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New advances in using seismic anisotropy, mineral physics and geodynamics to understand deformation in the lowermost mantle

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Abstract

The D'' region, which lies in the lowermost few hundred kilometres of the mantle, is a central cog in the Earth's heat engine, influencing convection in the underlying core and overlying mantle. In recent years dense seismic networks have revealed a wealth of information about the seismic properties of this region, which are distinct from those of the mantle above. Here we review observations of seismic anisotropy in this region. In the past it has been assumed that the region exhibits a simple form of transverse isotropy with a vertical symmetry axis (VTI anisotropy). We summarise new methodologies for characterising a more general style of anisotropy using observations from a range of azimuths. The observations can be then used to constrain the mineralogy of the region and its style of deformation by a lattice preferred orientation (LPO) of the constituent minerals. Of specific interest is the recent discovery of the stability of the post-perovskite phase in this region, which might explain many enigmatic properties of D'' . Mantle flow models based on density models derived from global tomographic seismic velocity models can be used to test plausible mineralogies, such as post-perovskite, and their deformation mechanisms. Here we show how linked predictions from mineral physics, geodynamical modelling and seismic observations can be used to better constrain the dynamics, mineralogy and physical properties of the lowermost mantle.

Keywords: D'' , lowermost mantle, mantle flow, anisotropy

1. Introduction

1.1. D'' and the lowermost mantle

The primary evidence for stratification of the Earth's interior comes from seismology. For nearly three quarters of a century seismologists have used changes in velocity gradients to map out the concentric shells that constitute the Earth's interior. Some changes are dramatic, like that seen at the core-mantle boundary (CMB), whilst others are more subtle,

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7 like that seen at the base of the lithosphere. Not long after Bullen's (1940) original classifi-
 8 cation of the lower mantle as the 'D' layer, it became apparent that the bottom few hundred
 9 kilometres of the mantle were seismically distinct from the bulk of the lower mantle. The
 10 lower mantle was split into D'—the top—and D''—the bottom (Bullen, 1949). Whilst much
 11 of the original nomenclature used to label the layers of the Earth has been abandoned, D''
 12 retains the name given to it over 60 years ago.

13 The D'' region encompasses a thermal boundary layer between the hot and vigorously
 14 convecting outer core and the colder, more slowly convecting mantle. It marks the terminus
 15 of downwelling mantle material and the place where upwelling plumes most probably origi-
 16 nate. It is often bounded by a seismic discontinuity that lies on average 250 km above the
 17 CMB (*e.g.*, Wysession et al., 1998), in many places contains ultra-low velocity zones at its
 18 base (*e.g.*, Garnero et al., 1998), and generally exhibits fine-scale structure revealed through
 19 scattered seismic energy (*e.g.*, Hedlin et al., 1997). The focus of this review is the obser-
 20 vation and interpretation of seismic anisotropy in this regio: in contrast to the overlying
 21 lower mantle, it exhibits significant seismic anisotropy (Meade et al., 1995; Montagner and
 22 Kennett, 1996; Panning and Romanowicz, 2006)

23 The implications of these observations are far reaching, as the CMB region plays a
 24 fundamental role in the dynamics of the mantle above and the core below. For example,
 25 core convection controls the generation of the Earth's magnetic field; mantle convection is
 26 the driving force behind plate tectonics. Making sense of the seismic observations requires
 27 a linked analysis of mineral physics, geodynamics and seismology. Here we present recent
 28 advances in each of these fields and show how they can be used to constrain the interpretation
 29 of measurements of seismic anisotropy.

30 1.2. Seismic anisotropy

31 Seismic anisotropy—the variation of seismic wave speed with direction—appears to be
 32 commonplace in the upper- and lowermost mantle (see *e.g.* Savage, 1999), and is probably
 33 present in the inner core (for a review, see Tromp, 2001). Anisotropy may be related to the
 34 inherent, wavelength-independent nature of the medium through which a wave travels, such
 35 as within the crystal structure of many minerals in the Earth; or it may be due to extrinsic,
 36 wavelength-dependent ordering of heterogeneous material, such as sedimentary layering in
 37 basins. In either case, the propagation of an elastic wave through the medium is described
 38 by the elasticity tensor.

The elasticity tensor c_{ijkl} gives the relationship between the applied stress σ_{ij} and the
 resulting strain ϵ_{kl} according to a linear relationship (Hooke's Law $\sigma_{ij} = c_{ijkl} \epsilon_{kl}$; for instance,
 see Nye, 1985 or Hudson, 1980a). The infinitesimal strain is

$$\epsilon_{kl} = \frac{1}{2} \left(\frac{\partial u_k}{\partial x_l} + \frac{\partial u_l}{\partial x_k} \right),$$

where u_n is displacement and x_n is the corresponding cartesian direction. The $3 \times 3 \times 3 \times 3$
 c_{ijkl} tensor can be reduced by symmetry ($\sigma_{ij} = \sigma_{ji}$) to a 6×6 matrix using the Voigt
 notation,

$$ij \rightarrow i, \quad kl \rightarrow j, \quad c_{ijkl} \rightarrow C_{ij},$$

2

$$C_{ij} = \begin{bmatrix} C_{11} & C_{12} & C_{13} & C_{14} & C_{15} & C_{16} \\ & C_{22} & C_{23} & C_{24} & C_{25} & C_{26} \\ & & C_{33} & C_{34} & C_{35} & C_{36} \\ & & & C_{44} & C_{45} & C_{46} \\ & & & & C_{55} & C_{56} \\ & & & & & C_{66} \end{bmatrix} .$$

39 The matrix is symmetrical, hence the lower elements are not shown, and there are 21 inde-
 40 pendent elastic constants which describe a minimally symmetrical, fully anisotropic system,
 41 an example of which would be a triclinic crystal. Increasing symmetry within a system
 42 reduces the number of independent elastic constants. For orthorhombic symmetries, there
 43 are nine; for hexagonal symmetry, there are five (C_{11} , C_{33} , C_{44} , C_{66} and C_{13}); for cubic there
 44 are three (C_{11} , C_{44} and C_{12}); and for isotropic media, there are only two (C_{11} and C_{44}). (For
 45 this special case, $C_{11} = C_{22} = C_{33}$, $C_{12} = C_{13} = C_{23}$, and $C_{44} = C_{55} = C_{66} = (C_{11} - C_{12})/2$.)
 46 A visual summary of the independent terms in the matrix C_{ij} for each crystal symmetry
 47 class can be found on p. 148 in Royer and Dieulesaint (2000).

48 Because the full tensor is so complicated, it is usual to make assumptions about the
 49 kind of symmetry present in the Earth; hexagonal symmetries are a good approximation
 50 where sedimentary layering or oriented cracks or inclusions are present. Where the layering
 51 is horizontal, the hexagonal symmetry can be described by a vertical axis of rotational
 52 symmetry; if it is inclined, then so is the symmetry axis (Figure 1). The plane normal to the
 53 symmetry axis is the plane of isotropy. When the plane of isotropy is horizontal (the axis of
 54 symmetry vertical), this is often referred to as vertical transverse isotropy (VTI), whereas a
 55 more general case where the plane inclined is termed tilted transverse isotropy (TTI).

In order to calculate the phase velocity along any particular direction given an elastic tensor, one solves the Christoffel equation,

$$\det|c_{ijkl} n_i n_j - \rho v_n^2 \delta_{il}| = 0 ,$$

56 where n_i is the unit normal to the plane wavefront, ρ is the density, v_n is the phase velocity
 57 along the plane wavefront normal, and δ is the Kronecker delta. The three eigenvalues of
 58 the solution correspond to the P and S wave velocities, V_P , V_{S1} and V_{S2} , along this direction
 59 (strictly, to the phase velocities of the quasi-compressional and -shear waves, which are not
 60 necessarily parallel and orthogonal respectively to n_i).

61 1.3. Shear wave splitting

62 Shear wave splitting occurs when a transverse wave travels through an anisotropic
 63 medium. Analogous to optic birefringence, this creates two orthogonally-polarised waves
 64 (the fast wave, S_1 and slow, S_2) (Figure 2). Depending on the distance travelled in the
 65 anisotropic medium, s , and the two velocities, V_{S1} and V_{S2} , the slow wave will be delayed
 66 by some time $\delta t = s \left(\frac{1}{V_{S2}} - \frac{1}{V_{S1}} \right)$. The measured polarisation of S_1 is termed the fast orien-
 67 tation, ϕ , and this is measured at the seismic station, hence ϕ is usually in the geographic
 68 frame and measured as an azimuth from north. The fast orientation in the ray frame, ϕ' , is
 69 measured relative to the intersection between the Earth radial plane (vertical) and the ray

70 normal plane, and therefore ϕ' is constant whilst the ray is not being actively split in an
 71 anisotropic region.

72 The strength of the S-wave anisotropy along a certain direction in the anisotropic medium
 73 is generally expressed as $\delta V_S = 2(V_{S1} - V_{S2})/(V_{S1} + V_{S2}) \approx (V_S \delta t)/s$. Hence in making
 74 measurements of splitting, normally one must assume a background ‘average’ V_S (from global
 75 1-D or tomographic models) and distance travelled in the anisotropic region, in order to
 76 calculate δV_S , with these uncertainties inherent. There is clearly a tradeoff between the path
 77 length in the anisotropic region and the strength of the anisotropy in that direction, hence
 78 in D'' —where the layer thickness determines the path length—our knowledge of δV_S in any
 79 particular direction is limited by the uncertainty in exactly where in the lowermost mantle
 80 the anisotropy lies.

81 The elasticity tensor can be visualised by examining V_P and V_S as a function of direction.
 82 We present the elastic behaviour of materials using upper hemisphere diagrams, explained
 83 in Figure 3. For all directions, we calculate the phase velocities as described above and show
 84 V_P and δV_S with colour. Additionally, the orientation of the fast shear wave, S_1 , is shown
 85 by black ticks. In these diagrams, we show the variation in elastic properties with respect
 86 to the three cartesian axes, 1, 2 and 3. Figure 3 shows the elastic constants for a set of
 87 mantle peridotites taken from Mainprice and Silver (1993). The 1–2 plane corresponds to
 88 the foliation in the sample, which probably results from a shear fabric. The 1-direction is
 89 aligned with the lineation, which probably shows the shear direction.

90 2. Measuring seismic anisotropy

91 The measurement of seismic anisotropy in the Earth has become routine for a limited
 92 number of techniques. In the deep mantle, work has mostly been directed towards observing
 93 the primary, unambiguous product of the presence of anisotropy: shear wave splitting in
 94 phases which traverse the D'' region. However new approaches are becoming available which
 95 can directly invert for anisotropic structure within the lowermost mantle using a broader
 96 range of data. Previous reviews of observations of D'' anisotropy are in Lay et al. (1998),
 97 Kendall (2000), Moore et al. (2004) and Wookey and Kendall (2007)

98 2.1. Correcting for the upper mantle

99 Measuring anisotropy in the deepest part of the mantle is not straightforward, as the
 100 upper mantle is known to be widely anisotropic itself (for a review, see Savage, 1999). The
 101 most common means of accounting for the effect of upper mantle anisotropy on D'' -traversing
 102 phases is to use a correction based on SKS splitting measurements. This phase traverses
 103 the outer core as a P wave and converts to a vertically polarised S wave (SV) at the CMB,
 104 hence is unsplit upon re-entering the lower mantle (Figure 4). Making the assumption of
 105 lower mantle isotropy, SKS should only split when encountering D'' and the upper mantle.

106 SKS studies are now numerous and successfully explain many features of upper mantle
 107 dynamics, on the basis that SKS’s path length in D'' is relatively small because the phase
 108 travels nearly vertically, and anisotropy in the lowermost mantle should not affect splitting
 109 in SKS much. Niu and Perez (2004) and Restivo and Helffrich (2006) compared SKS and

110 SKKS phases globally to investigate whether the lowermost mantle has an effect on such
111 phases. In some individual cases in regions of high shear velocity, such as beneath eastern
112 Canada, some discrepancy between SKS and SKKS was seen, which the authors attribute
113 to D'' anisotropy related to LPO of post-perovskite or some other non-VTI mechanism.
114 Overall, however, they found no significant departure from a mechanism in which SKS is
115 not split in D'' . This implies one of three things: anisotropy is not strong in D'' , which does
116 not appear to be the case from other measurements; anisotropy in D'' is not strong enough
117 to be noticeable for near-vertical rays like SKS-SKKS, which have a relatively short path
118 there; or the style of anisotropy (*e.g.*, VTI) means that radially polarised rays are not split,
119 as azimuthal anisotropy may cause splitting in SKS-SKKS phases (Hall et al., 2004). This
120 presents a puzzle for future studies of lowermost mantle anisotropy, as shall be explored.

121 If we continue with the assumption that SKS splitting reflects only upper mantle aniso-
122 tropy, then it can be used to remove the receiver-side splitting which occurs in a D'' -traversing
123 phase when reaching the seismometer. The ray paths in the upper mantle of S, ScS and Sdiff
124 are close to that of SKS for the distances discussed here, and their Fresnel zones at periods
125 of 10 s all overlap significantly down to ~ 300 km, so the effect of heterogeneity beneath
126 the receiver is addressed. This does not account for anisotropy beneath the earthquake,
127 however. One approach to address this is to use very deep-focus events (*e.g.*, >500 km),
128 which presumably do not experience much of the upper mantle anisotropic fabric as olivine
129 is only stable down to ~ 410 km. However, Wookey et al. (2005a), Rokosky et al. (2006)
130 and Wookey and Kendall (2008), for instance, show that there is observable splitting be-
131 neath even some deep events (<600 km), so this assumption may increase uncertainties in
132 observations of lowermost mantle splitting where no source-side corrections are made.

133 Further difficulties with SKS splitting-based corrections when examining lowermost mantle-
134 traversing phases are that in order to adequately correct for anisotropy beneath the receiver,
135 one must have a good knowledge of the type of anisotropy present there, as dipping or mul-
136 tiple layers of anisotropy will lead to observed splitting having a strong dependence of the
137 incoming polarisation of S-ScS-Sdiff. Choosing recording stations with many SKS measure-
138 ments from a wide range of backazimuths can help alleviate this. A 90° or 180° periodicity
139 in the splitting parameters ϕ and δt compared to the backazimuth betray the presence
140 of complex upper mantle anisotropy (Silver and Savage, 1994), which should be avoided.
141 Equally, stations which show little or no splitting across all backazimuths may be used with
142 no correction. For especially well studied regions, it may be possible to correct for even
143 complicated types of anisotropy (Wookey and Kendall, 2008), but the ability to uniquely
144 interpret such SKS splitting measurements is rare.

145 An additional factor to consider in using SKS measurements as an upper mantle correc-
146 tion is that S and SKS phases are of different slowness, so their incidence angles beneath
147 the receiver differ by up to $\sim 20^\circ$, depending on the epicentral distances being investigated.
148 In general, this will lead to a difference in the splitting accrued along the rays in the upper
149 mantle, hence an SKS-derived correction may not be appropriate. However, for an assumed
150 hexagonal anisotropy with a horizontal symmetry axis beneath the station, the difference
151 is small, and it appears in many studies the correction is adequate. Figure 5 shows the
152 receiver-side upper mantle splitting which occurs in SKS and S in a 250 km-thick anisotro-

153 pic layer. The elastic constants are of those shown in Figure 3 (Mainprice and Silver, 1993)
 154 with an imposed hexagonal symmetry. For SKS in the distance range $90^\circ \leq \Delta \leq 120^\circ$ (typ-
 155 ical for upper mantle SKS splitting studies), the range of incidence angles is small ($10\text{--}6^\circ$),
 156 and consequently there is almost no variation of splitting parameters with backazimuth. For
 157 S in the distance range $60^\circ \leq \Delta \leq 80^\circ$, incidence angles are $\sim 23\text{--}18^\circ$, and splitting in S
 158 shows some small variation with backazimuth. However, because the style of anisotropy is
 159 relatively simple, the difference in splitting parameters between S and SKS is very small—
 160 the fast orientations ϕ are indistinguishable, and the delay times are less than 0.3 s different,
 161 which is similar to the typical error in δt .

162 2.2. SH-SV traveltimes analysis

163 The most straightforward way to infer anisotropy in D'' is to compare the arrival times of
 164 the two components of a shear phase when polarised horizontally (SH) and vertically (SV)
 165 (or, respectively, the tangential and radial components), after correcting for upper mantle
 166 anisotropy. The phases studied are usually S, ScS and Sdiff, and the assumption is made
 167 that the wave travels approximately horizontally (CMB-parallel) when bottoming in D'' .
 168 Therefore, if SH arrives first, one can infer that along this azimuth the velocity is faster
 169 in the tangential direction than the radial ($V_{\text{SH}} > V_{\text{SV}}$). Figure 6 gives an example of this
 170 method.

171 In any study, constraining the source of the anisotropy to D'' is the main difficulty. There
 172 is good reason to suggest that the lower mantle above D'' is isotropic (*e.g.*, Meade et al.
 173 1995; Montagner and Kennett 1996; Panning and Romanowicz 2006), therefore taking pairs
 174 of phases—where one spends some time in D'' and the other avoids it—can be used to remove
 175 upper mantle effects. Figure 4 shows ray paths for the major phases used: S, ScS, and Sdiff.

176 Some of the earliest studies (*e.g.*, Lay and Young, 1991; Vinnik et al., 1995) inferred
 177 anisotropy by looking at the retardation (relative to SHdiff), amplitudes and phase shifts of
 178 SV waves diffracted along the CMB (SVdiff). However, anisotropy is not the only possible
 179 cause of these effects for waves diffracted past distances of $\Delta \gtrsim 95^\circ$, as shown by Maupin
 180 (1994) and Komatitsch et al. (2010). They model shear wave propagation in isotropic Earth
 181 models using the Langer approximation with perturbation theory, and spectral element
 182 method respectively, to show the early onset of SHdiff relative to SVdiff because of SV's
 183 coupling with the outer core, hence caution is needed in ascribing anisotropy to D'' on
 184 the basis of measurements of Sdiff at large distances: detailed full-waveform modelling and
 185 accurate isotropic Earth models are needed.

186 The majority of observations comparing SH and SV traveltimes show $V_{\text{SH}} > V_{\text{SV}}$, with
 187 $0.5\% \leq \delta V_{\text{S}} \leq 3\%$, particularly in higher-than-average V_{S} regions, such as beneath subduc-
 188 tion zones. Table 1 and Figure 7 summarise the observations for regional measurements
 189 of splitting in D'' . In general, however, it seems that around the Pacific rim, $V_{\text{SH}} > V_{\text{SV}}$.
 190 Beneath the central Pacific, the pattern is more variable: some studies find $V_{\text{SH}} > V_{\text{SV}}$, some
 191 $V_{\text{SH}} < V_{\text{SV}}$.

192 2.3. Global inversion for anisotropy

193 An extension of the above technique that can be made—in terms of searching for a VTI
 194 structure—is to produce a global inversion for a ratio of V_{SH} and V_{SV} ; usually the parameter
 195 $\xi = V_{SH}^2/V_{SV}^2$ is sought. Whilst global 1-D models of V_S such as PREM (Dziewonski and
 196 Anderson, 1981) sometimes include radial anisotropy in the upper mantle, at greater depths
 197 the inversions are generally isotropic. Montagner and Kennett (1996) used normal mode and
 198 body wave data to infer that $\xi > 1$ (*i.e.*, $V_{SH} > V_{SV}$) in D'' on a global scale. This matches
 199 the majority of local observations of SH-SV traveltimes. Recently, Panning and Romanowicz
 200 (2004, 2006) have inverted a global dataset of long-period three-component S waveforms to
 201 obtain a 3-D model of V_P , V_S , source parameters and ξ throughout the entire mantle. Any
 202 such study will be prone to difficulties in correcting for the strongly anisotropic crust and
 203 upper mantle, however, so great care is necessary to ensure that this does not contaminate
 204 the resulting model (Lekic et al., 2010). Equally, such models will necessarily suffer from
 205 sampling bias associated with the location of earthquakes and seismometers because of
 206 potentially limited azimuthal coverage of D'' . With observations along only one ray path, it
 207 is not possible to resolve whether VTI is a good approximation. However, the model agrees
 208 with regional observations, showing $V_{SH} > V_{SV}$ where V_S is higher than average, especially
 209 around the Pacific rim subduction zones. Where V_S is relatively low, such as beneath the
 210 central Pacific and beneath Africa, $V_{SV} > V_{SH}$. Similarly to the work of Montagner and
 211 Kennett (1996), it also predicts $\xi > 1$ for D'' on average (Figure 8). Kustowski et al. (2008)
 212 invert surface and body waves for 3-D anisotropic mantle velocities using similar data, but
 213 find strong tradeoffs in the lowermost mantle between V_S and ξ , and the anisotropic structure
 214 in D'' correlates poorly between the two models. It seems that at present there is still some
 215 room to improve on current global models.

216 2.4. Regional full-waveform inversion

217 An alternative to producing a global map of anisotropy is to conduct regional full-
 218 waveform inversion of seismic data from phases which traverse D'' . However, current studies
 219 are limited to assuming VTI in the lowermost mantle for computational and theoretical
 220 convenience. Using Tonga–USA raypaths, Kawai and Geller (2010) employ a full-waveform
 221 inversion for ξ beneath the central Pacific and find that $\xi < 1$ in D'' , though there is little
 222 sensitivity to structure below about 150 km above the CMB. This agrees with other studies
 223 along similar raypaths, with $\xi \approx 0.97$, which is at the lower end of the range of values found
 224 previously. Here, it was necessary to impose a discontinuity of arbitrary depth at the top of
 225 the model, and upper mantle anisotropy was not included, so this may have a large impact
 226 on the uncertainty.

227 2.5. Waveform analysis

228 Whilst relatively straightforward to implement, a weakness of any study which compares
 229 SH and SV waves is the assumption of VTI. Recently, efforts have been made to relax this
 230 constraint and infer more complex type of anisotropy.

231 An approach used by Garnero et al. (2004a) and Maupin et al. (2005) is regional forward
 232 waveform modelling of S–ScS waves beneath the Cocos plate and the Caribbean. They infer

233 small deviations of a TI symmetry of $\leq 20^\circ$ away from VTI as the raypaths move east to
 234 west across the region. Using an SH-SV traveltime approach, this would and does appear as
 235 $V_{SH} > V_{SV}$, though energy will appear on both radial and transverse components for both
 236 fast and slow arrivals.

237 2.6. Measurements of shear wave splitting

238 Another recent advance towards allowing more complex forms of anisotropy to be studied
 239 is to apply the measurement of both ϕ and δt by grid search over the splitting parameters
 240 (Fukao, 1984; Silver and Chan, 1991) to lower mantle-traversing shear phases (Figure 9).
 241 (This and other techniques such as the splitting intensity method (Chevrot, 2000; Vinnik
 242 et al., 1989) are summarised by Long (2009)). This allows one to determine a more general
 243 form of anisotropy, as the fast orientation is not limited to being either parallel or perpen-
 244 dicular to the CMB. In principle, with measurements along one azimuth, one can distinguish
 245 whether VTI is a possible mechanism for D'' anisotropy or not, two azimuths can define a
 246 TTI-type fabric, whilst three can define an orthorhombic symmetry of anisotropy.

247 One application of the measurement of shear wave splitting is to examine differential
 248 splitting between the S and ScS, usually investigated at epicentral distances $55^\circ < \Delta < 82^\circ$
 249 (with details of the method given by Wookey et al. (2005a)). Here, ScS samples D'' , S turns
 250 above it, and both phases share a very similar path in the upper mantle. Because the ScS
 251 phase is approximately horizontal for most of its travel in D'' at these distances, the ray
 252 frame fast orientation ϕ' (also ϕ^*) is used (Wookey et al., 2005a). This measures the angle
 253 away from the Earth radial direction (*i.e.*, vertical) when looking along the ray. Hence, for
 254 VTI with $V_{SH} > V_{SV}$, $\phi' = 90^\circ$. If $\phi' \neq 90^\circ$, then another mechanism such as TTI must be
 255 responsible.

256 Single-azimuth S–ScS studies beneath the northwest Pacific (Wookey et al., 2005a), Co-
 257 cos plate Rokosky et al. (2006) and southeast Asia (Thomas et al., 2007) have been con-
 258 ducted. Beneath the Cocos plate and southeast Asia, whilst there is some variability, in
 259 general fast directions do not depart much from being horizontal. Wookey et al. (2005a),
 260 however, found that the fast orientations dipped southeast towards the central Pacific by
 261 about 45° , which is a significant departure within the stated error of 7° . Assuming a TTI
 262 fabric, this actually provides a lower limit to the dip of the plane of isotropy, so clearly VTI
 263 in this region cannot explain the observations.

264 Recently, studies using two azimuths of S–ScS paths have been conducted. Beneath
 265 northern Siberia, Wookey and Kendall (2008) find that for waves travelling north from Hindu
 266 Kush events to stations in Canada, $\phi' = 89^\circ$ (the fast orientation is approximately horizontal
 267 in D''), whilst east-west paths from the Kuril arc to stations in Germany show $\phi' = 35^\circ$ (the
 268 fast direction dips 55° to the south). Beneath the Caribbean and North America, Nowacki
 269 et al. (2010) examine three regions with uncertainties of $\leq 10^\circ$ for all azimuths. For ray paths
 270 travelling north to stations in North America from events in South America, $\phi' \approx 90^\circ$, within
 271 error, which agrees with previous single-azimuth observations (Kendall and Nangini, 1996;
 272 Garnero and Lay, 2003; Garnero et al., 2004a). However, ray paths which cross these are not
 273 compatible with VTI: paths travelling northeast from the East Pacific Rise show $\phi' = -42^\circ$
 274 (dipping to the southeast), whilst those travelling northwest from the Mid-Atlantic Ridge

275 show $\phi' = 45^\circ$ (dipping south). A third region off the coast of northwest USA shows two
 276 paths with fast orientations $\geq 10^\circ$ different to horizontal.

277 In the cases outlined above, where $\phi' \approx 45^\circ$, the traditional SH-SV traveltime method
 278 would not observe any effects of anisotropy (Wookey and Kendall, 2007) (Figure 10). Equally,
 279 cases where $0^\circ < \phi' < 45^\circ$ cannot be distinguished from simple VTI where $V_{SH} > V_{SV}$. Hence
 280 the importance of not only resolving the fast orientation, but also incorporating a large range
 281 of azimuths, is hard to understate if we wish to make inferences about the nature and ori-
 282 gin of seismic anisotropy from analysis of shear waves. It seems that, in contrast to our
 283 previously simple idea of horizontal fast directions beneath subduction zones, and vertical
 284 ones beneath upwellings, the the picture is more complex. If VTI is not a good approxima-
 285 tion to the type of anisotropy in D'' , then multiple-azimuth studies must become the norm,
 286 otherwise we are at the mercy of the specific, single event-receiver geometry as to whether
 287 we can resolve the true effect of CMB dynamics. At the same time, however, the Earth
 288 does not give up its secrets easily, as the location of landmasses and large earthquakes poses
 289 limitations on which regions of the lowermost mantle we can probe at present.

290 Given that several studies have now implied that D'' does not everywhere show VTI-type
 291 behaviour, it is prudent to assess the discrepancy between this knowledge and the conclusions
 292 of Niu and Perez (2004) and Restivo and Helffrich (2006) (Section 2.1). Because azimuthal
 293 anisotropy appears to be present beneath at least Siberia, the Caribbean, western USA, the
 294 eastern and northwest Pacific and southern Africa, we should expect that studies comparing
 295 SKS and SKKS should exhibit differential splitting between the two phases which emerge
 296 from the outer core in these regions. In fact, as pointed out, Long (2009) and Wang and
 297 Wen (2007) do observe this in regional studies. In addition, Restivo and Helffrich (2006),
 298 for example, also show strong anomalous splitting between the two phases beneath western
 299 USA and the eastern Pacific, whilst southern Africa is poorly sampled because of event-
 300 receiver geometries. Furthermore, the Caribbean is not well covered: anomalous splitting in
 301 SKS-SKKS is evident there also, even if the global trend does not show significant departure
 302 from VTI for the whole dataset. Another factor is that because SKS and SKKS are polarised
 303 vertically upon exiting the outer core, they will not be split by TTI where the dip direction
 304 is closely parallel or anti-parallel to the wave propagation direction. Perhaps the largest
 305 difference is that even SKKS at $\Delta = 110^\circ$ spends around 350 km in a 250 km-thick D''
 306 with $\langle V_S \rangle = 7.3 \text{ km s}^{-1}$, whereas ScS at 70° has a path over 1000 km. It may therefore
 307 be not so surprising that SKS-SKKS differential splitting is hard to observe. However, the
 308 small number of cases where it is seen (5% of observations by Restivo and Helffrich (2006))
 309 requires a good explanation that is still lacking.

310 **3. Chemistry and mineralogy of the lower mantle**

311 The properties of the lowermost mantle are of course determined by the bulk compo-
 312 sition and which phases are stable at the pressures and temperatures there. In order to
 313 interpret seismic observations using geodynamic inferences, we must understand the single-
 314 and polycrystal behaviour of the solid phases present, and the possibility of the presence
 315 of melt. There are a number of steps which are necessary to use mineral physics data to

316 predict flow from anisotropy. Firstly, which phases are present must be established. Then,
 317 single-crystal elastic properties and deformation mechanisms must be evaluated. These can
 318 then be used to determine polycrystalline behaviour in deformation, which can allow an
 319 aggregate anisotropic fabric to be predicted on the basis of a given deformation history.
 320 Often it is hard to separate these in experiments, for instance, which involve many crystals,
 321 and authors attempt to find single-crystal properties from polycrystalline measurements.
 322 However successful modelling of texturing and hence anisotropy requires knowledge of all of
 323 these properties.

324 Lowermost mantle mineralogy can be investigated with mineral physics experiments
 325 at CMB pressures and temperatures using apparatuses such as the laser-heated diamond
 326 anvil cell (LHDAC), but there are of course limitations. An important source of error in
 327 experiments is the pressure scales used (the Au scale of Tsuchiya (2003), versus the MgO
 328 standard of Speziale et al. (2001), amongst others). This means the stated pressure, and
 329 hence depth, of the transition from pv to ppv in experiments can range by as much as
 330 ± 10 GPa (± 200 km in the lower mantle) depending on the scale, which is an ongoing
 331 problem (Hirose, 2007). Another significant source of error comes from the high thermal
 332 gradients created in the cell by focussed laser heating and diamond's excellent thermal
 333 conduction.

334 Numerical calculations of the properties of materials at high pressure and temperature
 335 are another important technique. As for physical experiments, however, uncertainties are
 336 present, due to the approximations necessary in performing the calculations. Density func-
 337 tional theory (DFT; Kohn and Sham, 1965) provides the basis for most of the studies
 338 we mention, which determines material properties by solving Schrödinger's wave equation.
 339 DFT gives an exact solution to the problem, but relies on an unknown term (the exchange-
 340 correlation energy). Different approximations to this term lead to different biases in the
 341 calculations. For a review, see Perdew and Ruzsinszky (2010).

342 *3.1. Composition and D'' mineralogy*

343 The Earth's mantle is generally believed to be pyrolitic in composition (Ringwood, 1962;
 344 McDonough and Sun, 1995). This chemistry determines which mineral phases are present
 345 under the conditions of the lowermost mantle, though some experimental evidence suggests
 346 that a representative pyrolitic material, the KLB-1 peridotite, may not alone be able to
 347 reproduce the seismically-observed density in the lower mantle (Ricolleau et al., 2009). In-
 348 put of other material such as mid-ocean ridge basalt (MORB) from subducting slabs must
 349 therefore play a role.

350 The phases present above D'' in a pyrolite composition are orthorhombic MgSiO_3 per-
 351 ovskite, with the likely incorporation of some Fe and Al (pv; Figure 11), cubic $(\text{Mg,Fe})\text{O}$
 352 (ferropericlase, fpc) and CaSiO_3 -perovskite (Ca-pv). Experiments suggest they are in the
 353 proportions 75, 20 and 5 % respectively (Kesson et al., 1998; Murakami et al., 2005) (Fig-
 354 ure 12). For MORB, which is much richer in Al and Si, experiments show a very different
 355 mineralogy (Hirose et al., 1999; Ono et al., 2001; Hirose et al., 2005), with about 40 % pv,
 356 no fpc and 20 % Ca-pv. Significant amounts of a Na- and Al-rich phase, and a silica phase
 357 (~ 20 % each) are present.

358 In 2004, several authors discovered another phase transition in MgSiO_3 to the orthorhombic
 359 CaIrO_3 structure at about 125 GPa (around 2700 km depth) and 2500 K (Murakami
 360 et al., 2004; Oganov and Ono, 2004). The post-perovskite phase (ppv) has a structure of
 361 layers of SiO_6 octahedra parallel to (010), intercalated with layers of Mg ions (Figure 11,
 362 right).

363 Recently, studies have been carried out on pyrolite and MORB samples up to CMB
 364 conditions. In pyrolite, Murakami et al. (2005) observe the pv–ppv transition at ~ 113 GPa
 365 (equivalent to ~ 2500 km) and 2500 K, where the phase assemblage is ppv (72%), fpc (21%)
 366 and tetragonal or cubic Ca-pv (7%). In MORB compositions, Ono and Oganov (2005)
 367 investigated pressures up to 143 GPa (Au standard) and temperatures of 3000 K. They
 368 observed ppv, Ca-pv, $\alpha\text{-PbO}_2$ -type (also called columbite) silica and a CaTi_2O_4 -type alumi-
 369 nous phase. Ohta et al. (2008) also investigated MORB samples with similar results, except
 370 they found a Ca-ferrite (CaFe_2O_4)-type aluminous phase at lowermost mantle conditions.
 371 They suggest a transition in silica from the CaCl_2 to $\alpha\text{-PbO}_2$ structure at around 115 GPa
 372 and 2000 K. Figure 12 summarises our current understanding of the phase proportions in
 373 the lower mantle.

374 Whilst we do not focus in this review on the gross variability of the phase assemblage
 375 at D'' conditions because of compositional changes other than pyrolite versus MORB, it
 376 is obviously important in the behaviour of the lowermost mantle, and there is increasing
 377 evidence that chemical heterogeneity must play a part in creating the seismic variability
 378 observed in D'' (e.g., Simmons et al., 2009).

379 3.1.1. Pv–ppv phase boundary

380 How much pv or ppv is present in the lowermost mantle is still unresolved. For pure
 381 MgSiO_3 , the phase boundary of course sharp and occurs at $\sim 110\text{--}120$ GPa, or 2400–2600 km,
 382 hence D'' would be mainly composed of ppv. However with realistic amounts of Fe and Al,
 383 the phase boundary will be spread out over a range of pressures. Whether the region of
 384 costability is extended upward in the Earth by the addition of Fe and Al, or downwards,
 385 depends on the partition coefficient of the element between the two phases. If Fe, for
 386 instance, partitions more favourably into pv, then it will be stabilised down into the ppv
 387 stability field, and costability of the two phases will occur to greater depths than for the
 388 pure Mg endmember. Partitioning into ppv would conversely increase the mixed phase
 389 region upwards into pv’s stability field. Thus this controls the amount of pv and ppv which
 390 are present in D'' . Additionally, Fe^{2+} and Fe^{3+} will behave differently, and how much iron
 391 is ferrous (Fe^{2+}) depends on the oxidation state of the lowermost mantle. It might also be
 392 that if another phase like fpc is present into which Fe (or Al) partitions preferentially over
 393 pv and ppv, then this will buffer the Fe content and decrease the width of the two-phase
 394 region.

395 Pv and ppv do include Fe and Al in their structure in a pyrolitic composition (Murakami
 396 et al., 2005), so the phase boundary between pv and ppv in various compositions is important.
 397 Whilst progress is being made, there has yet to emerge a consensus on the partitioning of
 398 Fe in particular between fpc and ppv, versus fpc and pv, hence there remains uncertainty
 399 in the pressure range across which pv and ppv are both stable. It seems that the partition

400 coefficient of Fe between pv and ppv, $K_{\text{Fe}}^{\text{pv/ppv}}$, is strongly dependent on Fe and Al content of
 401 the phases. Recent work at CMB conditions suggests $K_{\text{Fe}}^{\text{pv/ppv}} \approx 4$ (see Andrault et al., 2010,
 402 and their introduction for a recent concise review), and the phase boundary is predicted to
 403 be about 15 GPa or 300 km thick. Catalli et al. (2009) measure the transition width to be
 404 about 20 GPa (~ 400 km) in a synthesised sample of $(\text{Mg}_{0.9}\text{Fe}_{0.1})(\text{Al}_{0.1}\text{Si}_{0.9})\text{O}_3$, and less than
 405 that in a sample without Al ($(\text{Mg}_{0.91}\text{Fe}_{0.09})\text{SiO}_3$), though this of course does not include the
 406 buffering effects of any other phases which are present in the Earth. Both studies suggest
 407 costability begins at pressures equivalent to 400–600 km above the CMB.

408 Sinmyo et al.’s 2008 study highlights the uncertainties in the measurements of K_{D} , finding
 409 that the large temperature gradient in the sample may cause the variability between stud-
 410 ies. Further, uncertainties in the pressure scales mean it is hard to define at exactly what
 411 depth the beginning of the mixed-phase region starts. Notably, actual peridotite samples
 412 (Murakami et al., 2005) apparently contain ppv at D'' conditions.

413 An additional factor to consider is that the phase proportion curve may not be linear
 414 across the transition, so larger or smaller amounts of ppv may be present than expected for
 415 a given pressure. One attempt to quantify this (Hernlund, 2010) suggests ppv is likely to
 416 exist in significant proportions ($>50\%$ of the mantle) after just a few tens of kilometres of
 417 the transition.

418 Measurements of the Clapeyron slope of the pv–ppv show it likely lies in the range 7–14
 419 MPa K^{-1} (Oganov and Ono, 2004; Tsuchiya et al., 2004; Ono and Oganov, 2005; Hirose
 420 et al., 2006; Tateno et al., 2009). This positive value implies that colder areas of the low-
 421 ermost mantle will be enriched in ppv relative to hotter ones, and also offers the possibility
 422 that because of the steep geotherm near the CMB, so-called ‘double-crossings’ of the phase
 423 boundary might occur, leading to lenses of ppv-rich mantle bounded above and below by
 424 pv-rich areas (Hernlund et al., 2005; Wookey et al., 2005b). The effect this might have on the
 425 development of anisotropy from LPO of ppv is intriguing but poorly understood at present.

426 3.2. Single-crystal elasticity of D'' minerals

427 With knowledge of the approximate proportions of phases present in the lowermost man-
 428 tle, an understanding of the individual minerals’ properties and relative stabilities is neces-
 429 sary to make predictions about the behaviour of seismic waves passing through this region.
 430 Hence there has been much interest in using both experimental and theoretical methods to
 431 investigate these properties. Recent reviews of some of the work done on lowermost mantle
 432 phases—mainly pv, ppv and fpc—can be found in Hirose (2007), Shim (2008), Ohtani and
 433 Sakai (2008) and Trønnes (2010), amongst others. Here we discuss the most basic property
 434 of the phases in D'' for our purposes, their elasticity, which provides a first-order idea of
 435 their contribution to seismic anisotropy.

436 3.2.1. Perovskite

437 Magnesium silicate perovskite (with about 10% Fe and a few percent Al in the structure)
 438 is the most abundant mineral phase in the Earth, and is likely present in some portions of
 439 the bottom few hundred kilometres of the mantle. Because pv and ppv make up most of the
 440 lower mantle, they are the primary phases to affect seismic waves, and thus most important

441 to understand well. Although perfect perovskites are cubic, pv is orthorhombic due to the
 442 rotation of the SiO_6 octahedra (Figure 11, left).

443 Single-crystal elastic constants for pv at lowermost mantle conditions are shown in Figure
 444 13. Elastic constants for pv have been calculated by Oganov et al. (2001), Wentzcovitch et al.
 445 (2004), Wookey et al. (2005b) and Wentzcovitch et al. (2006) at CMB pressure, the latter
 446 two at high T . Figure 13 shows that there is some discrepancy between the calculations,
 447 which appears to be due to differences in the C_{12} , C_{22} and C_{33} terms. The maximum δV_S is
 448 between about 13–20 %, which is moderately but not very strongly anisotropic.

449 3.2.2. Post-perovskite

450 With the discovery of ppv (Iitaka et al., 2004; Murakami et al., 2004; Oganov and Ono,
 451 2004; Tsuchiya et al., 2004), there has been an understandable focus on its elasticity, phase
 452 stability, and so on, as explanations of lowermost mantle observations.

453 Intuitively, the orthorhombic ppv structure should be more seismically anisotropic than
 454 pv due to the layering of the SiO_6 octahedra, and this appears to be the case: the b-axis is
 455 more compressible than the a- and c-axes (Guignot et al., 2007; Mao et al., 2010). Elastic
 456 constants at D'' P and T have been calculated from experiments for ppv (Mao et al., 2010);
 457 *ab initio* calculations have recently been made by Wookey et al. (2005b), Stackhouse et al.
 458 (2005b) and Wentzcovitch et al. (2006).

459 Figure 14 shows the elastic anisotropy for ppv at high temperature, comparing the the-
 460 oretical calculations (MgSiO_3) at 4000 K to those of Mao et al. (2010) ($(\text{Mg}_{0.6}\text{Fe}_{0.4})\text{SiO}_3$) at
 461 2000 K. It is clear that there is some variation between the calculations. The experimentally-
 462 derived results show the largest δV_S , with $\delta V_S = 42\%$ along [010]. Otherwise, the pattern is
 463 quite similar between the studies of Stackhouse et al. (2005b) and Mao et al. (2010), despite
 464 the difference in Mg#. This agrees with the analysis of Wookey and Kendall (2007), who
 465 suggest from combining *ab initio* elastic constants for the MgSiO_3 , FeSiO_3 (Stackhouse et al.,
 466 2006) and AlSiO_3 (Stackhouse et al., 2005a) ppv endmembers in pyrolitic proportions that
 467 they do not differ significantly from those of pure Mg case. The general pattern of anisotropy
 468 differs slightly when considering the constants of Wentzcovitch et al. (2006), mainly due to
 469 differences in C_{11} , C_{33} and C_{13} ; the reason for this discrepancy is still unclear and hopefully
 470 future work will better constrain our knowledge of the single-crystal elasticity of ppv. It is
 471 notable that theoretical calculations with realistic amounts of Fe and Al in Mg-pv and -ppv
 472 are difficult because the number of atoms in the simulations becomes large, hence the effect
 473 of their incorporation is uncertain.

474 3.2.3. Ferropericlase

475 As the second most abundant mineral phase in the lowermost mantle, fpc is an important
 476 control on the behaviour of seismic waves in D'' . Assuming a pyrolitic mantle, an approxi-
 477 mate Mg# of 0.9 with Fe# = 0.1 is the likely composition. (Mg,Fe)O is stable throughout
 478 the lower mantle, though much recent interest has been shown in a possible change of its
 479 properties due to the change in the spin state in Fe which may occur at midmantle pressure
 480 and temperatures. We do not discuss in detail the spin transition in fpc further as it appears
 481 this occurs higher in the mantle than D'' (~ 2200 km; *e.g.*, Komabayashi et al., 2010); of

482 relevance is that Fe in fpc is likely in the low-spin state in the lowermost mantle. (For a
483 recent review of the spin transition in fpc, see Lin and Tsuchiya, 2008.)

484 Because fpc is cubic, the three constants required to describe the elastic behaviour of
485 the structure are C_{11} , C_{12} and C_{44} . Single-crystal elastic constants for fpc ($\text{Mg}_{0.9}\text{Fe}_{0.1}\text{O}$)
486 have recently been determined from experiment by Marquardt et al. (2009) up to 81 GPa
487 (~ 1900 km) at ambient temperatures. Karki et al. (1999) calculate the elastic constants
488 up to 150 GPa (greater than mantle depths) and 3000 K using *ab initio* methods for the
489 pure Mg endmember, whilst Koci et al. (2007) perform calculations at 0 K up to 150 GPa
490 for a range of Fe proportions up to 25 % ($(\text{Mg}_{0.75}\text{Fe}_{0.25})\text{O}$). Figure 15 shows a selection of
491 single-crystal elastic constants for MgO from theoretical calculations and ($\text{Mg}_{0.9}\text{Fe}_{0.1}\text{O}$)

492 It appears that the main effect of Fe in fpc is to decrease C_{11} and C_{44} , and increase
493 C_{12} (Figure 15; Koci et al., 2007), which in general will decrease the anisotropy of the
494 crystal (C_{12} becomes closer to $(C_{11} - 2C_{44})$, as for the isotropic case). Little work has been
495 conducted with Fe in the structure at high pressure, however, so these results are for high-
496 or intermediate-spin states of Fe, and it is not clear what effect low-spin Fe might have
497 on the anisotropy of fpc. As with pv and ppv, a large unknown at present is the partition
498 coefficient between these phases, hence our knowledge of the likely Fe content of any of them
499 at a particular pressure and temperature is limited.

500 3.2.4. Other phases

501 Whilst pv–ppv and fpc are the dominant phases in a pyrolitic composition at D'' condi-
502 tions, Ca-pv along with silica and aluminous phases are present in much larger proportions
503 in a MORB composition, hence knowledge of these phases is still important.

504 Ca-pv is predicted to undergo a transition from cubic to tetragonal due to rotation of
505 the SiO_6 octahedra at around 2000–2500 K at the CMB on the basis of *ab initio* molecular
506 dynamics (MD) simulations (Adams and Oganov, 2006; Stixrude et al., 2007), so potentially
507 in cold regions of the mantle this lower symmetry phase may exist. In contrast, Li et al.
508 (2006b) suggest—also from MD—that the tetragonal phase is stable throughout the lower
509 mantle. However, experiments at both pressures and temperatures of the lowermost mantle
510 have yet to be conducted, so the phase diagram of Ca-pv is uncertain. Li et al. (2006a),
511 Adams and Oganov (2006) and Stixrude et al. (2007) report elastic constants for Ca-pv at
512 CMB conditions. Cubic Ca-pv appears to be moderately anisotropic, showing maximum δV_S
513 of $\sim 20\%$, comparable to ppv and fpc, however the fact that it is a minor constituent of the
514 lowermost mantle means it is often neglected as a possible contributor to seismic anisotropy.

515 The silica phases most likely present in D'' are in the orthorhombic CaCl_2 or $\alpha\text{-PbO}_2$
516 (also called columbite) forms, with the transition occurring at about 110–120 GPa (2500–
517 2600 km). The implications for the presence of mainly the $\alpha\text{-PbO}_2$ -type in D'' are not clear,
518 as there are as yet no measurements of velocities or elastic constants for it at lowermost
519 mantle temperatures and pressures. Karki et al. (1997a) do report constants at high pressure
520 and 0 K from *ab initio* calculations (based on structure parameters reported in Karki et al.
521 (1997b)). At least at 0 K, the $\alpha\text{-PbO}_2$ -type silica shows a maximum δV_S of $\sim 15\%$, so
522 appears unlikely to be a major candidate anisotropic phase in D'', given its low abundance.
523 Future high- T work to elucidate the properties of free silica in the lowermost mantle will

524 have important repercussions for models where subducted MORB at the CMB plays a large
525 role in seismic anisotropy.

526 *3.3. Lattice preferred orientation and slip systems in D'' phases*

527 In order to generate anisotropy, individual anisotropic crystals must be aligned over
528 large lengthscales in a lattice- (or crystal-) preferred orientation (LPO, or CPO) (Figure
529 16A). Assuming that the phase undergoes deformation which is accommodated by slip on a
530 crystallographic plane (such as dislocation glide), the relative strengths of the slip systems
531 active in the crystal determine how the mineral aligns. Furthermore, how an aggregate of
532 individual crystals deforms depends on the phases present and their orientations.

533 At present, our understanding of slip systems and aggregate texture development for
534 mono- and polymineralic assemblages of phases at CMB conditions is poor, mainly because
535 it is currently impossible to recreate mantle temperatures, pressure (both very large) and
536 strain rates (very low) on large polycrystalline samples in the laboratory. However, various
537 experimental and theoretical methods have been used to examine the likely deformation
538 mechanisms.

539 There are two main approaches to evaluating the LPO caused by deformation in mantle
540 minerals. Firstly, one can investigate the phases at D'' conditions in the LHDAC, compress-
541 ing the sample by increasing the confining pressure during the course of the experiment,
542 leading to uniaxial deformation in the cell. Typically, radial X-ray diffraction data are taken
543 and the intensity of the individual diffraction lines is taken to correspond to the number
544 of crystals which are aligned in the orientation appropriate to cause the diffraction. The
545 ellipticity of the diffraction rings is a measure of the differential stress within the sample.
546 Thus a pole figure (orientation distribution function, ODF) can be calculated for the crystal-
547 lographic directions and a dominant slip system inferred. There are a number of limitations
548 to this technique, however—primarily, the sample size is very small (a few μm^3), hence the
549 amount of shortening is limited, and the sample is rarely actually at D'' temperatures when
550 observations are made: it is usually heated beforehand for some time, but is cooling when
551 lattice parameters are measured.

552 Alternatively, one can look at structural analogues of lowermost mantle phases which are
553 stable at conditions more easily achieved in the laboratory. Hence larger samples ($\sim 20 \text{ mm}^3$)
554 can be compressed, and the texture created examined directly. CaIrO_3 , MgGeO_3 and
555 MnGeO_3 have been used in this way, for instance, to investigate the slip system in ppv
556 as they share the same structure. So far, the Kawai and D-DIA (differential-DIA) appa-
557 ratures have been used to compress samples with a shear plane imposed at an angle to
558 the compression direction. (For a review of terminology and methods, see Durham et al.
559 (2002).) The sample is typically sheared to a shear strain of $\gamma \sim \mathcal{O}(1)$, and the sample
560 recovered and analysed with electron backscatter diffraction (EBSD) to determine the crys-
561 tallographic orientation of potentially thousands of crystals. An ODF can be calculated,
562 and slip systems inferred. Note that in such experiments, complex behaviour of polycrys-
563 talline material can be investigated, and several slip systems may operate. It is also notable
564 that the presence of other phases as compared to a single-phase assemblage can change the
565 deformation behaviour of an aggregate. This means that our long-term understanding of

566 how material deforms in D'' must rely on calculations or experiments on likely lowermost
567 mantle compositions.

568 Theoretical methods are also used to investigate deformation mechanisms, typically using
569 the generalised stacking fault (GSF) within a Peierls-Nabarro dislocation model. Often, *ab*
570 *initio* methods are used to find the GSF energy, feeding the Peierls-Nabarro model. Walker
571 et al. (2010) summarise the main techniques used. Others, such as Oganov et al. (2005),
572 use metadynamics to find new structures by perturbing the structure being studied, and
573 allowing it to relax to another, effectively pushing the structure over an energy barrier to a
574 new arrangement.

575 The purpose for this review of understanding single-crystal deformation mechanisms is
576 that we require such knowledge in order to infer deformation from measurements of seismic
577 anisotropy. With values for the relative strengths of slip systems, one can predict the
578 aggregate ODF and subsequent anisotropy of a polycrystalline assemblage. The predicted
579 slip systems may be used, for example, in a viscoplastic self-consistent model (Lebensohn
580 and Tomé, 1993; Wenk et al., 1991) and subjected to a known strain history, resulting in
581 predictions which can be compared to observations.

582 3.3.1. Perovskite

583 For pv, theoretical calculations have been combined with experiment to determine the
584 relative strengths of the dominant slip systems by Mainprice et al. (2008). Using a Peierls-
585 Nabarro dislocation model, they infer that the $[010](100)$ system is easiest at lowermost
586 mantle conditions. This agrees qualitatively with experiments performed at lower pressures
587 than present at the CMB (Cordier et al., 2004; Merkel et al., 2003), though high-temperature
588 studies are still awaited. Even with 100% alignment of the phase, the maximum δV_S is $\sim 2\%$,
589 which is significantly less than is the case for ppv or fpc. Hence it seems that, compared to
590 fpc and ppv, pv is a poor candidate phase to explain the near-ubiquitous observation of D''
591 anisotropy.

592 3.3.2. Post-perovskite

593 Table 2 summarises the experimental studies to date on slip systems in ppv and its
594 structural analogues. It is clear that little consensus exists regarding the dominant slip
595 system, with slip on (100) , (010) , (001) and $\{110\}$ all suggested by at least one study.
596 However, there is agreement for the slip system in CaIrO_3 . Recent DAC and large-volume
597 deformation experiments seem to confirm (010) as the likely slip plane for relatively large
598 strains, with perhaps $[100]$ the slip direction. Most studies also detect a different texturing
599 associated with the transformation from the pv to ppv structure—a so-called ‘transformation
600 texture’—consistent with slip on $\langle \bar{1}10 \rangle \{110\}$ (Walte et al., 2009; Okada et al., 2010; Hirose
601 et al., 2010). However, whether CaIrO_3 is a ‘good’ analogue for ppv—in the sense that it
602 deforms in the same way—is under debate (Walte et al., 2009; Hirose et al., 2010; Miyagi
603 et al., 2010; Mao et al., 2010; Okada et al., 2010). Hence whilst the advantages of using
604 relatively large, polycrystalline samples are obvious, care is needed in directly applying the
605 results of analogues to the case of the lowermost mantle.

606 Earliest theoretical work suggested on the basis of structural arguments that slip on (010)
 607 should be easiest, as this is the plane in which the SiO_6 octahedra lie, and indeed this agrees
 608 with experiments on CaIrO_3 . Carrez et al. (2007) suggest the system $[100](010)$ on the basis
 609 of Peierls-Nabarro modelling. Metsue et al. (2009) also find the same, though point out that
 610 despite the similarity between the predicted slip systems in ppv and CaIrO_3 , the starting
 611 single-crystal properties for the two phases are quite different, so drawing conclusions from
 612 such bases is difficult.

613 The observed ‘transformation texture’ of slip on $\{110\}$ (*e.g.*, Walte et al., 2009; Okada
 614 et al., 2010) adds complexity to our picture of the relation of deformation to anisotropy. If
 615 it is replicated in the pv–ppv transition, then it may be that descending mantle will acquire
 616 a certain texture for a time, which changes as strain increases. Hence future work to pin
 617 down whether such a process occurs in the Earth is important.

618 3.3.3. Ferropericlase

619 As the reader might have come to expect, great difficulties in experiments and theoretical
 620 calculations at extreme conditions mean there is disagreement between authors regarding
 621 the likely slip system in fpc. For NaCl-type cubic crystals, slip along $\langle 110 \rangle$ is expected to
 622 dominate, hence one might expect $\{110\}$ to be the likely slip planes for fpc (Karato, 1998).
 623 However, other slip planes may also be dominant, and high temperatures will affect the
 624 activation energies of the slip planes. *Ab initio* calculations for MgO and Peierls-Nabarro
 625 modelling (Carrez et al., 2009) suggests that the active slip system at low temperature
 626 is $\frac{1}{2}\langle 110 \rangle\{110\}$, though the $\frac{1}{2}\langle 110 \rangle\{100\}$ system becomes relatively easier with increasing
 627 pressure.

628 Experiments on the pure-Mg endmember at 47 GPa and ambient temperature by Merkel
 629 et al. (2002) in the LHDAC suggest slip on $\{110\}$. Contrasting results were found by Long
 630 et al. (2006), who used a large-volume press to deform a sample at 300 MPa and ~ 1400 K
 631 for a range of compositions ($0 \leq \text{Mg}\# \leq 1$). For pure MgO, $[001]$ tends to align with the
 632 shear direction, whilst $[110]$ aligns for FeO. Even for $\gamma \approx 4$, though, the development of
 633 LPO was fairly weak.

634 Yamazaki and Karato (2002) used compositions of $\text{Mg}\# = 0.25$ and 1.0 at $P = 300$ MPa,
 635 $T \approx 1000$ K with a very similar experimental setup to that of Long et al. (2006). They find
 636 slip on $\{100\}$ or $\{111\}$ is likely.

637 Whilst knowledge of individual slip systems is important, in the long term we require
 638 experiments and calculations on polycrystalline, multi-phase assemblages of the kind we
 639 expect to exist at D'' , as experience suggests monomineralic assemblages at vastly different
 640 conditions are not necessarily accurate proxies for the real thing. An improvement would be
 641 knowledge of the relative strengths of the several slip systems operating in the single crystal
 642 of any given phase. This would then allow one to calculate the development of texture under
 643 a known strain. An issue which seems very difficult to resolve experimentally is the vast
 644 difference in strain rates between studies and the Earth. It seems likely that strain rates in
 645 the deep mantle are $\dot{\epsilon} \approx \mathcal{O}(10^{-16})\text{--}\mathcal{O}(10^{-14}) \text{ s}^{-1}$, whilst at present we achieve $\dot{\epsilon} \gtrsim 10^{-4} \text{ s}^{-1}$,
 646 so whether we can ever recreate such strains is a hard question to answer positively.

647 **4. Shape-preferred orientation**

648 Thus far we have only considered the LPO of mineral phases as a potential cause of
649 lower mantle anisotropy. An entirely separate cause of anisotropy is the sub-wavelength
650 layering or ordering of material with contrasting elastic properties (Figure 16B and 16C).
651 The anisotropy may be due to the periodic layering of different materials or the preferred
652 alignment of inclusions like melt pockets.

653 If SPO is the cause of lowermost mantle anisotropy, it may still be a result of deformation
654 processes. To infer the link between deformation and observed anisotropy we must appeal
655 to effective medium theories that predict the anisotropy. A number of approaches exist,
656 but they can be divided into those that assume constant strain (*e.g.*, Hudson, 1980b) or
657 those that assume constant stress (*e.g.*, Tandon and Weng, 1984; Sayers, 1992). A further
658 complication involves the degree of interconnectivity between fluid inclusions, which leads
659 to frequency dependent anisotropy (for a review see Hall and Kendall, 2001). Assuming an
660 effective medium theory, an aggregate elastic tensor can be constructed and then used to
661 predict the seismic observables along a given ray path. Holtzman and Kendall (2010) de-
662 scribe such an approach for linking a number of anisotropy mechanisms to strain partitioning
663 at plate boundaries.

664 Spheroidal inclusions lead to a hexagonal symmetry or TTI (see examples in Figure
665 16B and 16C). A more complex orthorhombic medium results if the inclusions are scalene
666 ellipsoids (three axes of different lengths). However, on the basis of natural samples, which
667 tend to contain either elongate (prolate spheroidal) or flat (oblate spheroidal) inclusions, it
668 seems that in most settings one axis will be significantly different from the other two. An
669 example of each are L- and S-tectonites in subduction settings (Tikoff and Fossen, 1999).

670 With respect to the lower mantle, Kendall and Silver (1996; 1998), for example, model
671 the effects of spheroidal inclusions of contrasting velocity. They show that small volume-
672 fractions of oblate or disk-shaped inclusions of melt are highly efficient in generating seismic
673 anisotropy. In order for periodic layering or aligned inclusions to produce an effective an-
674 isotropy, and not simply heterogeneity, the wavelength of the layering must be less than
675 the dominant seismic wavelength. Indeed a way of discriminating between LPO and SPO
676 anisotropy may be through observations of frequency dependent effects. For example, small-
677 scale heterogeneity may scatter high-frequency seismic energy, but such a medium may be
678 effectively anisotropic to long wavelength energy (Rümpker et al., 1999).

679 Also compatible with observations might be the complementary presence of both SPO
680 and LPO. If, for instance, strain partitions into one weaker phase in a multi-phase mixture
681 (*e.g.*, a solid and liquid, or two solid phases with contrasting strengths; *e.g.*, Ammann et al.,
682 2010), then we might expect shear bands to form, as is frequently observed in surface geology.
683 If the bands are of the appropriate length scale, they might have an SPO contribution to
684 seismic anisotropy, whilst the highly deforming material in the bands—or even outside, for
685 the case of melt-rich bands—may still deform to produce LPO. Hence the division between
686 LPO and SPO is not necessarily clear whilst our knowledge of the lowermost mantle is at
687 this limited stage.

688 A major unknown in this sort of analysis is that the plausibility of melt in the lowermost

689 mantle is still speculative. Furthermore, much work is needed to better establish the material
690 properties of such melt, be they primordial in origin, the remains of subducted palaeo-oceanic
691 crust (basalt) or material derived from the outer core.

692 5. Geodynamics

693 While knowledge of the deformation mechanism of lowermost mantle materials is limited
694 (see section 3.3), one approach to assessing how likely they are to be realistic is to consider
695 the first-order flow expected just above the CMB. Topography on the CMB is limited to
696 a few kilometres at most (*e.g.*, Tanaka, 2010), and the outer core is liquid with a free-slip
697 surface above, so it seems highly likely that flow just above the CMB is mainly horizontal.
698 If we assume this, we might be able to mark as unlikely some of the proposed deformation
699 mechanisms for ppv, and then use the remainder to suggest slightly more nuanced flow
700 situations in D'' . We explore this further in section 6.

701 Global models of mantle flow have matured rapidly with increasing computer power and
702 new techniques over recent years, and inferring the first-order flow field at the CMB by
703 including geophysical observables such as recent plate motions and likely phase stabilities
704 and rheologies is now possible. Alongside this, models of mantle flow have developed which
705 are derived from seismic tomography, with the constraints of mineral physics, geoid and
706 plate motion data.

707 Where there is good evidence from seismic wave speed tomography (*e.g.*, Ritsema et al.,
708 1999; Montelli et al., 2004) of subducting slabs reaching the lowermost mantle, such as
709 the Farallon slab beneath North America, we can make slightly more detailed inferences
710 regarding the likely large-scale flow field. A simple approach used frequently (*e.g.*, Wookey
711 and Kendall, 2007; Yamazaki and Karato, 2007; Miyagi et al., 2010) is to assume horizontal
712 flow occurs at the CMB, and hence slip systems which produce fast orientations within the
713 slip plane are the likeliest to match the majority of observations which suggest $V_{SH} > V_{SV}$
714 in D'' . As section 2.6 shows, however, requiring horizontal fast directions in all directions
715 does not match with observations, so such assumptions must be revisited.

716 One constraint on the kind of deformation experienced in such a situation is to construct
717 models of mantle flow with an imposed subduction of a thermally negatively buoyant slab.
718 McNamara et al. (2003), for example, use a general 2D cylindrical model with diffusion
719 and dislocation creep to search the parameter space of variables such as slab thickness and
720 strength, and relative activation energies of the two creep regimes. They find that dislocation
721 creep dominates around the slab, and at the base of the mantle beneath the slab, whilst
722 the rest of the mantle is likely deforming in diffusion creep, hence not producing significant
723 LPO. They also claim that LPO in such a model requires $\gamma \gtrsim 4$ to develop. With this
724 method, where the whole Earth's mantle is modelled, but without imposing the constraints
725 of observed plate motions, the results can be qualitatively, and to some extent quantitatively
726 compared to deformation mechanisms in lowermost mantle mineral phases.

727 In order to construct models which are useful in understanding how the mantle flows in
728 D'' , a huge number of parameters are necessary, only some of which are known well. One-
729 dimensional radial viscosity profiles (*e.g.*, Mitrovica and Forte, 2004), for instance, place

730 a strong control on the depth and extent of subduction, which would then affect the flow
731 field above the CMB. Although these are constrained from present-day observables (mainly
732 isostatic glacial rebound of the surface for shallow depths, and mineral physics data much
733 deeper), obviously there is likely to be lateral variations in viscosity as well—such as that
734 introduced by a cold slab—which can only be modelled with accurate understanding of
735 the effect on viscosity of temperature, composition, mineralogy, and so forth. Other large
736 unknowns are the temperature at the CMB and the effect of composition and temperature
737 on the density of mantle phases.

738 In some studies (*e.g.*, Wenk et al., 2006; Merkel et al., 2006, 2007), workers take ‘gen-
739 eral’ models of flow of this kind and test for the type of anisotropy produced by a given
740 deformation mechanism when traced through the flow field. Assuming a certain flow field
741 as suggested by the convection model, they trace particles through the field and apply a
742 viscoplastic self-consistent (VPSC) model (*e.g.*, Lebensohn and Tomé, 1993, Wenk et al.,
743 1991) to calculate the texture developed for a polycrystalline aggregate using a set of slip
744 system activities relevant to the phases being tested. The resulting aggregate elastic tensor
745 is constructed from the single crystal constants and the orientation distribution function
746 (ODF) of the phases in the aggregate, and can then be compared with seismic observations
747 from similar settings—that is, beneath subducting slabs.

748 Another approach to modelling flow in the mantle is to seek a ‘true’ picture of what
749 is happening at present. Using seismic travel time picks, plate motion reconstructions
750 (Lithgow-Bertelloni and Richards, 1998), gravity measurements, dynamic topography and
751 other constraints, various authors (*e.g.*, Tackley, 2000; Trampert et al., 2004; Simmons et al.,
752 2009) have attempted to invert for the present-day or recent flow field in the mantle. Much
753 of this work depends on the particular relationship between seismic wave speed and density
754 in order to assess whether only thermal, or thermal and compositional effects are being seen
755 by the seismic velocities. With knowledge of the density anomalies which are thermal and
756 compositional (or mineralogical), one can produce a model of mantle flow. This seems a
757 promising approach to take, if we wish to assess whether we can use measurements of aniso-
758 tropy to determine flow in the mantle. For instance, if the flow is fairly constant over time
759 and shear strains are fairly large ($\gtrsim 1$, perhaps) then current mineral physics understanding
760 suggests we could observe LPO, providing the strain rate is high enough and dislocation
761 creep is occurring. If, on the other hand, strain rates predicted by such inversions are much
762 lower, then perhaps SPO is the likely mechanism.

763 A further step to take with such an approach is to directly incorporate experimentally or
764 theoretically derived slip system activities for a mono- or polymineraleic assemblage of grains
765 and perform VPSC calculations as above. The texture will be more complicated, and likely
766 weaker, but in theory more ‘realistic’. This does depend hugely on the flow model being
767 used, though tests on producing a synthetic seismic model from a global flow model by Bull
768 et al. (2010) suggest that the input and recovered strain fields are usually $<20^\circ$ apart. This
769 is encouraging from the perspective of hoping to be able to one day map deformation from
770 anisotropy, but adequate seismic coverage will long be a problem, as discussed in section
771 6.1.

772 6. Linking observations to physical processes

773 If the measurement of seismic anisotropy is to be useful in studying the dynamics of the
 774 lowermost mantle, then we need a close understanding of the rheology of mantle materials
 775 at CMB conditions. Section 3 discussed that we are still some way from fully understanding
 776 how to ‘measure’ dynamics in D'' using seismic anisotropy, but we are now at the stage
 777 where our inferences are informed by a great deal of work on the properties of lowermost
 778 mantle minerals. In the first instance, seismic anisotropy can be used to evaluate a number
 779 of different mechanisms which might cause it.

780 6.1. Inferring SPO and TTI

781 A simple mechanism to produce lower mantle anisotropy which cannot at present be
 782 ruled out is SPO. This has been the preferred interpretation in a number of studies (*e.g.*,
 783 Kendall and Silver, 1998; Lay et al., 1998; Karato, 1998), which model the expected bulk
 784 anisotropy for isotropic inclusions of material with a contrasting V_S in an isotropic medium.
 785 Kendall and Silver (1998), for instance, use the effective medium theory of Tandon and
 786 Weng (1984) to predict the shear wave splitting caused by horizontal rays travelling through
 787 a medium with oriented spheroidal inclusions. Whilst high-velocity inclusions are unlikely
 788 to be a mechanism which can match the observations (as the inclusions would need to have
 789 $V_{S_{inc}} \gtrsim 13 \text{ km s}^{-1}$), melt-filled inclusions ($V_{S_{inc}} = 0$) can produce $\delta V_S = 2\%$ with a melt
 790 fraction of just 0.01% for oblate spheroidal inclusions. Moore et al. (2004) show a D''
 791 with horizontal sub-wavelength layering of heterogeneous material can produce synthetics
 792 compatible with observations in certain regions. Both studies suggest that SPO—especially
 793 of melt—is an efficient way of producing anisotropy without much reducing the bulk average
 794 V_S (Kendall and Silver, 1996).

795 If we assume that SPO is the cause for an observed anisotropy, then this usually implies
 796 that the style of anisotropy is TTI (see section 4). Because of the high symmetry of TTI,
 797 two near-perpendicular azimuths of shear waves are sufficient to characterise the orientation
 798 of the symmetry axis (or plane of isotropy), as five independent elastic constants describe
 799 such a system and the local $\langle V_S \rangle$ can be assumed.

800 One simplistic way to infer the orientation of the TTI fabric is to assume a case where
 801 Thomsen’s (1986) parameters $\delta \approx \epsilon$, hence the fast orientation of a wave split by such a
 802 medium is always in the plane of isotropy for waves not perpendicular to the plane. Therefore
 803 a simple geometrical calculation to find the common plane of the fast orientations in the ray
 804 frame ϕ' can be used. Nowacki et al. (2010) use this to calculate the TTI planes of isotropy
 805 beneath the Caribbean and western USA (Figure 17). Figure 18 illustrates the nominally
 806 simple geometry for region ‘E’ in this study.

807 An alternative method used by Wookey and Kendall (2008) to estimate the orientation of
 808 the TTI plane of isotropy for two orthogonal ray paths beneath Siberia can be summarised
 809 as: (1) take a set of elastic constants C_{ij} for a TI system, with vertical V_S and V_P defined
 810 by a global 1-D velocity model (Kennett et al., 1995); (2) rotate these constants about
 811 all three cartesian axes and compute δV_S (and hence δt) and ϕ' at each point; (3) output
 812 the orientations which produce $(\phi', \delta t)$ which are compatible with the observations. This

813 inversion has the advantage that it can be simply extended for any set of elastic constants,
 814 and lies between analytic solutions from shear wave splitting measurements and inversions
 815 for the full elastic tensor, which would likely be poorly constrained.

816 6.2. Implications of SPO and TTI

817 If our assumption that the lowermost mantle shows a variable TTI type of anisotropy is
 818 correct—and it is worth noting that no studies as yet are incompatible with this symmetry—
 819 then what does this imply for the dynamics within and above D''? As discussed in the
 820 previous section, various authors have shown that SPO of melt pockets (or other low V_S
 821 inclusions) at the CMB could cause this, and this then begs the question as to where these
 822 melts come from. A possibility mooted by Knittle and Jeanloz (1987) was that reaction
 823 between core and mantle materials would lead to inclusions of Fe-rich products (*e.g.*, FeO,
 824 FeSi) in D'' (Kendall and Silver, 1998). However, the bulk reduction in V_{SH} from this does not
 825 match observations, hence is an unlikely scenario. As mentioned in section 4, Stixrude et al.
 826 (2009), for example, suggest that silicate melts might be present in the lowermost mantle
 827 at temperatures as low as 4000 K. Just 0.01% melt could be compatible with observations
 828 given the bulk sound velocity is predicted to be around 10.9 km s^{-1} .

829 If such models are accurate, then we require knowledge of how the inclusions—partially
 830 or wholly molten, or simply of contrasting velocity—align in response to flow, to make
 831 geodynamical inferences. To first order, weaker inclusions in a stronger matrix align parallel
 832 to the strain ellipse's long axis (*i.e.*, the shear plane) when the strain is high ($\gamma > 1$). Hence
 833 for the cases where we have two azimuths (in the Caribbean and Siberia), we would predict
 834 flow dipping between $26\text{--}55^\circ$ roughly to the south in D''. These steep angles seem somewhat
 835 unlikely for high strains, given that flow right at the CMB must be horizontal, but cannot
 836 necessarily be precluded.

837 Contrary to this first-order approximation, weak inclusions apparently rotate when sheared
 838 so that they are no longer parallel to the finite strain ellipse, as noted by Karato (1998).
 839 Numerous experiments—chiefly on olivine-MORB samples—indicate that shear bands of
 840 melt align antithetic to the shear plane at an angle of $\sim 20\text{--}40^\circ$ (Kohlstedt and Zimmerman,
 841 1996; Holtzman et al., 2003a,b). Taking the example of the regions studied by Wookey and
 842 Kendall (2008) and Nowacki et al. (2010), this melt orientation predicts horizontal shear to
 843 the north or northwest in western USA, and gently dipping flow to the south elsewhere in the
 844 Caribbean and Siberia. Figure 19 shows this situation with the shear wave anisotropy pre-
 845 dicted by sensible lowermost mantle parameters, where melt inclusions dip 25° southward,
 846 but due to northward flow. In the Caribbean, geodynamical calculations of the flow beneath
 847 subducting slabs would generally agree rather with east–west flow for a north–south-striking
 848 plate (McNamara et al., 2003), but at least this model seems physically possible.

849 The known mineral phases present at the CMB do not show hexagonal symmetry, how-
 850 ever an alternative explanation for TTI would be the alignment of one crystallographic axis
 851 of some anisotropic mineral phase, with the other axes random. As an artificial example,
 852 Figure 20 shows the case where an aggregate of ppv shows alignment of c-axes, but the a-
 853 and b-axes are otherwise randomly oriented. This might correspond to slip on the (001)

854 plane along both the [100] and [010] directions. This leads to TI with the symmetry axis
 855 parallel to the c-axis, where the fast shear wave is within the TI plane.

856 *6.3. Inferring orthorhombic and higher symmetries*

857 Whilst at present TTI cannot be ruled out as causative of the observed seismic aniso-
 858 tropy in D'' , a more general orthorhombic symmetry—such as that caused by alignment of
 859 orthorhombic crystals—is a more likely mechanism. Equally, cubic and lower symmetries can
 860 also produce the observed patterns of anisotropy. However, it is unlikely that distinguishing
 861 such a highly symmetric type of anisotropy will be possible with the current earthquake
 862 and seismometer geometries for some time, so assuming that orthorhombic anisotropy is the
 863 lowest symmetry likely to exist is, for now, a necessary step.

864 So far, no studies have been able to uniquely infer the orientation of an orthorhombic
 865 symmetry, because only measurements of D'' anisotropy along two directions have been
 866 made. However, Wookey and Kendall (2008) and Nowacki et al. (2010) use two azimuths
 867 and the technique described in Section 6.1 to test the orientations of different candidate
 868 orthorhombic systems beneath the Caribbean and Siberia. In the case of using two azimuths
 869 of measurements, one normally finds that two sets of planes are compatible. Figure 21 shows
 870 an example of fitting possible orientations of different (orthorhombic) elastic constants to
 871 measurements made beneath the three regions of Nowacki et al. (2010). They use a set
 872 of constants obtained by Yamazaki et al. (2006), who deform CaIrO_3 (same structure as
 873 MgSiO_3 -post-perovskite), and find that the [100](010) slip system is dominant. The elastic
 874 constants are referenced to the shear plane and slip direction imposed upon the deformation,
 875 so we can directly infer in which direction a material which behaves in this way is being
 876 sheared.

877 *6.4. Inferring deformation in D''*

878 We measure D'' anisotropy in the hope that it can provide information about the manner
 879 in which it is deforming, and hence how the mantle moves at depths. In order to estimate flow
 880 or strain from anisotropy, we must integrate our understanding of the cause of anisotropy, the
 881 orientation of the assumed anisotropy type, our knowledge of the rheology of the medium,
 882 and the response of the shear direction to the potentially changing flow field. Figure 22
 883 illustrates the many steps involved in getting from observations to predictions of deformation,
 884 and the many assumptions which are made along the way.

885 At present, the response of D'' materials to deformation is not well known, hence early
 886 attempts at inferring flow from measurements of seismic anisotropy were necessarily general.
 887 Beneath the circum-Pacific subduction zones where flow is assumed to be horizontal at the
 888 CMB, the global ξ models of Panning and Romanowicz (2004, 2006) show $V_{\text{SH}} > V_{\text{SV}}$, and
 889 thus it has been interpreted that likely mechanisms in response to shear in D'' mineral should
 890 produce fast orientations parallel to the shear plane. This then may lead to the inference
 891 that beneath the central Pacific, the change of $\xi > 1$ to $\xi < 1$ corresponds to vertical flow
 892 (*e.g.*, Kawai and Geller, 2010) or some sort of shearing in different horizontal directions (*e.g.*,
 893 Pulliam and Sen, 1998). Clearly, whilst there is short scale variability in the signal anyway,

894 determining the first-order flow field from an educated guess is an understandable first step
895 which we should attempt to improve upon.

896 In fact, this point highlights one of the current shortcomings in our addressing of the
897 problem of using seismic anisotropy to map deformation. At present, we are limited to
898 using ‘best guess’ estimates of the flow field in certain areas at the CMB (specifically, where
899 the ancient Farallon slab is presumed to be sinking to the CMB beneath North and Central
900 America, and to some extent other circum-Pacific subduction zones) to argue for and against
901 different mechanisms for producing seismic anisotropy. For instance, Yamazaki and Karato
902 (2007) prefer an explanation for D'' anisotropy of the LPO of a mixture of $(\text{Mg,Fe})\text{O}$ and
903 MgSiO_3 -post-perovskite because horizontal shear would give a horizontally-polarised fast
904 shear wave for this case, which is the sort of deformation postulated beneath deep slabs.
905 They then argue that SPO of melt inclusions oriented vertically is the likeliest case for the
906 central Pacific, because flow there is probably vertical and in higher-temperature material.
907 If the CMB is considered an impenetrable free slip surface, then why should flow not also be
908 mainly vertical in the very lowermost mantle beneath a downwelling as well as an upwelling?
909 Whilst these first-order explanations are sensible, they are only an initial idea about flow,
910 hence using this to constrain LPO and infer the presence of melt makes a large stride in
911 assumptions which we must eventually address with direct observations of lowermost mantle
912 rheology.

913 Nonetheless, many authors have inferred different flow regimes at the CMB based on
914 seismic anisotropy. Early work (*e.g.*, Vinnik et al. 1995; Lay and Young 1991; Ritsema et al.
915 1998) attributed anisotropy to stratification or LPO on the basis of the expected flow field
916 near the CMB. Later, Kendall and Silver (1996), for instance, identify slab material which is
917 laid down in piles parallel to the CMB as a cause of SPO. Recently, dual-azimuth splitting
918 measurements were used in combination with global V_S tomography to infer that north-
919 south flow beneath Siberia is the likely cause of anisotropy due to LPO of ppv (Wookey and
920 Kendall, 2008). Similarly, Nowacki et al. (2010) infer that an LPO of ppv whereby the (001)
921 planes align parallel to shear is most likely beneath the Farallon slab because of first-order
922 flow arguments, and then extend the argument to suggest that shear planes dip towards the
923 downwelling centre, analogous to the situation in mid-ocean spreading centres (Blackman
924 et al., 1996), and supported by general-case geodynamic calculations (McNamara et al.,
925 2002)

926 Future advances in incorporating all our current understanding of the behaviour of the
927 constituents of the lowermost mantle into linking observations and dynamics will become in-
928 crementally better. These early attempts at measuring the flow of the deepest mantle should
929 be surpassed as we use new information which becomes available from increasingly advanced
930 experimental and numerical techniques for studying seismic anisotropy, flow, geodynamics
931 and mineral physics.

932 7. Conclusions and future directions

933 In this review, we have presented the current state of studies which aim to use seismic
934 anisotropy to discover the flow in the deepest mantle, and the many other fields which feed

935 into this. It seems that we are moving from an early phase of D'' study into a more mature
 936 field, where the number of observations is now becoming limited by the location of seismic
 937 stations. As we look to the future, projects to increase global coverage of seismometers will
 938 benefit all studies of the Earth's interior, but especially that of the lowermost mantle. With
 939 this increased coverage, the prospect of using more advanced techniques to take advantage is
 940 an exciting one which may yet yield even harder questions that we currently try to answer.

941 One such technique that must be further explored with new datasets is the full inver-
 942 sion for the elastic tensor using the full seismic waveform. Recent advances towards this
 943 necessarily assume a simple anisotropy, but this can be relaxed as data coverage improves.
 944 However, as for global inversions for simple anisotropy, upper mantle and crustal corrections
 945 will be a problem. At the same time, existing global datasets—as used for global tomog-
 946 raphy, for example—might be exploited to move from regional shear wave splitting studies
 947 to global ones. This will require either a new, robust way of analysing shear wave splitting,
 948 which is still the most unequivocal of observations of anisotropy, or the further automation
 949 and quality control of standard techniques. Shear wave splitting ‘tomography’ is another
 950 technique which will likely prove important in the future.

951 Whilst seismological observations will be our primary test of models of D'' flow and
 952 anisotropy for some time, advances must be made in mineral physics and geodynamics if we
 953 are to improve. Studies of deformation in likely lowermost mantle mineral assemblages will
 954 hopefully go some way in the future to reducing the ambiguity regarding how to translate
 955 anisotropy to flow, and global mantle flow models may be able to become predictors of
 956 anisotropy with such knowledge.

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1335 **Figure and table captions**

Figure 1: Transverse isotropy, or hexagonal symmetry, and wave propagation through such a medium. On the left, the rotational axis of symmetry is vertical, leading to vertical transverse isotropy (VTI). On the right, the axis is tilted away from the vertical, leading to tilted transverse isotropy (TTI), or simply a general case of transverse isotropy (TI). Waves within the plane of isotropy are split into orthogonal fast (blue) and slow (red) waves. The dip θ and azimuth a (the dip direction) of the plane of isotropy define the TTI orientation.

Figure 2: Shear wave splitting in an anisotropic medium. The unsplit incoming shear wave encounters the anisotropic medium, and is split into two orthogonal waves, fast (S_1 , blue) and slow (S_2 , red). The delay between the two is measured as δt , and the fast orientation in the ray frame (measured relative to the vertical) is ϕ' .

Figure 3: Representation of elasticity tensor by the variation of V_P and V_S with direction. The leftmost diagram explains the wave anisotropy plots on the right. The tensor in the three cartesian directions 1, 2 and 3 is represented by an upper hemisphere projection of the variation of wave speed with direction. The top of the projection is the 1-direction, left the 2-direction, and out of the page the 3-direction. At each point (each inclination from the 3-axis, i , and azimuth clockwise away from 1 in the 1–2 plane, a), V_P (km s^{-1}) and δV_S (%) are shown by colour according to the scale at the bottom. On the δV_S plot, the orientation of the fast shear wave as projected onto the upper hemisphere is shown by the black ticks. Shown are the average C_{ij} for a selection of five kimberlites from Mainprice and Silver (1993), where the X-, Y- and Z-directions are oriented to the 1-, 2- and 3-directions respectively.

Figure 4: Raypaths of some of the body wave phases used to study D'' anisotropy.

Figure 5: Shear wave splitting parameters of SKS and S phases from upper mantle anisotropy. The two phases have slightly different slownesses, corresponding to a different incidence angle beneath the station. The upper hemisphere phase velocity plots, left, show the case of TI with a symmetry axis parallel to 1 (representing north). The 2-axis points west and 3 is up (out of the page). The elastic constants are those of Mainprice and Silver (1993) as shown in Figure 3, but with an imposed hexagonal symmetry. The circles at the centre of the δV_S plot show the range of incidence angles of SKS (red, innermost), S (blue, outermost) and ScS (black) phases at distances described in the text. The splitting parameters corresponding to these distances and backazimuths and a 250 km-thick layer are shown on the right for SKS (red) and S (blue). There is almost no variation in SKS, and for ϕ the two phases experience indistinguishable splitting. For δt , the largest difference is about 0.3s, and within typical errors the two phases would exhibit the same splitting parameters. The parameters for ScS lie between the two other phases.

Figure 6: SH-SV traveltime analysis, Figure 5 from Garnero and Lay (1997). The authors examine shear waves travelling along the CMB beneath Alaska from two events in 1970 and 1973, at distances $90.0^\circ \leq \Delta \leq 97.8^\circ$. The onset of the S wave on the transverse component (SH) is around 4 s before that of the radial component (SV). Because there is minimal energy on the transverse component for the SKS arrival, it appears that negligible upper mantle anisotropy affects the signal. Hence the authors conclude that the two components have experienced different velocities in the lowermost mantle ($V_{SH} > V_{SV}$).

Figure 7: Summary of previous studies of D'' anisotropy. Numbered regions corresponding to Table 1 are shown in outline, plotted on top of a global tomographic model of V_S at 2750 km (Becker and Boschi, 2002) (colour indicates the variation away from PREM (Dziewonski and Anderson, 1981) as per the legend). Regions where the dominant signal is $V_{SH} > V_{SV}$ are shown in blue; those where $V_{SH} < V_{SV}$ are in purple. Where a region is shown with red and blue stripes, both situations have been seen, as well as isotropy. Yellow areas indicate regions where the orientation of an assumed TTI fabric has been determined: this symbol shows the dip direction of the plane of isotropy with a tick of varying length, as shown in the legend (longer is steeper dip). In regions where one azimuth of raypaths show fast directions which are not CMB-parallel or -perpendicular, they also have a dip symbol as for the TTI regions, with the long bar parallel to the ray path in D'' . Regions with no fill show isotropy, and grey-filled regions show complex isotropy, either from SKS-SKKS differential splitting (see Table 1), or because no studies comparing V_{SH} to V_{SV} have been undertaken.

Figure 8: Average depth profile of $\xi = V_{SH}^2/V_{SV}^2$ from the SAW642AN model of Panning and Romanowicz (2006) (red) and S362WMANI of Kustowski et al. (2008) (blue). For SAW642AN The uppermost and lowermost mantle show $\xi > 1$, whilst most of the lower mantle is approximately isotropic. S362WMANI does not show the same dominant signal in D'' .

Figure 9: Example of a shear wave splitting measurement, slightly modified from Supplementary Figure 3 of Nowacki et al. (2010). The measurement is made at FCC (Fort Churchill, Manitoba, Canada) on the ScS phase from an 645 km-deep earthquake beneath Brazil at 13:27 on 21 July, 2007, and pre-corrected for upper mantle anisotropy beneath the receiver. Panel A shows the original three-component seismogram, with the predicted ScS arrival time for a 1-D global velocity model, and the arrival itself. Second panel (B) shows the horizontal components when rotated to the fast orientation ϕ , as found in the analysis, before and after time-shifting the slow component forward by the delay time found in the analysis. Lower left (C) shows the fast and slow waves before (upper left) and after (upper right) shifting by δt . The lower subpanels show the horizontal particle motion before and after correction with the optimum ($\phi, \delta t$). Last panel (D) shows the λ_2 surface (corresponding to misfit) in ϕ - δt space, with the optimum splitting parameters given by the blue cross, and surrounding 95% confidence interval (thick contour). Subplots to the right show the result of cluster analysis (Teanby et al., 2004)—the single cluster shows this is a stable result.

Figure 10: Comparison of SH-SV traveltime analysis and shear wave splitting for a transversely isotropic (TI) medium. On the left (A), the plane of isotropy is shown by the grey circle, dipping at an angle from the horizontal. This defines the orientation of the anisotropy. The ray frame fast orientation of the split shear wave, ϕ' , is controlled by the angle between the ray and the dip direction of the plane of isotropy, α , so that ϕ' is along the line of intersection between the plane of isotropy and the plane normal to the ray path. On the right (B) is shown the radial (R) and transverse (T) components of the split shear wave for various ϕ' . For all cases $\delta t = 1.5$ s, as shown by the dashed lines. Measuring the delay time directly on the two components only gives the correct amount and orientation of splitting for the special cases of $\phi' = 0^\circ$ or 90° . Within $\sim 15^\circ$ of 0 or 90° , such measurements are still useful for detecting the presence of anisotropy, but do not provide much information about the symmetry. Slightly modified from Wookey and Kendall (2007).

Figure 11: Structure of MgSiO₃-perovskite and -post-perovskite. Yellow spheres are Mg ions; SiO₆ octahedra are shown in blue.

Figure 12: Proportions of phases present in the lower mantle for pyrolite and MORB compositions (after Ono and Oganov (2005) and Hirose (2006), and partly based on Trønnes (2010)). Yellow regions show aluminous phase regions, whilst grey regions show phases of silica. Sloping phase boundaries represent the range of depths over which the transition between the phases probably occurs. Ca(Fe,Ti)₂O₄-type Al-bearing phase refers to the uncertainty over the structure of the phase. Abbreviations are: Ca-pv: CaSiO₃-perovskite; pv: (Mg,Fe)(Si,Al)SiO₃-perovskite; ppv: (Mg,Fe)(Si,Al)SiO₃-post-perovskite; st: staurolite; α -PbO₂: SiO₂ in the α -PbO₂ form (also called columbite structure).

Figure 13: Elastic P and S wave anisotropy for pv from calculations at lower mantle conditions. (Top: Wookey et al. (2005b); bottom: Wentzcovitch et al. (2006).) Plots on the left show upper hemisphere, equal area projections of V_P with direction within the orthorhombic crystal. The 1, 2 and 3 axes are shown, corresponding to the [100], [010] and [001] directions respectively: 1 is up, 2 is left and 3 is out of the page. Colour indicates V_P as shown in the scale bar at the bottom. Plots on the right show δV_S (colour as per the scale bar) and the fast shear wave orientation with direction (black ticks). Because of the orthorhombic symmetry, each plot only varies within each quadrant.

Figure 14: Elastic P and S wave anisotropy for ppv from experiments and calculations at $T = 4000$ K (top to bottom: Stackhouse et al., 2005b; Wentzcovitch et al., 2006; Mao et al., 2010). Features as for Figure 13.

Figure 15: Elastic P and S wave anisotropy for fpc from *ab initio* calculations and experiment at lower mantle conditions. The three axes (1, 2 and 3) each corresponds to the $\langle 100 \rangle$ directions—because of the cubic symmetry the plots only vary within each eighth of the upper hemisphere.

Figure 16: Lattice preferred orientation (LPO) of crystals (A) and shape preferred orientation (SPO) of prolate (B) and oblate (C) slower isotropic inclusions in a faster anisotropic matrix (schematic). Spheres above are 3-D versions of the plots explained in Figure 3. They show the amount of shear wave anisotropy δV_S by colour, and the fast shear wave orientation by black ticks. Note that the colour scales are different. Blue arrows show a direction of flow which may align the crystals or inclusions, and thus how this might be interpreted from measuring the anisotropy.

Figure 17: Inferred TTI planes beneath the Caribbean, taken from Supplementary Information to Nowacki et al. (2010). The bar symbols show the direction of dip with the short tick, with the dip in degrees of the plane of isotropy given by the numbers. Beneath, colour shows the variation of V_S in the S20RTS model (Ritsema et al., 1999) at 2750 km depth. The coloured areas labelled ‘W’, ‘S’ and ‘E’ show the approximate horizontal region of sensitivity of ScS at 2750 km. Thin black lines show individual raypaths of ScS in the bottom 250 km of the mantle.

Figure 18: TTI plane of isotropy in region ‘E’ of Nowacki et al. (2010), shown by schematic layering of the material. Rays from South America travelling north show $\phi' \approx 90^\circ$, whilst those from the Mid-Atlantic Ridge (MAR) travelling northwest exhibit $\phi' = 45^\circ$. Assuming hexagonal symmetry where $\delta \approx \epsilon$, the fast orientation is in the plane of isotropy in each case. Whilst TTI is a possible explanation, it is only one type of anisotropy which can produce the observations with two azimuths of waves.

Figure 19: Shear wave anisotropy for horizontal (left) and inclined (right) melt inclusions in D'' . The cartoons below show the alignment of oblate spheroids which respond to the motion of the mantle differently. In both cases, the sense of shear is top to the north (approximately right here), shown by the arrow. On the left, the inclusions are aligned parallel to the horizontal flow and produce VTI. On the right, the melt inclusions dip at 25° towards the sense of shear, opposite the sense of flow. For most azimuths of horizontally-propagating shear waves, this produces splitting with the fast orientation parallel to the alignment of the oblate inclusions. As discussed in the text, this is compatible with observations beneath Siberia and the Caribbean. The elastic constants are calculated using effective medium theory (Tandon and Weng, 1984) for an arbitrary set of lowermost mantle-like properties (matrix: $V_P = 14 \text{ km s}^{-1}$, $V_S = 7.3 \text{ km s}^{-1}$, $\rho = 5500 \text{ kg m}^3$; inclusions: $V_P = 7 \text{ km s}^{-1}$, $V_S = 0 \text{ km s}^{-1}$, $\rho = 5500 \text{ kg m}^3$, aspect ratio = 0.01, volume fraction = 0.005).

Figure 20: Variation of shear wave splitting with direction for MgSiO_3 post-perovskite (elastic constants of Stackhouse et al. 2005b at 3000 K). Colour indicates the strength of shear wave anisotropy in a given direction (δV_S) as per the scale bar. The black bars show the orientation of the fast shear wave. The crystallographic directions are indicated. (A) Shear wave splitting for unaltered single-crystal constants. There is strong ($\delta V_S = 20\%$) anisotropy for rays along $[100]$ and $\langle 111 \rangle$. (B) Anisotropy for a planar average of the constants when rotated around $[001]$. Strong ($\delta V_S = 15\%$) splitting occurs within the plane normal to $[001]$, with fast directions also in the plane. However, this corresponds to an aggregate of perfect alignment of $[001]$ directions of pure ppv, which does not occur in D'' .

Figure 21: Upper hemisphere diagrams showing shear planes and slip directions which are compatible with the measurements of sub-Caribbean D'' shear wave splitting of Nowacki et al. (2010). The schematic diagram on the left shows how to interpret the diagrams on the right: they show the upper hemisphere projection of the slip plane (coloured lines) and slip direction (black dots), hence the centre of the plots corresponds to the vertical direction; in this case the top of the diagrams is north. The elastic constants tested are those of Yamazaki et al. (2006), who deform ppv to produce an aggregate consistent with the dominant slip system in the crystal of $[100](010)$. Three regions ('W', 'S' and 'E') are shown. Lighter colours show that more alignment of the phase via the slip system is required to produce the observed splitting—larger splitting times are observed in region S. For these constants, orientations of the shear plane dipping south or southeast can produce the observed splitting in regions S and E; horizontal shear can explain the splitting in region W.

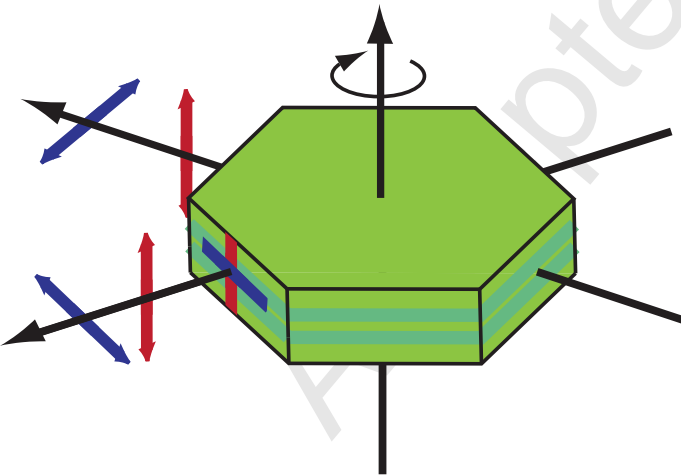
Figure 22: Flow chart showing the progression of calculations and assumptions required to predict flow from measurements of shear wave splitting.

Table 1: Summary of previous studies of anisotropy in the lowermost mantle.

Table 2: Summary of inferred slip systems in MgSiO_3 post-perovskite and structural analogues from deformation experiments using the diamond-anvil cell (DAC), laser-heated diamond-anvil cell (LHDAC), Kawai-type and deformation-DIA (D-DIA) apparatuses.

Figure 1

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VTI

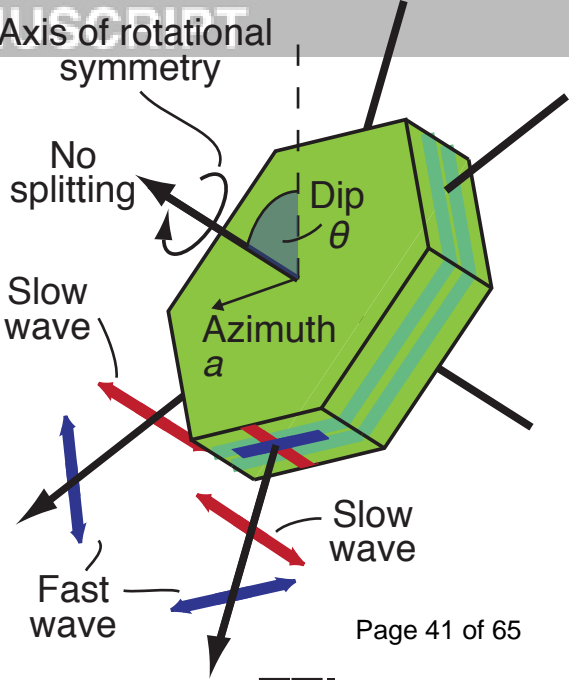
Axis of rotational symmetry

No splitting

Slow wave

Fast wave

Slow wave



TTI

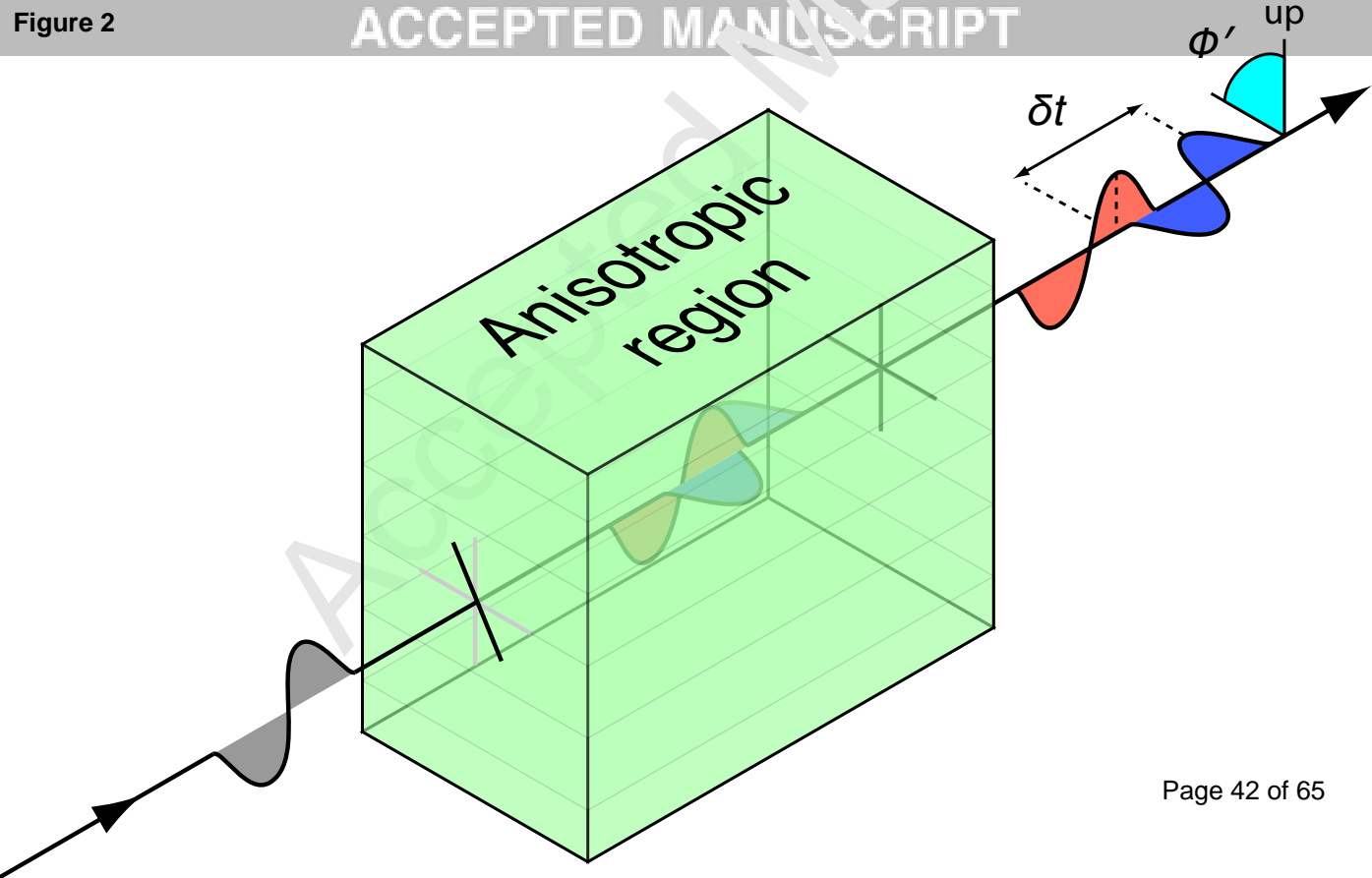


Figure 3

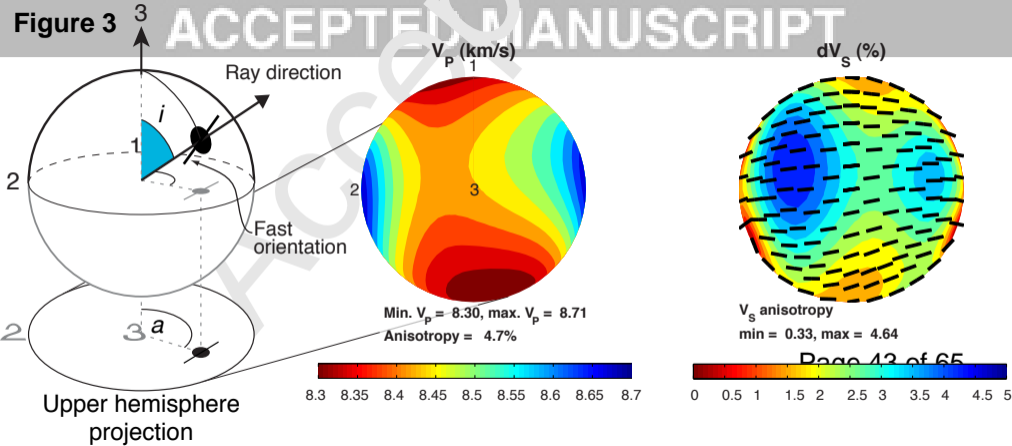


Figure 4

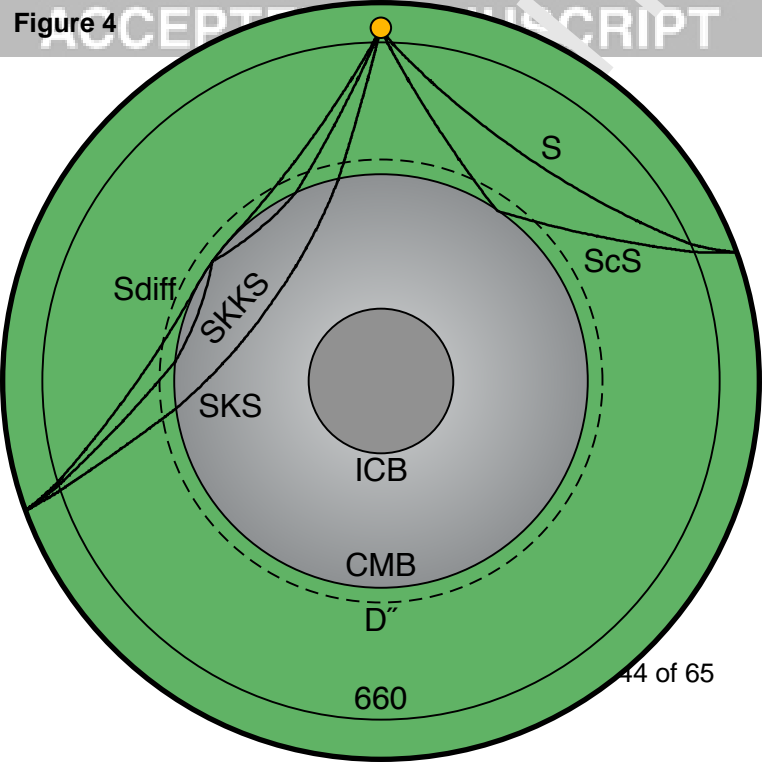
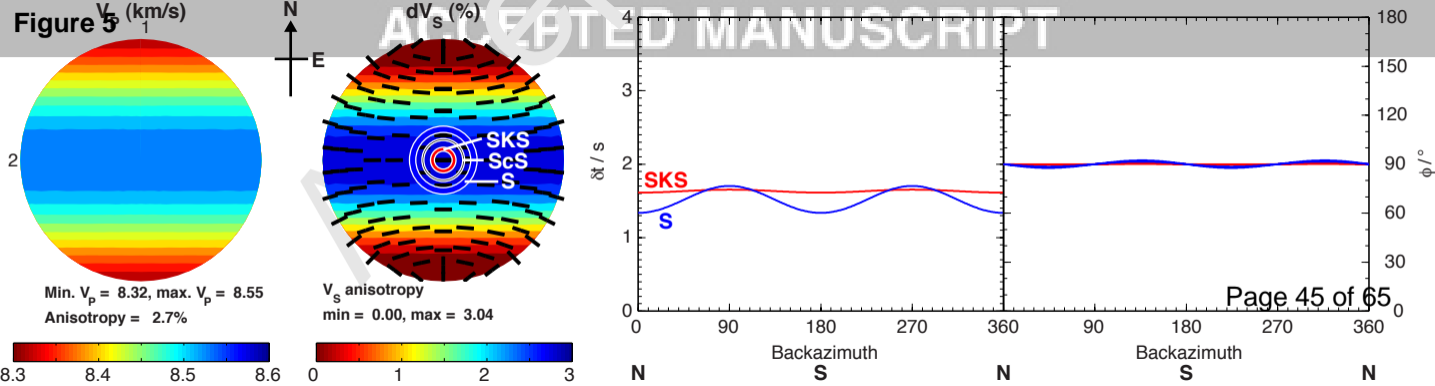
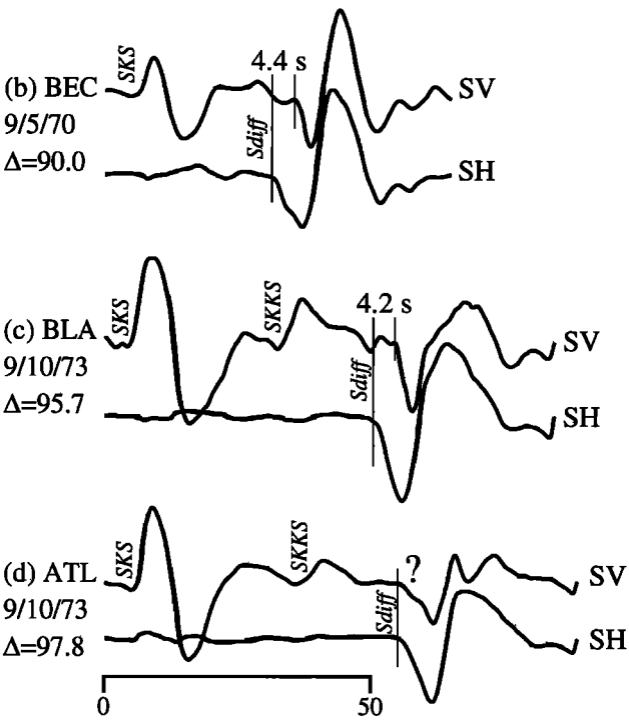
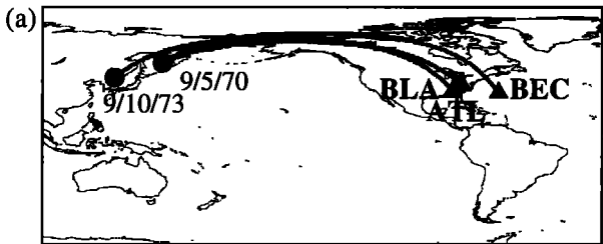
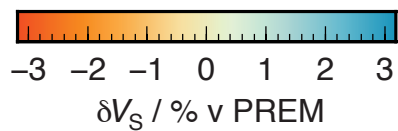
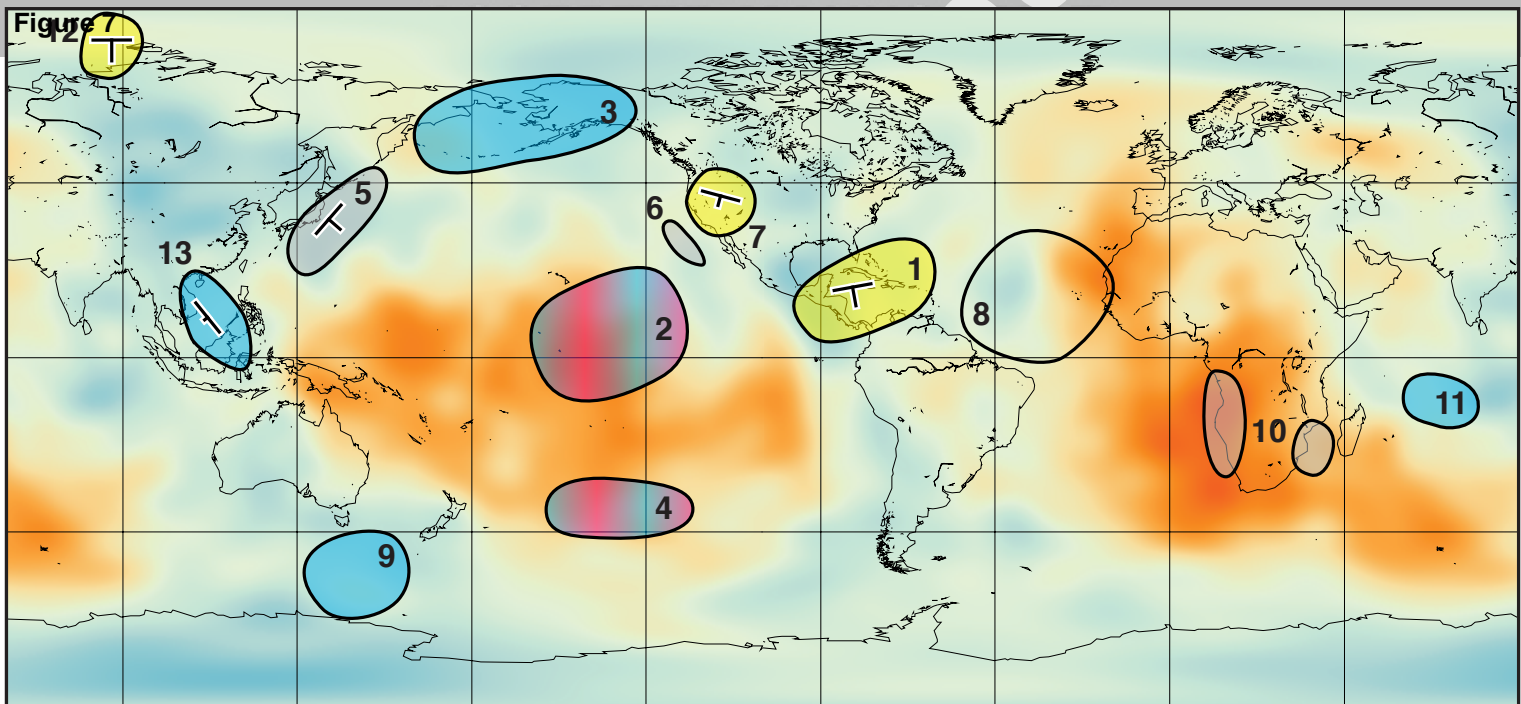


Figure 5







$V_{SH} > V_{SV}$

$V_{SH} < V_{SV}$

$V_{SH} > V_{SV}$
 $V_{SH} < V_{SV}$
 $V_{SH} \approx V_{SV}$

TTI:
 dip
 60°

TTI:
 dip
 40°

TTI:
 dip
 20°

See text

$V_{SH} \approx V_{SV}$
 Page 47 of 68

Figure 8⁰

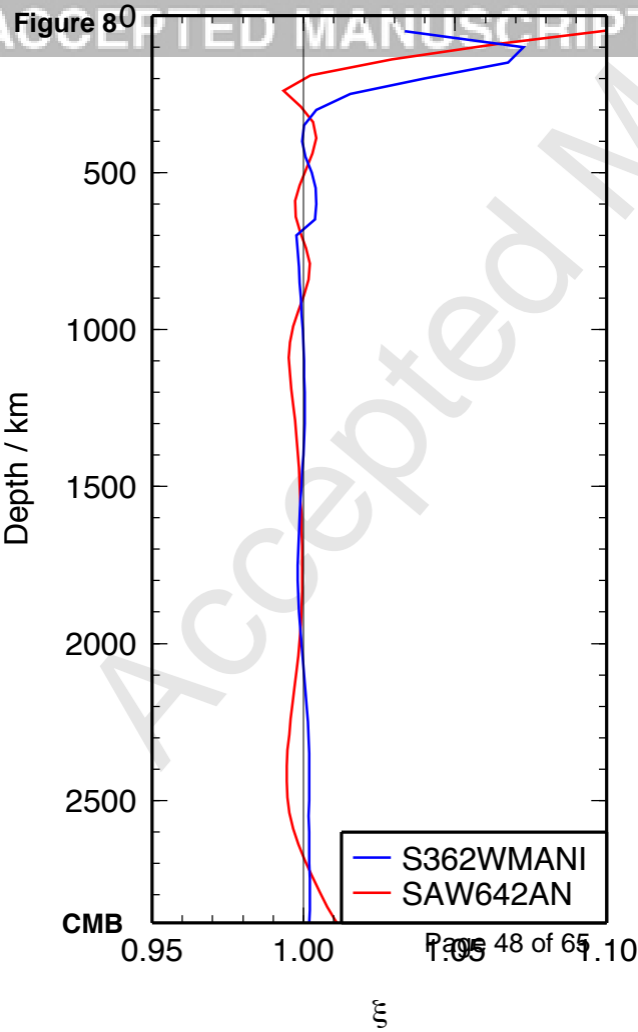
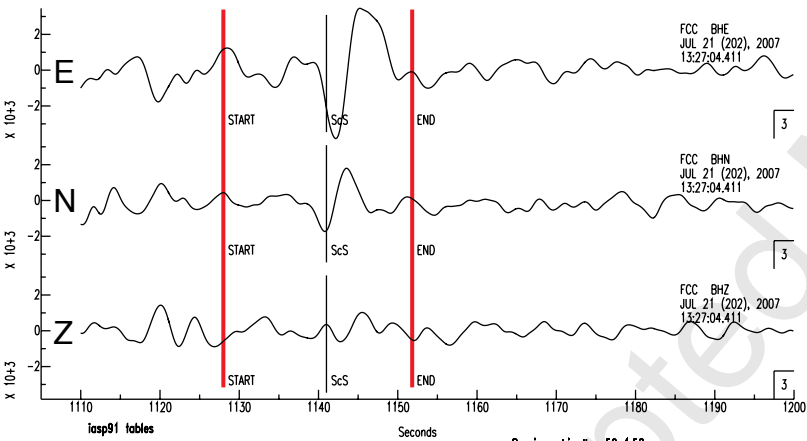


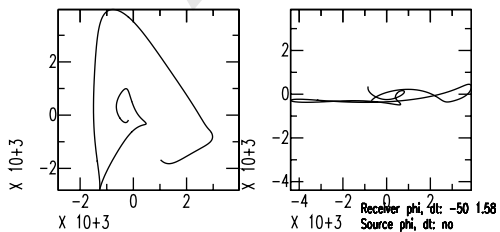
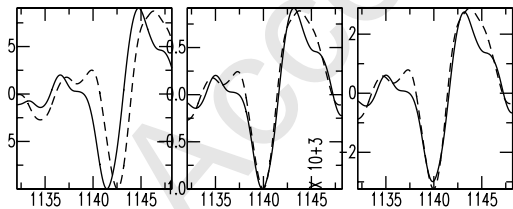
Figure 9

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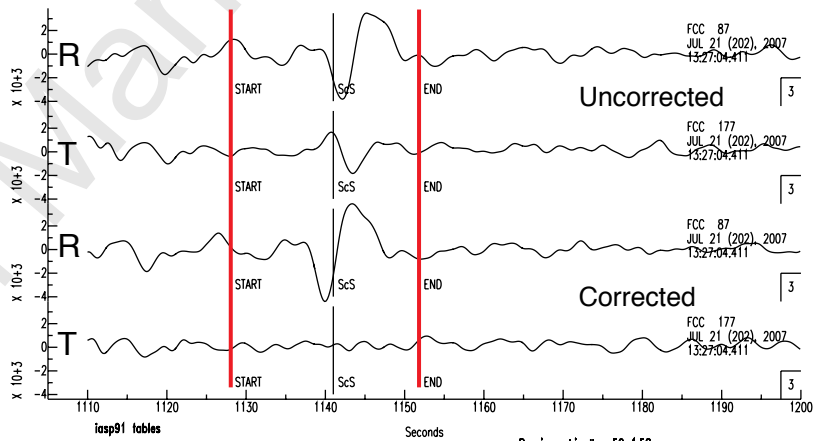
c

12 FCC PAS1 59.0000+/-3.7500 ILAG 2.5750+/-0.0813
 SPOL 87.2345+/-0.5918



b

Event:2007.202 Sta:FCC Dist:69.2 Az:347.5 Baz:155.7 -8.133N -71.272E 644.9km



d

2007.202.13.2 $\delta t = 2.5750 \pm 0.0813$ (s)
 fast = 39.0000 ± 3.7500 ($^\circ$)
 spol = 87.2345 ± 0.5918 ($^\circ$)

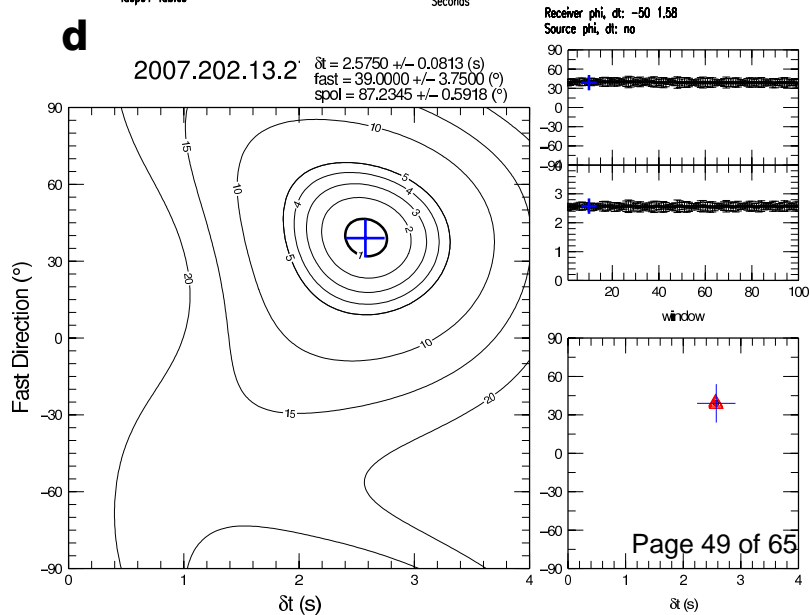
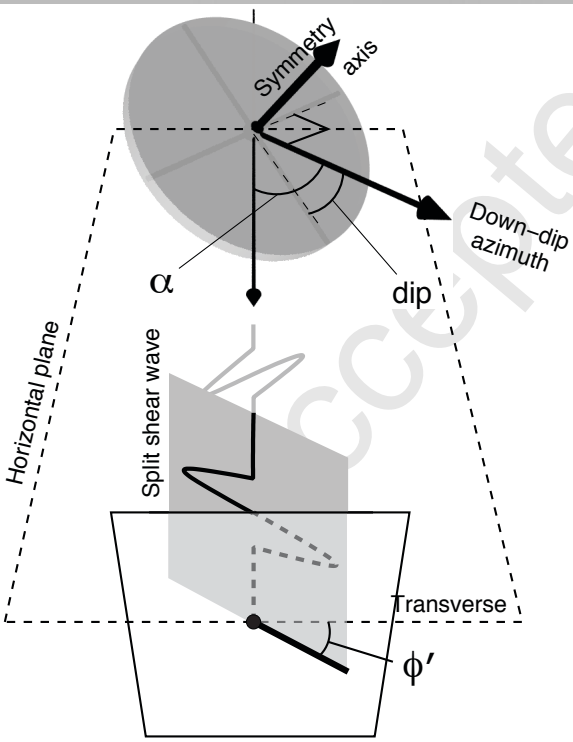


Figure 10



B

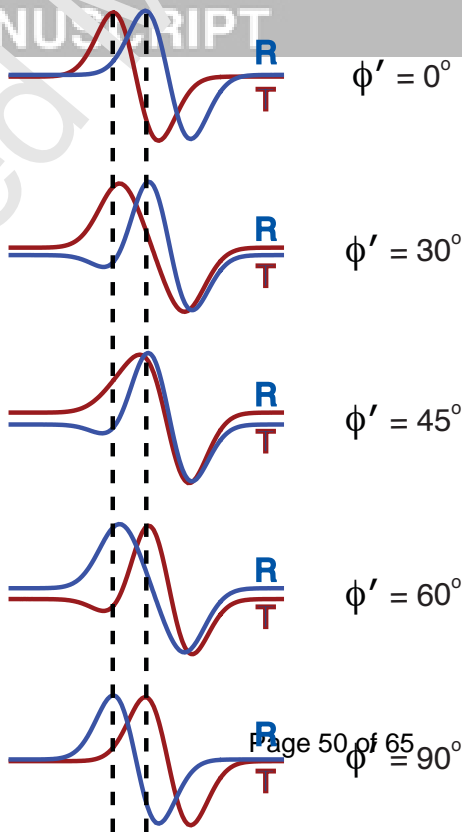
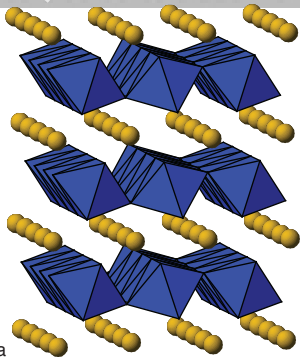
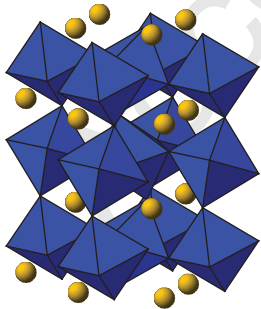
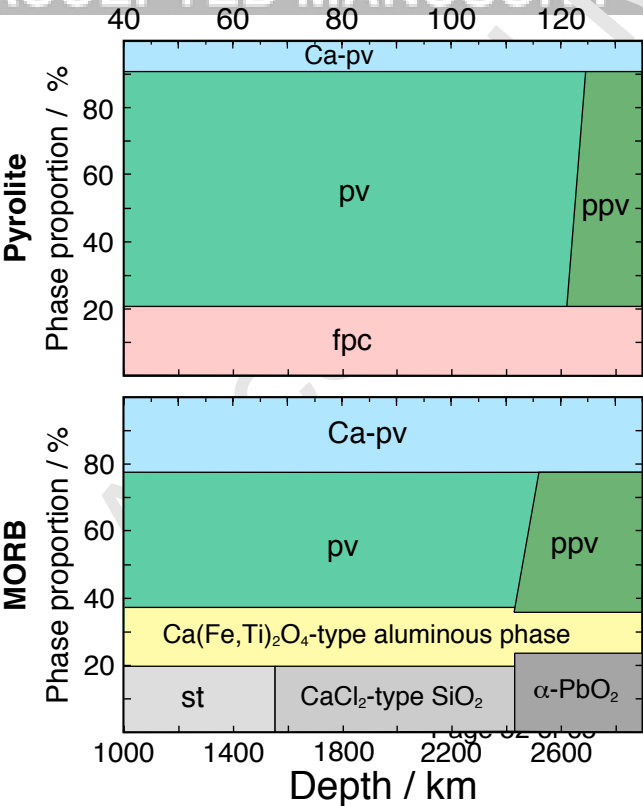


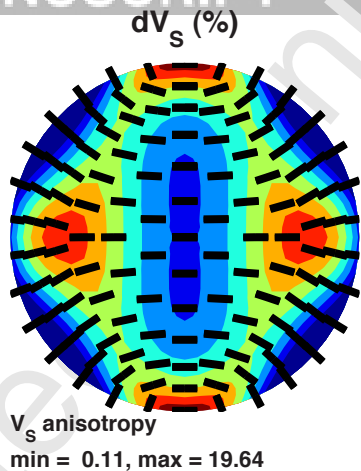
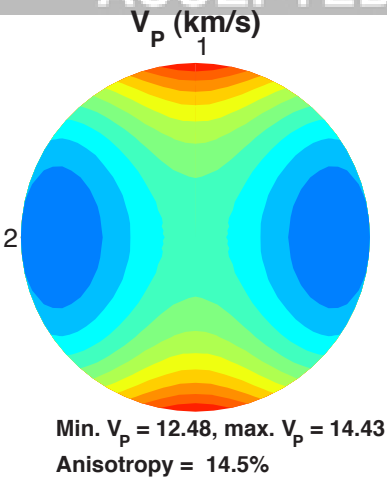
Figure 11 MgSiO_3 pv

MgSiO_3 ppv

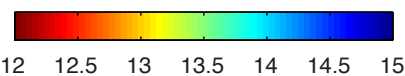
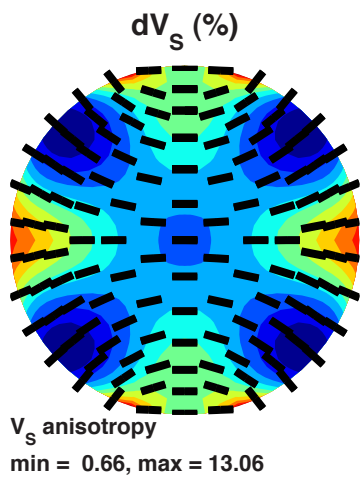
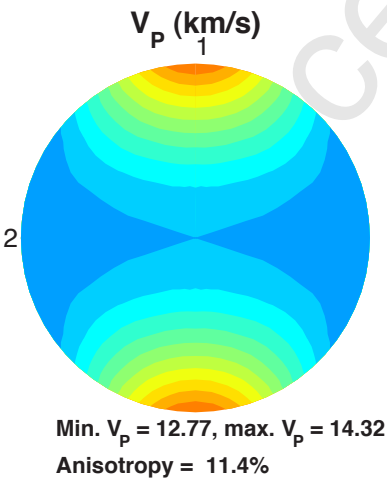
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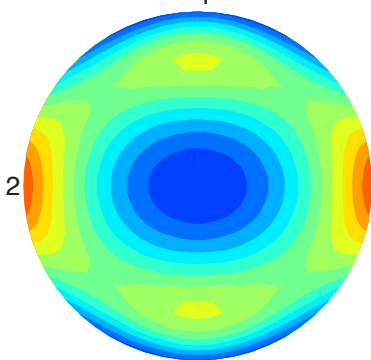




MgSiO₃, Wentzcovitch et al., 2006, P=125 GPa, T=2500 K

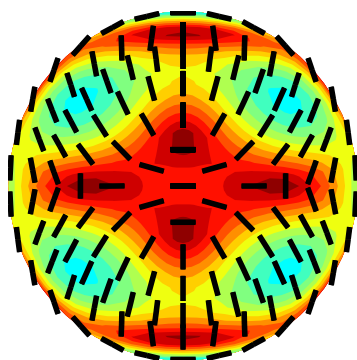


V_p (km/s)



Min. $V_p = 12.69$, max. $V_p = 14.66$
Anisotropy = 14.4%

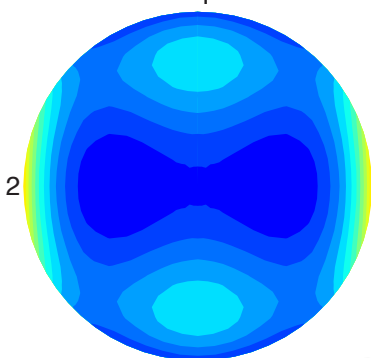
dV_s (%)



V_s anisotropy
min = 0.93, max = 24.41

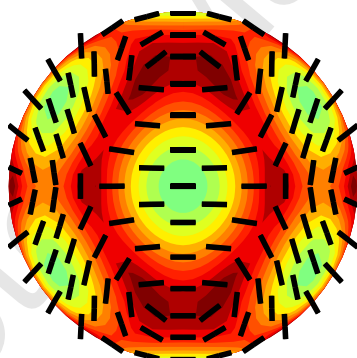
MgSiO_3 , Wentzcovitch et al., 2006, $P=140$ GPa, $T=4000$ K

V_p (km/s)



Min. $V_p = 13.08$, max. $V_p = 14.79$
Anisotropy = 12.3%

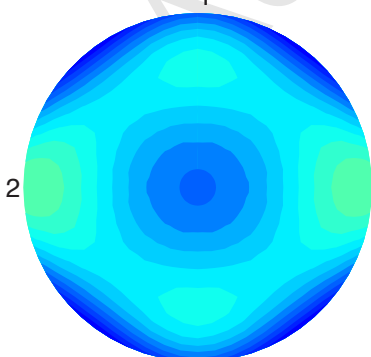
dV_s (%)



V_s anisotropy
min = 0.35, max = 19.70

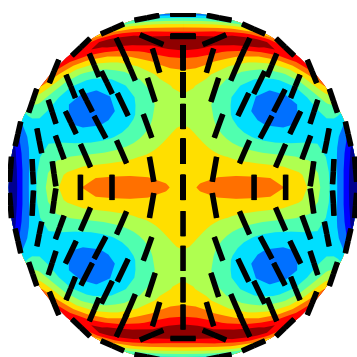
$(\text{Mg}_{0.6}\text{Fe}_{0.4})\text{SiO}_3$, Mao et al., 2010, $P=140$ GPa, $T=2000$ K

V_p (km/s)

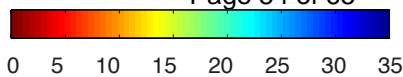
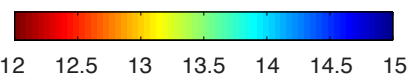


Min. $V_p = 13.65$, max. $V_p = 14.75$
Anisotropy = 7.7%

dV_s (%)

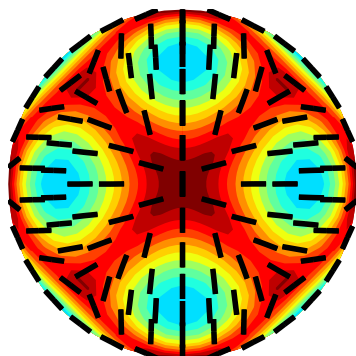
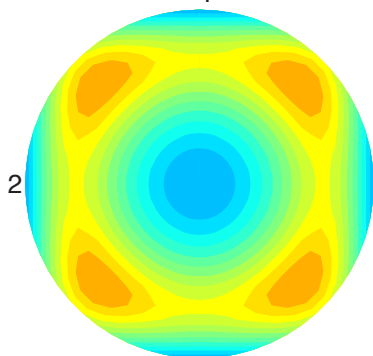


V_s anisotropy
min = 0.80, max = 42.31



V_p (km/s)
1

dV_s (%)



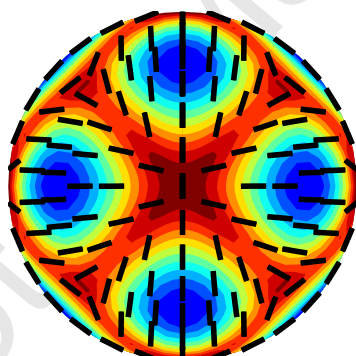
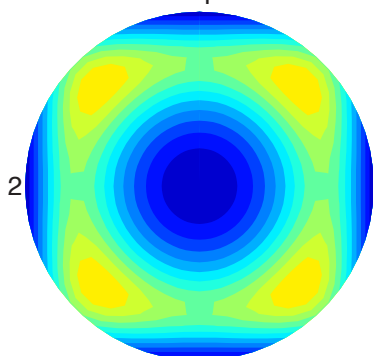
Min. V_p = 13.06, max. V_p = 15.21
Anisotropy = 15.2%

V_s anisotropy
min = 0.00, max = 36.91

MgO, Oganov & Dorogokupets, 2003, P=150 GPa, T=1398 K

V_p (km/s)
1

dV_s (%)



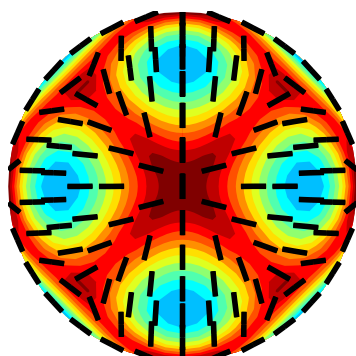
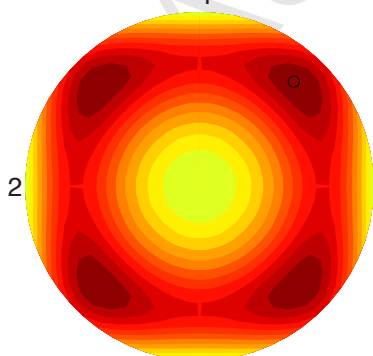
Min. V_p = 13.36, max. V_p = 16.41
Anisotropy = 20.5%

V_s anisotropy
min = 0.00, max = 48.73

(Mg_{0.9}Fe_{0.1})O, Marquardt et al., 2009, P=80 GPa, T=0 K

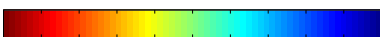
V_p (km/s)
1

dV_s (%)

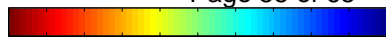


Min. V_p = 11.63, max. V_p = 13.75
Anisotropy = 16.7%

V_s anisotropy
min = 0.00, max = 38.45



12 13 14 15 16



0 10 20 30 30 50

Figure 16

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A S-wave Anisotropy

0% 3%

B S-wave Anisotropy

0% 0.5%

C S-wave Anisotropy

0% 5%

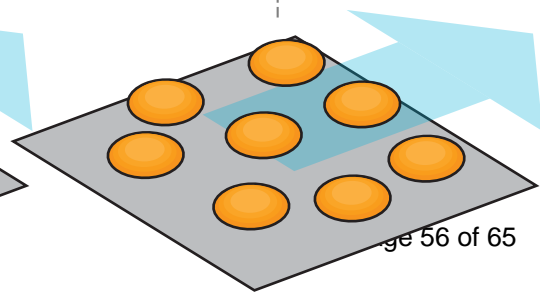
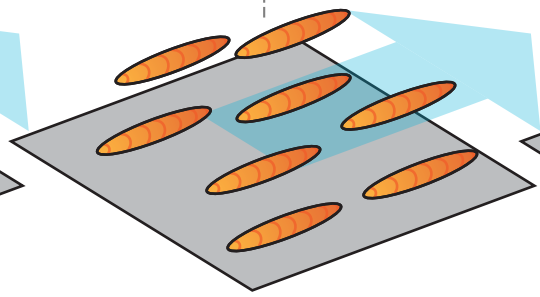
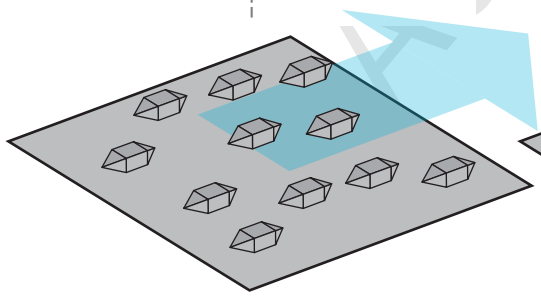
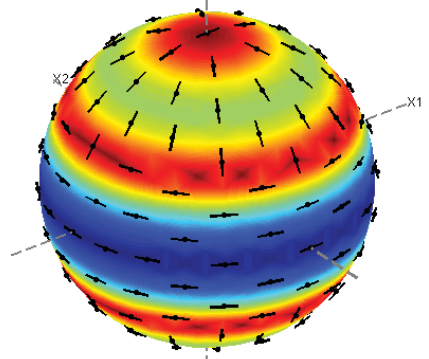
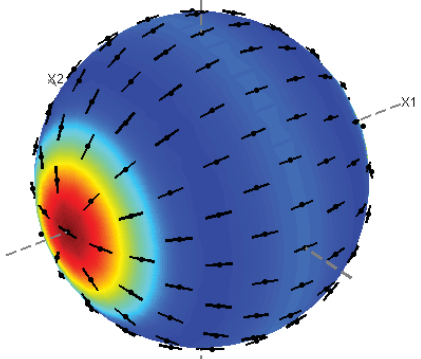
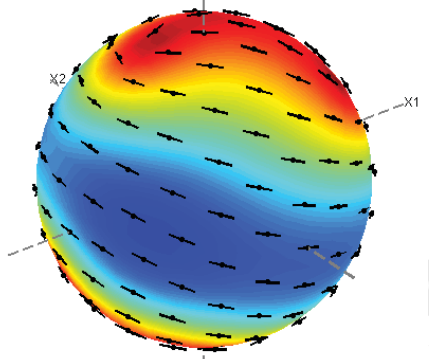
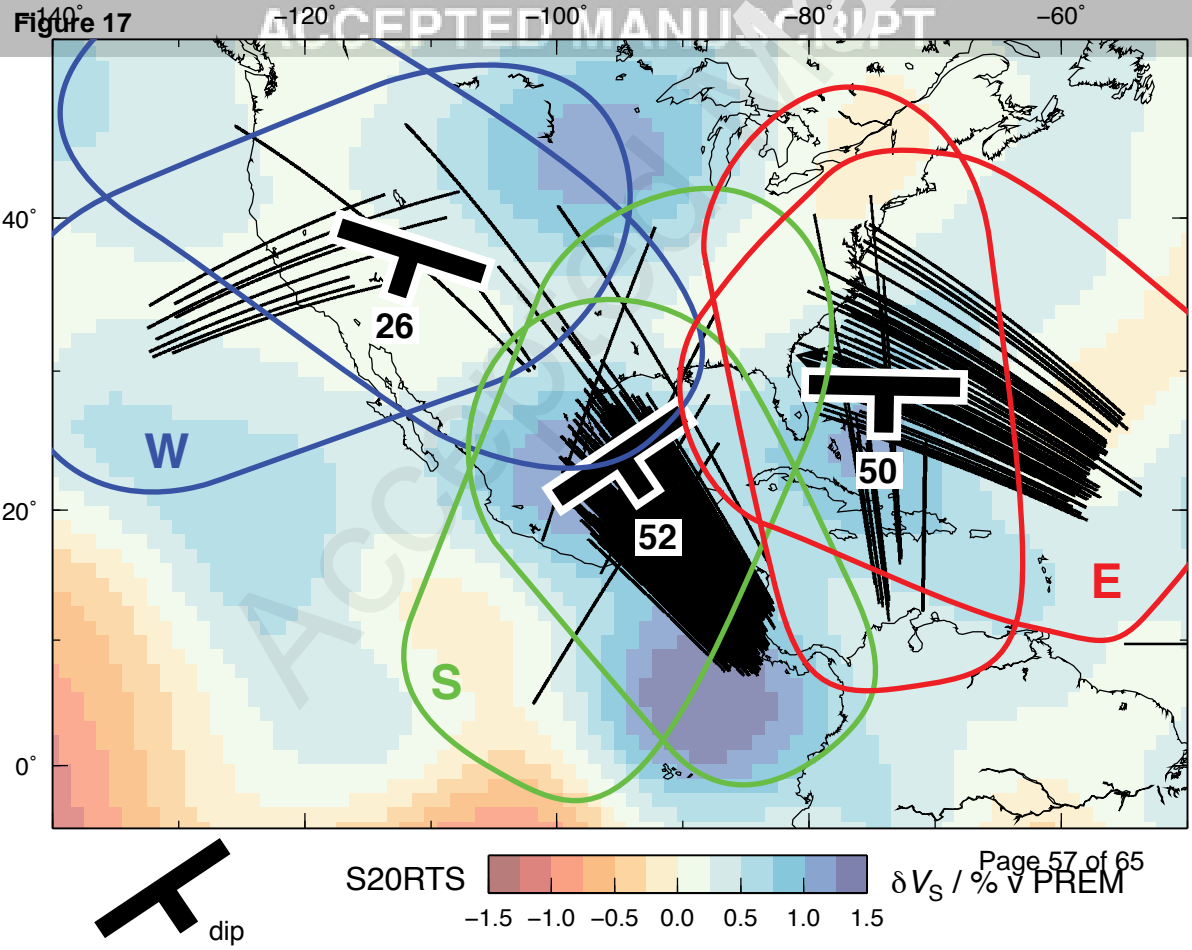


Figure 17

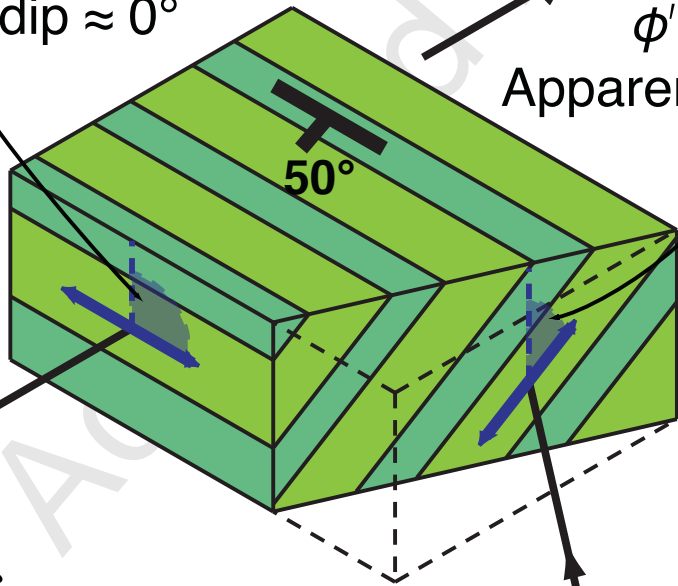


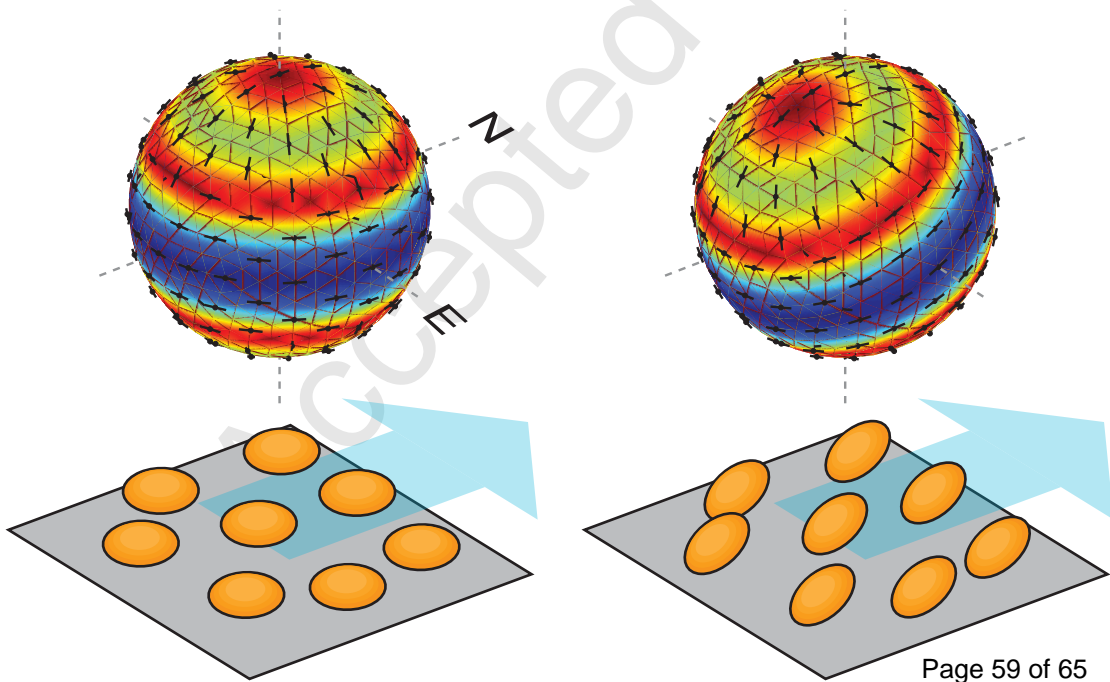
Apparent dip $\approx 0^\circ$

$\phi' \approx 45^\circ$
Apparent dip $\approx 45^\circ$

Rays from
South America

Rays from
MAR



0%  3%

Horizontal melt inclusions

Inclined melt inclusions

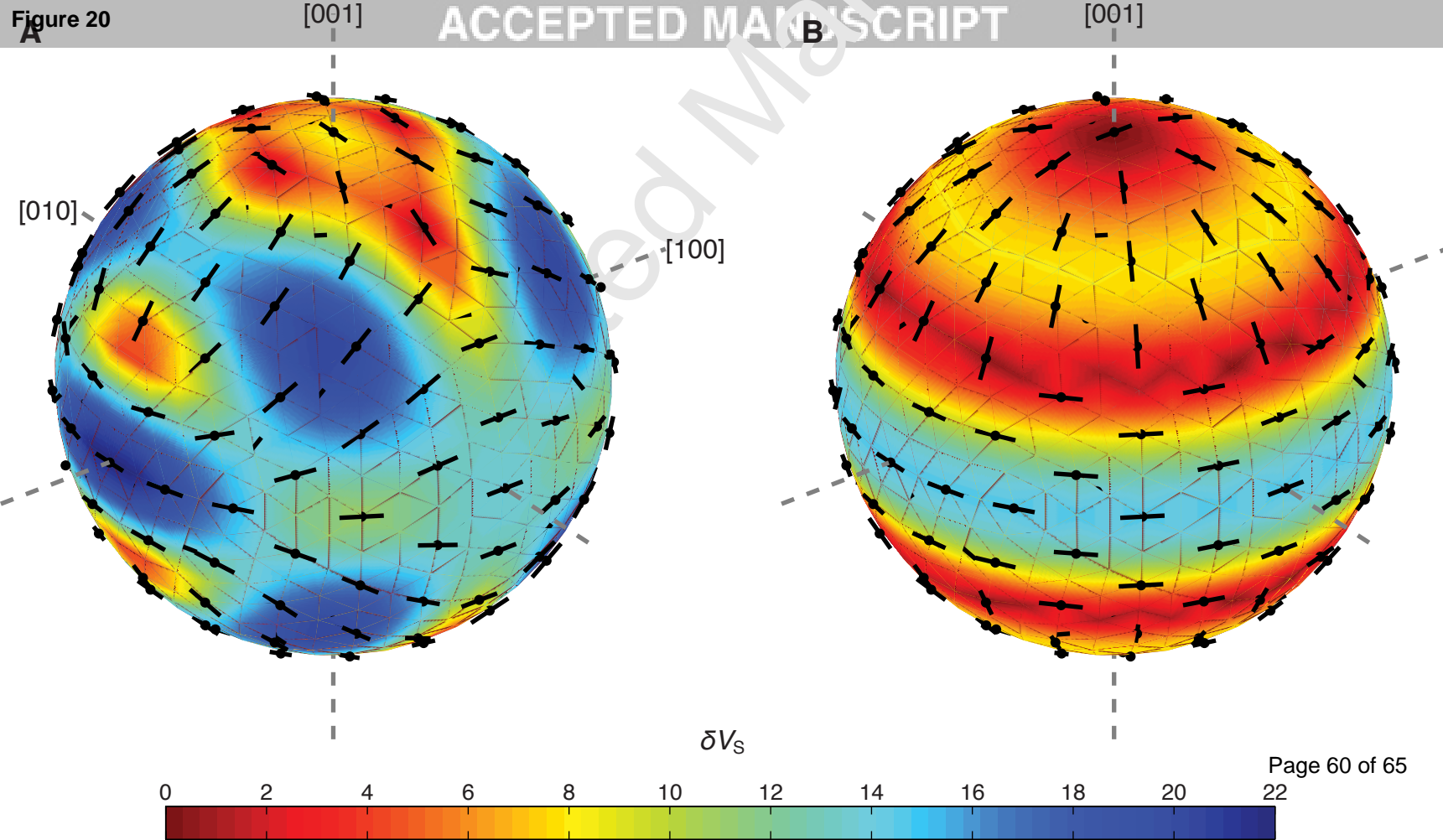
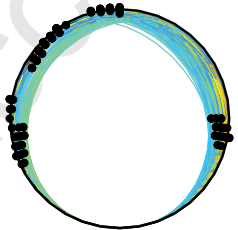
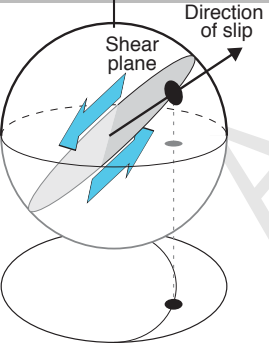


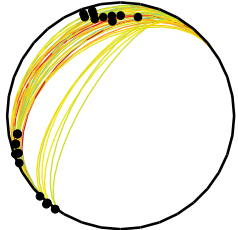
Figure 21

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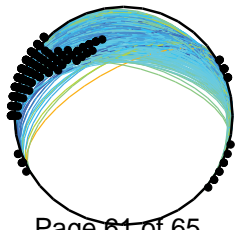
MgSiO₃-ppv
[100](010) slip system



W



S



E

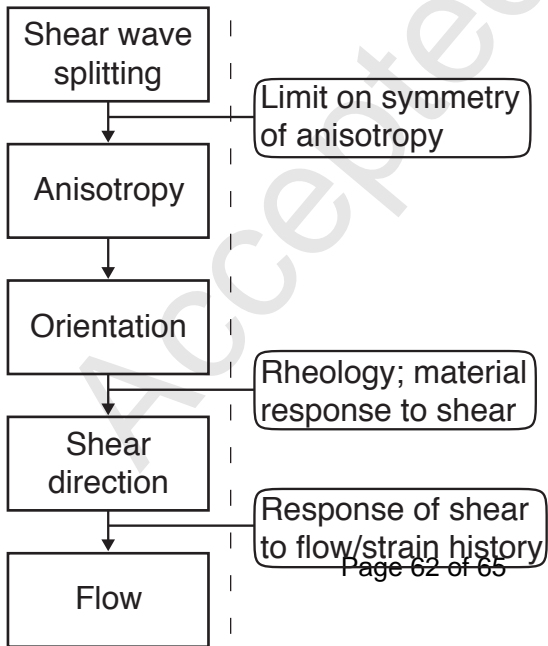


Table 1: Summary of previous studies of anisotropy in the lowermost mantle.

Study	Phases used	Observation	$\delta V_S / \%^a$	Suggested style of anisotropy
1. Caribbean				
Lay and Helmberger (1983)	ScS	$V_{SH} > V_{SV}$	5	Isotropic velocity structure
Kendall and Silver (1996)	S,Sdiff	$V_{SH} > V_{SV}$	1.8	VTI
Ding and Helmberger (1997)	ScS	$V_{SH} > V_{SV}$	2.5	VTI
Rokosky et al. (2004)	ScS	$V_{SH} > V_{SV}$	0.6	VTI
Garnero et al. (2004a)	S,ScS,Sdiff	$\leq 20^\circ$ dip east-west		TTI
Maupin et al. (2005)	S,ScS,Sdiff	$\leq 20^\circ$ dip east-west	1.5–2.2	TTI
Rokosky et al. (2006)	ScS	Mostly $V_{SH} > V_{SV}$	0.0–2.0 ^{b,c}	Varying TTI
Nowacki et al. (2010)	ScS	$\sim 50^\circ$ dip \sim south	0.8–1.5	TTI or orthorhombic
2. Central Pacific				
Vinnik et al. (1995)	Sdiff	$V_{SH} > V_{SV}$	0.6 ^b	VTI
Vinnik et al. (1998)	Sdiff	$V_{SH} > V_{SV}$	~ 10	VTI
Pulliam and Sen (1998)	S	$V_{SH} < V_{SV}$	–2	VTI
Ritsema et al. (1998)	S,Sdiff	$V_{SH} < V_{SV}$	–2.1––1.4	VTI
Kendall and Silver (1998)	S,Sdiff	$V_{SH} \approx V_{SV}$		Isotropic
Russell et al. (1998, 1999)	ScS	$V_{SH} > V_{SV}, V_{SH} < V_{SV}$	2–3	VTI
Fouch et al. (2001)	S,Sdiff	$V_{SH} > V_{SV}$	0.3–5.3	VTI
Kawai and Geller (2010)	S,ScS,SKS	$V_{SH} < V_{SV}$	–3	VTI
3. Alaska				
Lay and Young (1991)	S,ScS,Sdiff	$V_{SH} > V_{SV}$		VTI
Matzel et al. (1996)	S,ScS,Sdiff	$V_{SH} > V_{SV}$	1.5–3	VTI
Garnero and Lay (1997)	S,ScS,Sdiff	Mainly $V_{SH} > V_{SV}$	–1–3	VTI
Wyssession et al. (1999)	Sdiff	$V_{SH} > V_{SV}$	0.2–0.6	VTI or TTI
Fouch et al. (2001)	S,Sdiff	$V_{SH} > V_{SV}$	0–0.9	VTI
4. South East Pacific				
Ford et al. (2006)	S,Sdiff	$V_{SH} > V_{SV}, V_{SH} < V_{SV}$	–1.0–0.9	VTI
5. North West Pacific				
Wookey et al. (2005a)	ScS	$\sim 40^\circ$ dip southeast	0.8–2.3	TTI
6. East Pacific				
Long (2009)	SKS-SKKS	Differential $\delta t \approx 2$ s ^d	0.5 ^b	TTI
7. Western USA				
Nowacki et al. (2010)	ScS	26° dip southwest	1.2	VTI or TTI
8. Atlantic Ocean				
Garnero et al. (2004b)	S,Sdiff	$V_{SH} \approx V_{SV}$	≤ 0.5	Isotropy or weak VTI
9. Antarctic Ocean				
Usui et al. (2008)	S	$V_{SH} > V_{SV}$	1 ^b	VTI
10. Southern Africa				
Wang and Wen (2007)	SKS-SKKS	Differential $\delta t \approx 1$ s ^d	$\sim 2^b$	Varying HTI
11. Indian Ocean				
Ritsema (2000)	S	$V_{SH} > V_{SV}$	1.4–1.7	VTI
12. Siberia				
Thomas and Kendall (2002)	S,ScS,Sdiff	Mainly $V_{SH} > V_{SV}$	–0.8–1.4	Mainly VTI
Wookey and Kendall (2008)	ScS	55° dip \sim south	0.7–1.4	TTI or orthorhombic
13. Southeast Asia				
Thomas et al. (2007)	ScS	9° dip southwest	0.5	VTI or TTI

^a +ve: $V_{SH} > V_{SV}$; –ve: $V_{SH} < V_{SV}$ ^b Calculated from the study's stated δt using $\langle V_S \rangle$ from a global isotropic V_S model (Ritsema et al., 1999) for a uniform 250 km thick D'' layer.^c Upper limit on δt of 2.5 s imposed.^d Differential δt refers to $\delta t_{SKKS} - \delta t_{SKS}$.

Table 2: Summary of inferred slip systems in MgSiO_3 post-perovskite and structural analogues from deformation experiments using the diamond-anvil cell (DAC), laser-heated diamond-anvil cell (LHDAC), Kawai-type and deformation-DIA (D-DIA) apparatuses.

Study	Method	P (GPa)	T (K)	Differential stress (GPa)	Dominant slip system ^a	Remarks
(Mg, Fe)SiO₃						
Merkel et al. (2007)	LHDAC	145–157	1800	7–9	(100) or (110)	Mg#=0.9; opx starting material
Miyagi et al. (2010)	LHDAC	148–185	3500	5–10	[100](001) or [010](001)	Mg#=1.0; glass starting material
Mao et al. (2010)	LHDAC	140	2000	^b	{100} or {110}	Mg#=0.6; opx starting material
CaIrO₃						
Yamazaki et al. (2006)	Kawai	1	1173		[100](010)	$\gamma=0.4-1$ ^c
Walte et al. (2007)	D-DIA	3	1000	^b	[100](010)	$\gamma=0.8-1$
Niwa et al. (2007)	DAC	0–6	300	^b	(010)	
Miyagi et al. (2008)	D-DIA	2–6	300–1300	–2–2	[100](010)	
Walte et al. (2009)	D-DIA	1-3	1300	^b	[100]{010}	$\gamma=0.5-1$
MgGeO₃						
Merkel et al. (2006)	LHDAC	104–124	1600	3–8	(100) or (110)	opx starting material
Kubo et al. (2008)	LHDAC	83–99	1600	0.1–1	(010)	opx starting material
Okada et al. (2010)	LHDAC	78–110	300	1–3	(001)	4 runs: opx and pv starting material
MnGeO₃						
Hirose et al. (2010)	LHDAC	77–111	2000	2–10	(001)	opx starting material

^a Where no slip vector is given in the study, only the slip plane is shown.

^b Not stated.

^c Shear strain γ as stated in the study.

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