_0) + (\frac{dC_{ijkl}}{dT})(T - T_0) \]

where \( C_{ijkl}(P, T) \) are the elastic constants at upper mantle pressure \( P \) and temperature \( T \), \( C_{ijkl}(P_0, T_0) \) are the elastic constants at reference pressure \( P_0 = 0.1 \text{ MPa} \) and temperature \( T_0 = 25^\circ\text{C} \), \( \frac{dC_{ijkl}}{dP} \) is the first-order pressure derivative, and \( \frac{dC_{ijkl}}{dT} \) is the first-order temperature derivative. In using the above relationship, we have assumed a linear variation of \( C_{ijkl} \) with temperature and pressure. The seismic velocities also depend on the density of the minerals at upper-mantle pressure and temperature which can be calculated using

\[ \rho(P, T) = \rho_0 + (\rho_0/K)(P - P_0) - \alpha \rho_0(T - T_0) \]

where \( \rho(P, T) \) is the density at upper-mantle pressure \( P \) and temperature \( T \), \( \rho_0 \) is the density at reference pressure \( P_0 \) and temperature \( T_0 \), \( K \) is bulk modulus, and \( \alpha \) is the coefficient of thermal expansion, assumed to be isotropic for the present purpose.

To calculate the seismic properties of a polycrystal, one needs to evaluate the elastic properties of the aggregate. In the case of an aggregate with a crystallographic fabric, the anisotropy of the elastic properties of the single crystal must be taken into account. For each orientation \( g \), the single crystal properties have to be rotated into the specimen coordinate frame using the orientation matrix \( g_{ij} \)

\[ C_{ijkl}(g) = g_{im}g_{jn}g_{ko}g_{lp}C_{mnop}(g^0) \]

where \( C_{ijkl}(g) \) are the elastic properties in sample coordinates, \( g = g_{ij} = g(\varphi_1, \phi, \varphi_2) \) the measured orientation in sample co-ordinates, and \( C_{mnop}(g^0) \) is the elastic property in crystal coordinates.

The elastic properties of the polycrystal may be calculated by integration over all possible orientations of the ODF. Bunge (1985) has shown that integration is given as

\[ \langle C_{ijkl} \rangle^m = \int C_{ijkl}^m(g)f(g)dg \]

where \( \langle C_{ijkl} \rangle^m \) are the elastic properties of the aggregate of mineral \( m \). We define the Voigt average of the rock for \( m \) minerals as

\[ \langle C_{ijkl} \rangle^{\text{Voigt}} = \frac{1}{m} \sum \langle C_{ijkl} \rangle^m \]

The final step is the calculation of the three seismic phase velocities by solving the Christoffel equation

\[ \text{Det} |\langle C_{ijkl} \rangle^{\text{Voigt}}X_jX_i - \delta_{ik}\rho V^2| = 0 \]

where \( X_jX_i \) are the direction cosines of the wave propagation direction, \( \delta_{ik} \) is the Kronecker delta and \( V \) is one of the three seismic phase velocities. The procedures used here can calculate the Voigt, Reuss or Voigt–Reuss–Hill elastic averages. Measurements of the Twin Sisters Dunite by Crosson and Lin (1971) have shown that the seismic velocities calculated with the Voigt average give the closest agreement between velocities calculated using petrofabric data and laboratory measurements, and hence the Voigt average has been used in all calculations presented here. The choice of the averaging method affects the absolute velocity, but not the anisotropy of the sample.

3.4. P-wave velocities

The calculated P-wave velocities are presented on contoured projections (Fig. 3) in the same
PHYSICS OF THE EARTH AND PLANETARY INTERIORS

P-WAVE VELOCITIES

Contours (km/s)
8.15
8.20
8.25
8.30
8.35

Contours (km/s)
8.40
8.50
8.60

Contours (km/s)
8.30
8.40
8.50
8.60

Contours (km/s)
8.20
8.30
8.40
8.50
8.60
8.70

KBBF20
M57
KBBF4
KB117
LMA2.5

Fig. 3. Contoured P-wave velocity distributions for rocks using modal compositions given in Table 1 on an equal area projection. Lineation (X) is east-west and the foliation plane (XY) horizontal. The lowest contour value is a dashed line, higher values are solid. The location of the highest value is represented by a black square. Values in kilometres per second.

orientation as the pole figures. All samples show the same velocity distributions with the $V_p$ maximum being close to the lineation (X) or more precisely parallel to the maximum of the [100] pole figure of olivine. $V_p$ anisotropy ($\delta V_p$), defined here as $\delta V_p = [V_{p\text{max}} - V_{p\text{min}}]/\langle V_p \rangle$ with $\langle V_p \rangle = [V_{p\text{max}} + V_{p\text{min}}]/2$ is a parameter commonly used in petrophysics. The maximum and minimum values may occur anywhere on the projection sphere, hence $\delta V_p$ is the maximum $V_p$ anisotropy that can be observed. In general, the geometry of a body sampled by a $V_p$ seismic experiment will be limited by geometrical considerations and anisotropy measured is limited to a planar surface. For example, $P_n$-waves sample the azimuthal anisotropy of the upper mantle in a horizontal plane close to the Moho. The $\delta V_p$ for these samples, which are expected to be typical of the lithosphere, range from 4.6% for KBBF20 to 7.0% for KBBF4; consequently, the $P_n$ anisotropy of the continental upper mantle will be less than these values.

3.5. S-wave splitting delay times

We have calculated the S-wave splitting delay time for each of the nodules using the relationship

$$\delta t = L \delta V_s / \langle V_s \rangle$$

where $\delta t$ is the delay time, $L$ is the seismic path length, $\delta V_s$ is the S-wave anisotropy in the propagation direction $\delta V_s = [V_{s\text{max}} - V_{s\text{min}}]/\langle V_s \rangle$ and the mean S-wave velocity $\langle V_s \rangle = [V_{s\text{max}} + V_{s\text{min}}]/2$. Note that the S-wave anisotropy ($\delta V_s$) is defined for a given propagation direction, and hence could be directly determined in a seismic experiment. In calculating $\delta t$ we have assigned an arbitrary path length of 200 km; this path length can be thought of as a model lithosphere thickness. The contoured projections of delay time (Fig. 4) are in the same orientation as the pole figures and $V_p$ projections. All the samples show a similar distribution of delay times with a broad maximum in foliation (XY plane) and normal to the lineation (Y). All the samples show a similar pattern. The maximum delay time varies from 1.3 s for KBBF20 to 3.1 s for LMA2.5.
4. Effect of strain and fabric intensity on seismic anisotropy

Do these anisotropy variations correspond to changes in composition, structural orientation or finite strain? In order to evaluate the role of finite strain we need to establish a relationship between finite strain and seismic anisotropy. A crystallographic fabric of a polycrystal may be caused by a variety of physical processes including heterogeneous crystallization, annealing, viscous and plastic deformation. In rocks of the upper mantle the importance of plastic deformation at high temperature for fabric development has been emphasized by many workers (e.g. Mercier, 1985; Nicolas and Christensen, 1987). The relationship between fabric and seismic anisotropy in mantle rocks is also now well established (Crosson and Lin, 1971; Baker and Carter, 1972; Peselnick and Nicolas, 1978; Nicolas and Christensen, 1987). Recent studies of mantle seismic anisotropy have revealed significant variations with depth (e.g. seismic tomography; Montagner and Anderson, 1989) and location (e.g. SKS; Silver and Chan, 1991). We will evaluate the variation of seismic anisotropy with strain using experimental, simulated and natural fabrics.

4.1. Experimental deformation

The fabrics developed in experimentally deformed synthetic dunites of Nicolas et al. (1973) have proved to be a valuable test case for fabric simulation models (Etchecopar and Vasseur, 1987; Ribe and Yu, 1991; Wenk et al., 1991). The synthetic dunites with an average grain size of 0.25 mm were deformed at 1.3–1.5 GPa, 1200–1300°C at a constant strain rate of about $10^{-5}$ s$^{-1}$ in a Griggs solid medium piston cylinder deformation apparatus. The samples were deformed by uniaxial shortening to strains up to 58% (Table 2). In each sample the strain was heterogeneous due to axial temperature gradients. The strain in a given region of a sample was locally determined using initially spherical bubbles in the olivine grains as finite strain markers. The fabric strength as measured by the $J$ index (Fig. 5(A)) increases with axial strain. All fabric measurements have been made on the original grains or porphyroclasts, except at 58% strain where in addition the recrystallized grains have been measured. The recrystallized grains have a $J$ index which is significantly lower than the porphyroclasts. The values for the $V_p$ anisotropy ($\delta V_p$) and the S-wave anisotropy ($\delta V_s$) also increase in a similar manner to the $J$ index as a function of axial strain. There is an approximately linear relationship between the $J$ index and seismic anisotropy for both P- and S-waves (Fig. 6(A)), with $\delta V_p$ being more sensitive to strain than $\delta V_s$.

4.2. Numerical simulation

The experimental study was limited to relatively small strains (58%) whereas strains in the upper mantle may well exceed several hundred per cent. The mode of deformation in the upper mantle is unlikely to be uniaxial shortening. Studies of natural deformation of peridotite massifs (e.g. Nicolas et al., 1971) and basalt xenoliths (Mercier and Nicolas, 1975) show a strong non-coaxial deformation, which we can model as simple shear. In the absence of experimental data on the fabric evolution of olivine in simple shear we have used a numerical simulation of fabric development by Etchecopar and Vasseur (1987). The simulation has two important and original aspects for the present application; first, it is capable of modelling minerals with less than the five independent slip systems required by Von Mises criterion for a general constant volume deformation, and secondly it has the possibility of introducing an effective approximation to recrystallization. Olivine is one of those minerals that does not have five independent slip systems. From the experimental work of Nicolas et al. (1973) it is clear that recrystallization is an important process in the fabric evolution of olivine. Recrystallization is modelled by resetting the grain boundary coordinates to their initial orientations after some specified strain. For axial shortening, we have chosen 30% strain which corresponds to the strain at which Nicolas et al. (1973) started to observe recrystallization in their samples. In the first series of simulations for axial shortening we have either one slip system $[100](010)$ or two $[100](010)$
and [001](100)]. The [100](010) is the high-temperature system and [001](100)] the low-temperature system in olivine. The simulation and the experimental data have a similar trend with the $J$ index increasing with axial strain (Fig. 5(A)). Ideally the starting fabric for the simulation should

![KIMBERLITE NODULES - S-WAVE DELAY TIME](image)

Fig. 4. S-wave anisotropy ($\delta t$) parameter distributions for the rocks on an equal area projection. Conventions as in Fig. 3. Values in seconds.
be random and thus have by definition a $J$ index of 1. However, with only 210 grains being used in the simulation it is difficult to achieve a truly random distribution with such a small population. Interestingly, the starting fabric of the hot pressed dunite is not random ($J = 3.5$); the initially high $J$ index of the experimental samples may explain why the experimental $J$ values always remain significantly higher than the simulation values.

In simple shear we have specified recrystallization at intervals of shear strain $\gamma = 2$ as observed for ice experimentally deformed in torsion at high temperature by Bouchez and Duval (1982), as no experimental data are available for olivine. The $J$ index has been plotted (Fig. 5(B)) as a function of the principal finite strain axis $(1 + e_l)$ (e.g. Nye, 1957) of the strain ellipsoid; this parameter allows a comparison between axial shortening and simple shear. In simple shear, the $J$ index increases more rapidly with strain than for axial shortening and the difference between one and two slip systems is more marked. Two slip systems develop higher $J$ index values for a given strain than one.

4.3. Natural deformation

We have shown above that the experiments of Nicolas et al. (1973) provide a useful calibration of the $J$ index as a measure of fabric intensity. Using the olivine fabric data, we can also calculate the seismic properties as a function of strain and fabric intensity ($J$) (Fig. 6(A)). The seismic anisotropy of both $V_p$ and $V_s$ increases almost linearly with fabric intensity ($J$). A similar plot (Fig. 6(B)) for 25 naturally deformed peridotites and dunites also has a near-linear correlation between seismic anisotropy and fabric intensity.
although the scatter is greater. Here we used fabric intensity \((J)\) rather than strain, as we cannot measure the strain of the naturally deformed samples. We have drawn a least squares fitted straight line through both data sets (Figs. 6(A) and (B)) which describes better the experimental data \((r^2 = 0.90\) for \(V_p\) and 0.88 for \(V_s\)) than the naturally deformed olivine data \((r^2 = 0.48\) for \(V_p\) and 0.58 for \(V_s\)). For the naturally deformed olivine data, the anisotropy increases less rapidly with \(J\) index above \(J \approx 12\), suggesting a saturation of the seismic anisotropy at high values of \(J\) index. A frequency diagram for the 25 samples of fabric intensity \((J)\) (Fig. 6(C)) and \(V_s\) anisotropy (Fig. 6(D)) shows an asymmetric distribution with average values of \(J = 9.4\), \(\delta V_p = 11.7\%\) and \(\delta V_s = 9.0\%\). If we use the calibration curve given by the experiments of Nicolas et al. (1973) for the relationship between \(J\) index and strain (Fig. 5(A)), we deduce an average strain of 60\% shortening for the 25 samples, corresponding to a shear strain of 1. Many problems are encountered in extrapolating these average numbers to natural deformation. The deformation mode of the experiments was axial shortening, whereas in natural conditions a dominant non-coaxial or shearing component is observed. The slip system activity reflecting, in particular temperature conditions (Fig. 5(B)), may be different, probably compromising such deductions. Recrystallization was present in the experiments and proved very important in reducing the seismic anisotropy at higher strains. Recrystallization is still a poorly understood process in a quantitative manner, and

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![Fig. 6. Relationship between seismic anisotropy and fabric intensity. (A) Variation of fabric intensity \((J)\) index with \(V_p\) and \(V_s\) anisotropy for the experiments of Nicolas et al. (1973). (B) Variation of fabric intensity \((J)\) index with \(V_p\) and \(V_s\) anisotropy for 25 olivine aggregates. (C) Frequency diagram for fabric intensity \((J)\) index for 25 olivine aggregates. (D) Frequency diagram for maximum \(\delta V_s\) (%) for 25 olivine aggregates.](image-url)
as it takes place on grain boundaries it will undoubtedly be sensitive to presence of melt or chemically active fluids (e.g. H₂O and CO₂) which are likely to be present in the upper mantle but were absent in the experiments. Within the rather large uncertainty which constrains our current knowledge, it appears that the upper-mantle anisotropy is essentially dominated by olivine and has a limited strain memory which will be approximately 60% (or a shear strain of 1).

In this section we have shown that there is an approximately linear increase of seismic anisotropy with strain at low strains (< 60%). At high strains (or fabric intensities) the relationship is less clear, probably due to the onset of recrystallization. In experimentally deformed samples, recrystallization reduces the anisotropy. In natural samples the anisotropy saturates at high values of fabric intensity. In all cases the S-wave anisotropy appears less sensitive to strain (or fabric intensity) than P-wave anisotropy.

5. The seismic properties of the average of five mantle nodules

From the elastic constants we have determined from petrofabric analysis, we now seek to estimate the seismological properties of the average of these samples, and then make the assumption that these are representative of the subcontinental lithosphere. The average of five aggregates from 120 to 170 km depth is probably more representative of the mantle than individual specimens over the length scale sampled by SKS-waves. In doing so we have assumed that the lithosphere is homogeneous in terms of composition and deformation over at least a 50 km interval. We have calculated the average olivine pole figures of the five samples (Fig. 7(A)) which show that there is a high concentration of [100] close to the lineation (X) and [010] normal to the foliation (Z). The [001] axes are dispersed in the foliation plane (XY). The average sample has a

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**Fig. 7.** Average fabrics and anisotropies of the five kimberlite nodules. (A) Pole figures of olivine. Conventions as in Fig. 1. (B) Contoured P-wave distribution (km s⁻¹). (C) S-wave seismic anisotropy parameter \( \delta V_s \) (%). (D) Fastest S-wave polarization plane.
composition of 64.8% olivine, 31.2% enstatite and 4.0% garnet (Table 3). We are then interested in the properties of the average elastic stiffness tensor $\langle C_{ijkl} \rangle$. There are several properties we seek to investigate.

5.1. Effective elastic symmetry

We have investigated this by trying to fit the elastic stiffness coefficients $\langle C_{ijkl} \rangle$ to various symmetry systems. Various assumptions have been made in previous seismological studies about the symmetry elastic of the upper mantle; our data may place some constraints on the effective elastic symmetry. The three symmetry classes we have considered are isotropic, hexagonal and orthorhombic. Hexagonal is sometimes termed transverse isotropic in seismological studies. Following the scheme of Harder (1988), $\langle C_{ijkl} \rangle$ is first rotated into a standard coordinate frame, based on the eigenvectors of the tensor formed by the contraction $V_{ij} = C_{imnj}$. Next three inversions are performed for the three symmetry systems with two (isotropic), six (hexagonal) and nine (orthorhombic) free parameters to fit the 21 independent constants comprising the tensor. Misfit is measured by a normalized misfit parameter ($\epsilon$), which is corrected to take into account the variable number of degrees of freedom of the symmetry systems

$$\epsilon = \frac{\| \langle C_{ij} \rangle_{\text{symm}} - \langle C_{ij} \rangle_{\text{obs}} \|^2}{\| \langle C_{ij} \rangle_{\text{obs}} \|^2}$$

where the $\langle C_{ij} \rangle_{\text{symm}}$ tensor corresponds to the symmetry system and $\langle C_{ij} \rangle_{\text{obs}}$ is the average elastic tensor of the five kimberlite nodules. For isotropic symmetry, the quantity $(\epsilon_{\text{iso}})^{1/2}$ is a measure of the total anisotropy. We then define the ratio

$$S = \frac{(\epsilon_{\text{iso}} - \epsilon_{\text{hex}})}{(\epsilon_{\text{hex}} - \epsilon_{\text{orth}})}$$

as the symmetry index ($S$). It will be much greater than unity for hexagonal symmetry and much less than 1 for orthorhombic symmetry. From Table 3 we see the following results. The average tensor of the five kimberlite nodules has predominantly hexagonal symmetry, $S = 4.43$. For the olivine

<table>
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<th>TABLE 3</th>
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<td>Average seismic properties of five kimberlite nodules</td>
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<td>Depth range (km)</td>
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<td>$\delta V'_p$ (%)</td>
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<td>direction X</td>
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<td>Symmetry parameter, $S$</td>
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<td>olivine</td>
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component of these nodules, \( S = 2.7 \) is more orthorhombic than the hexagonal rock, and the orthopyroxene component has the most orthorhombic symmetry with \( S = 0.65 \). Nevertheless, the addition of the two orthorhombic-like components results in an aggregate symmetry which is more hexagonal in character. This is due to the two orthorhombic symmetry minerals both having a two-fold symmetry axis parallel to the foliation normal (Z-axis), [010] of olivine and [100] of orthopyroxene (see Figs. 1 and 2), and the other two-fold symmetry axes being non-parallel in the specimen foliation (XY plane). Although average elastic constants of the nodules are hexagonal (or transverse isotropic), the P-wave velocity distribution has orthorhombic symmetry (Fig. 7), the \( V_p \) delay time has roughly hexagonal symmetry and the \( V_{s1} \) polarization planes have monoclinic symmetry. Hexagonal symmetry in elastic properties results in a single axis of revolution, in this case nearly parallel to the Y-axis. It would appear that sampling only part of the elastic tensor, for example with \( V_p \), will give a false indication of the true elastic symmetry.

5.2. Magnitude of anisotropy

The magnitude of anisotropy in the upper mantle is especially important in the study of shear-wave splitting, as it trades off with path

Fig. 8. Delay time (\( \delta t \)) as a function anisotropic layer thickness. (A) ZY section. (B) XY section. Lines at layer thicknesses of 50–350 km at 50 km intervals.
length (Silver and Chan, 1991). The values for the average of five aggregates are shown in Table 3 and Fig. 7. The maximum value of $\delta V_s$ is 3.7% or 1.5 s delay time ($\delta t$) for a 200 km thick anisotropic layer. For the three principal structural directions ($X$, $Y$ and $Z$), the largest splitting (3.1%) occurs for the direction perpendicular to the lineation direction and within the foliation plane (Y-axis). There is roughly a factor of 2 difference between this direction and the other two ($X$ and $Z$). Thus, the maximum effect would be obtained if the deformation is such that the foliation plane is vertical and the lineation direction horizontal, and hence the vertical SKS propagation direction is parallel to $Y$. If, on the other hand, the foliation plane and the lineation direction were horizontal, then for vertical propagation, in a direction normal to the foliation plane ($Z$), one would then expect 1.6% or $\delta t = 0.6$ s for a 200 km layer.

A second important characteristic is that the faster S-wave is polarized in a plane containing the maximum concentration of olivine [100] axes and the lineation ($X$) (Figs. 7(A) and (D)). Hence, whatever the orientation of the propagation direction, the polarization of seismic waves should give information about the orientation of the lineation ($X$).

Typically, SKS delay times are about 1 s. If these samples represent the subcontinental mantle, then such a $\delta t$ requires an anisotropic lithosphere of 150 km (Fig. 8) with a Y-axis between vertical (inclination 90°) and 45° to the vertical. The P-wave properties of the average of five kimberlite nodules (Fig. 7(A)) is typical of many peridotites (e.g. Christensen, 1984; Nicolas and Christensen, 1987) with a maximum $V_p$ parallel to the lineation. The $V_p$ anisotropy is, however, low at 2.7%. The $P_n$ anisotropy ($\delta P_n$) is highest in the structural $XZ$ plane at 2.5% (Table 3), consistent with $P_n$ anisotropy beneath the continents of about 3% (e.g. Bamford, 1977; Vetter and Minster, 1981; Beghoul and Barazangi, 1990). Hence the same structural orientation ($Y$ vertical, $XZ$ horizontal) gives a maximum for SKS delay time and $P_n$ anisotropy.

If this is characteristic of the subcontinental lithosphere as a whole, then we can make the following statements. First, for those modes of deformation that place the foliation plane vertical and lineation horizontal (such as strike-slip motion) the delay time will be relatively high. For deformation in which the foliation plane is horizontal, such as asthenospheric flow, the effect will be roughly half as large.

5.3. Dependence of magnitude on composition

With the use of separate petrofabrics of olivine and orthopyroxene, it is possible to examine anisotropy as a function of composition. The measured fabrics have been used and the ratio of the olivine and orthopyroxene volume fractions varied to simulate compositional variation, assuming the fabric remains constant. In so doing we are assuming that there are no mechanical interactions between the olivine and orthopyroxene components during the fabric-forming process (i.e. plastic deformation, Wenk et al., 1991) and that fabric strength does not depend on volume fraction above the percolation threshold for olivine. We have varied the olivine component between 50 and 100%, whereas the kimberlite nodules we have studied contain between 56–80% olivine. One of the surprising features of the data is that the behaviour of P and S are very different (Fig. 9). P-velocity is a strong function of composition, with the maximum anisotropy going from 13.9% for 100% olivine to 6.0% for 50% olivine in LMA2.5 (Fig. 9). In contrast, the variation of maximum $\delta V_s$ is much smaller, from 7.1% to 9.5%. Thus, the intrinsic S-wave anisotropy is only a weak function of composition, although the ratio of $V_p/V_s$ anisotropy will be sensitive to composition. The physical reason for the relatively poor compositional sensitivity of $\delta V_s$ is due to the very similar orientations of the polarization planes produced by the petrofabrics of olivine and orthopyroxene resulting in constructive interference, and the low strength of the orthopyroxene petrofabrics. In a practical sense, then, the combination of SKS and $P_n$ observations could be used as a measure of olivine content. The $P_n$ anisotropy ($\delta P_n$) for each structural plane is given in Table 3. The effect of composition on the orientation of the maximum for shear-wave splitting can been seen in Fig. 9(C). For KBJ17 and
5 Kimberlite Nodules

![Graphs showing effects of composition on seismic anisotropy](image)

Fig. 9. Effect of composition on the seismic $V_p$ and $V_s$ anisotropy. (A) $V_p$ anisotropy as a function of olivine content. (B) $V_s$ anisotropy as a function of olivine content. (C) Maximum $\delta V_s$ orientation as a function of olivine content.

KBBF20 the increase in orthopyroxene tends to move the maximum $\delta V_s$ towards $Z$, for the other nodule fabrics there is almost no change.

These calculations show how the orthopyroxene affects the overall anisotropy of the aggregate. From Fig. 9 we can see that 30% orthopyroxene 70% olivine aggregate has an anisotropy which is lower by about 1% for $\delta V_s$ and 2% for $\delta V_p$ than the 100% olivine polycrystal. Data in Christensen and Lundquist (1982) show a similar reduction in the same compositional range for an ophiolite sample. We can correct the average anisotropy values of the 25 naturally deformed samples (Fig. 6) for 100% olivine $\delta V_s$ and $\delta V_p$ to a more typical mantle composition (e.g. 30% orthopyroxene) by using the reductions derived above from Fig. 9 of 2% for $\delta V_p$ and 1% for $\delta V_s$, giving 9–10% and 8%, respectively.

6. Discussion

The seismological properties of the mantle samples we have examined so far reveal some surprisingly simple systematics that, if confirmed by additional data, will greatly facilitate the interpretation of shear-wave splitting data.

6.1. Magnitude of $\delta V_s$

The most critical parameter for determining path length through an anisotropic layer is the assumed value of intrinsic shear-wave anisotropy $\delta V_s$. Based on existing analyses of ophiolites, this number has typically been assumed to be about 4% (Silver and Chan, 1991). However, application to the subcontinental mantle requires the assumption that deformation mechanisms in both regions are the same. The measurements of mantle nodule samples that directly sample the subcontinental mantle provide a much more direct measure of this parameter. We have shown that the magnitude of $\delta V_s$ appears to be less sensitive to olivine content than P-waves and thus would be little affected by chemical variations that might be found in the subcontinental mantle. The much stronger dependence of $P_n$ anisotropy ($\delta P_n$) on composition could explain the much smaller val-
ues of about 3% beneath continents (Bamford, 1977; Vetter and Minster, 1981; Beghoul and Barazangi, 1990), compared with 6% in the ocean basins due to increased olivine content just below the oceanic Moho, compared with the continental Moho. Here we have assumed that the orientation of the anisotropy has remained constant, at least in the vertical direction. Montagner and Tanimoto (1991) find that the anisotropy is constant in a vertical direction below the oceanic regions, but tends to rotate with depth below the continents. In such a situation the averaging of the anisotropy along the SKS ray path will result in a lower anisotropy below the continents relative to the oceans. Alternatively the higher $\delta P$, in the ocean basins may simply reflect a higher temperature regime with stronger deformation fabrics as suggested by the data from naturally deformed ophiolites in Fig. 6.

The simple dependence of $\delta V_s$ on orientation of the mantle samples suggests some intriguing features of anisotropy. Table 3 and Fig. 7 show that $\delta V_s$ can vary by a factor of 2, depending on orientation. For paths normal to the foliation plane ($Z$) or parallel to the lineation direction ($X$), $\delta V_s$ is 1.6–1.7% while for paths along the $Y$ direction $\delta V_s$ is 3.1%. If one assumes that the splitting delay times $\delta t$ are due to anisotropy in the lithosphere, then the path length has to be reasonably consistent with estimates of lithospheric thickness. Since $L = \delta t <V_s^> > / \delta V_s$, where $L$ is path length and $<V_s^>$ is isotropic shear velocity, a $\delta t$ of 1.7 s, the value obtained by Silver and Chan (1991) for the Red Lake, Ontario, Canadian Shield, with $\delta V_s = 1.6\%$ corresponds to a thickness of over 500 km, which is clearly inconsistent with other observations of lithospheric thickness. Using the higher value ($\delta V_s = 3.1\%$)

Fig. 10. Schematic summary of the interpretation of SKS-wave polarizations due to transcurrent movement in the subcontinental lithosphere. Inspired by a strike-parallel motion interpretation of the Appalachian mountain belt by Vauchez and Nicolas (1991). $V_s^>$ is the unpolarized S-wave in the isotropic mantle. $V_s^1$ and $V_s^2$ are the two orthogonally polarized S-waves which reach the continental station after traversing the anisotropic subcontinental lithosphere. The velocity of $V_s^1 > V_s^2$. $V_s^1$ is polarized parallel to the movement direction ($X$), which is represented by a transcurrent vertical ductile fault. The kimberlite nodule has its foliation plane ($XY$) vertical and parallel to fault plane. The lineation ($X$) is horizontal.
gives 250 km (see Fig. 8), which, while slightly higher than that quoted by Silver and Chan (1991) (~200 km), still appears to be consistent with the rather high velocities found to great depths beneath the Canadian Shield (Grand, 1987). Thus, we are required to use the higher value. This, in turn constrains the foliation plane to be vertical (Z horizontal), the lineation direction to be horizontal and the Y direction vertical, for a vertically propagating shear wave. We summarize the proposed interpretation in Fig. 10. Such a deformation would be most easily caused by transcurrent motion, rather than simple shortening as predicted by Vauchez and Nicolas (1991) for mountain belts.

In the above, we are assuming that the anisotropy (and presumably fabric strength) are on average constant throughout the anisotropic layer being used to define the seismic lithosphere. The frequency distributions of fabric intensity (J) and anisotropy of the samples studied to date (Fig. 6(C)) suggest that it may be possible to make such an assumption as the distribution is narrow. Also, we are assuming that all the anisotropy is in the lithosphere. Observations by Montagner and Tanimoto (1991) suggest that there is some anisotropy below shield areas down to 300 or 400 km. If we make a two-layer anisotropic model with the lithosphere as the upper layer and a lower layer with a weaker anisotropy, in the most likely case the anisotropies of the two layers will not be parallel, resulting in destructive interference. The averaging of the anisotropy along the SKS ray path will result in a lower anisotropy observed at the surface and hence our thickness estimates are a lower bound of the true value.

6.2. Relation between strain and fast polarization direction

For vertical propagation, and assuming that the conclusions of the previous sections are correct, any deformation for which \( X \) or \( Z \) is vertical will have half the delay time for an equivalent path length. The one type of deformation where such a configuration is expected is asthenospheric flow associated with absolute plate motion (APM), in which \( Z \) is vertical and the foliation (XY) horizontal. However, the style of deformation may in fact be quite different. Indications that this may be true are given by the 25 ophiolites, for which the average olivine fabrics are stronger \((J = 9.4, \text{whereas the average of the kimberlites } J = 8.3)\). Further the anisotropy of the ophiolites \((\delta V)\) is about 8%, twice the value of the kimberlite nodules (3.7%). It is not clear that such a distinction is meaningful as the lithosphere beneath the Canadian Shield has a seismic anisotropy \((\delta V) \approx 4%\) which we propose is linked to transcurrent movement in the subcontinental mantle over 250 km depth. At the present time, such lithosphere is considered to be relatively cold (with fast seismic velocities, e.g. Grand, 1987) and essentially mechanically elastic. The fossil mantle deformation that we are inferring is known to be parallel to Precambrian crustal deformation (Silver and Chan, 1991). To achieve such plastic deformations which produced the presently observed anisotropy, the mantle must have been significantly hotter than today. In such a situation we are perhaps looking at a fossil hot ductile asthenosphere which today is cold lithosphere. In this case we may wonder why the fabric strength of the kimberlite nodules is significantly lower than that of the ophiolite samples? Microstructural studies have all concluded that the granular nodules have undergone a long period of postdeformational annealing in the upper mantle (Boullier and Nicolas, 1975; Gueguen, 1977; Mercier, 1979; Drury and Van Roermund, 1989); it seems likely that the annealing process involving grain boundary migration is responsible for reducing the fabric strength. No experimental data under annealing conditions are available at present; however Nicolas et al. (1973) have shown that dynamic recrystallization under differential stress causes a reduction of fabric strength. Dynamic recrystallization involves some combination of rotational and grain boundary migrational mechanisms, hence the Nicolas et al. (1973) experiments should provide some indication of the fabric evolution in annealing with no stress. Note that the experiments of Nicolas et al. (1973) were at higher stresses that those usually accepted for the con-
vecting mantle. The role of processes such as grain boundary migration at low stresses may be more important with implications for the petrofabric evolution.

The $\delta V_s$ are observed to be high only for a very restricted orientation normal to the lineation within the foliation plane ($Y$-axis) (Figs. 7 and 8). Current measurements restrict the orientation of the $\delta V_s$ propagation direction for high $\delta t$ to within 45° of the $Y$-axis for a lithosphere thickness of 150 km ($\delta t = 1$ s) and considerably closer to the $Y$-axis for greater thicknesses or delay times. Further measurement may change this observation, but all the olivine and orthopyroxene fabrics of kimberlite nodules measured to date have essentially the same fabrics (e.g. Boullier and Nicolas, 1975; Mercier, 1985) and hence we do not expect this situation to change dramatically. However, we observe that olivine fabrics of plastically deformed samples from peridotite massifs tend to have girdles of (010) from $Y$ to $Z$, indicating (0kl) slip (e.g. Nicolas, 1989, p. 26). Kimberlite nodules have a strong concentration of (010) at $Z$ (Figs. 1 and 7(A)), indicating (010) slip; these differences may be the results of the physical conditions of deformation (temperature, pressure) or perhaps high-temperature postdeformational annealing of the kimberlite nodules over long time scales. The cause of the pronounced point maximum of (010) at $Z$ in the kimberlite nodules remains unexplained; however, due to this characteristic fabric the seismic properties have very simple systematics. A change from an (010) point maximum at $Z$ to a $YZ$ girdle will change the orientation of the maximum $\delta V_s$ (and $\delta t$) from $Y$ to somewhere in the $YZ$ plane, hence considerably changing the interpretation of SKS observations. Changes in composition for certain fabric types (e.g. KBBF20 and KBJ17, Fig. 9(C)) may also change the maximum $\delta V_s$ from $Y$ to closer to $Z$. At the present time we suggest that the nodules are representative of the subcontinental mantle of cratonic regions; further observations are needed to extend the sampling to other geodynamic environments.

7. Conclusions

From this initial study of five peridotite composition kimberlite nodules the average S-wave anisotropy is 3.7%. The polarization plane of the fastest S-wave is parallel to the lineation. The maximum S-wave delay time is parallel to the $Y$-axis, i.e. in the foliation ($XY$ plane) and normal to the lineation ($X$). Although the data set is small, all the petrofabrics are similar in these olivine-dominated (56–80%) rocks. In particular, the olivine [010]-axes form a point maximum parallel to $Z$ (normal to the foliation), rather than the typical girdle in the $YZ$ plane (normal to the lineation) found in ophiolites. The topological simplicity of the olivine petrofabrics is responsible for the single maximum in S-wave delay time parallel to $Y$. To obtain the observed SKS delay times of 1–2 s for reasonable anisotropic lithosphere thicknesses (150–250 km) the maximum S-wave delay time of the nodules ($Y$) has to be parallel to the vertical SKS ray path. Transcurrent flow in the mantle would produce a vertical foliation ($XY$ plane), a horizontal lineation ($X'$) and vertical $Y$-axis (Fig. 10). A transcurrent mantle flow with a strong mechanical coupling with the continental crust is consistent with the observation that the polarization plane of the fast SKS-wave is parallel to the trend of the crustal geology (Vinnik et al., 1984, 1989; Silver and Chan, 1988, 1991; Savage et al., 1990).

The fabric strength of peridotite ophiolites has been shown to be approximately twice as strong as this preliminary survey of kimberlite nodules. Further the topology of the ophiolite fabrics is more complex than that of the nodules. These differences will undoubtedly be reflected in differences in the seismic anisotropy of the asthenosphere and the subcontinental lithosphere, of which the ophiolites and kimberlite nodules are representative samples. In the present investigation we have not explored the possibility of a two-layer anisotropic lithosphere–asthenosphere model, although it remains a future objective.
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