Hemispherical structure in inner core velocity anisotropy

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We investigate the hemispherical pattern in inner core velocity anisotropy using a new and independent high-quality data set of PKPbc-PKPdf and PKPab-PKPdf body wave observations. Our data show no evidence for a tilted anisotropy axis with respect to the Earth’s rotation axis and is significantly better fit by a model with 4.4% anisotropy in the western hemisphere and only 1.0% in the eastern hemisphere than by a model of uniform anisotropy. We carry out variance minimization and find the boundaries between the hemispheres at lines of constant longitude at 14°E and 151°W. Variance minimization enables us to extract the imprint of hemispherical structure from all of the data, and not just polar paths, resulting in boundaries which are nearly 30° from the results of previous studies. The high quality of the data allows us to provide robust evidence that the isotropic velocity in the eastern hemisphere is 0.2% lower than in the western hemisphere in the top 660 km of the inner core. Our data set also suggests an increase in inner core anisotropy in the eastern hemisphere from 1% to around 6% at depths deeper than 660 km, indicating that the hemispherical pattern in anisotropy may disappear at greater depths. The presence of hemispherical structure rules out mechanisms for creating anisotropy which are unable to sustain longitudinal variations in the inner core. Furthermore, steady inner core superrotation of the order of 0.1°/year would eradicate the hemispherical differences, though inner core oscillation would still be permissible.


1. Introduction

Earth’s inner core, discovered in 1936 [Lehmann, 1936] is one of the most challenging regions of the Earth to study. A solid body, of radius 1221.5 km [Dziewonski and Anderson, 1981], it consists of an iron-nickel alloy, with a small amount of lighter elements. The composition of these light elements, together with the crystalline phase present in the inner core are still a matter of debate [see, e.g., Oganov et al., 2005]. Buoyancy of light elements released as the inner core solidifies, together with the release of latent heat during the solidification process, drive convection in the fluid outer core [Gubbins et al., 2003]. This convection sustains the geodynamo, which causes Earth’s magnetic field.

Seismology has provided much of the information upon which we base our understanding of the properties of the inner core. The inner core is anisotropic: the speed of seismic waves in the inner core is dependent on the direction of wave propagation (see Song [1997], Tromp [2001], and Souriau [2007] for reviews). The inner core may also be rotating relative to the rest of the Earth [Song and Richards, 1996] although this is currently an unresolved issue [Laske and Masters, 1999; Poupinet et al., 2000]. While progress is being made in imaging hemispherical inner core structures with free Earth oscillations [Irving et al., 2009; Deuss et al., 2010], our current understanding of the lateral velocity and anisotropy structures in the inner core is still largely dependent on measurements of body waves which traverse the inner core.

The first proposal that the inner core was not homogeneous was made by Poupinet et al. [1983], who found that rays which traveled through the inner core from the south pole to the north pole had shorter travel times than those which traveled along the equator. This idea was revisited by Morelli et al. [1986], who suggested that the anomalous travel times were caused by 3% cylindrical anisotropy in the inner core. This suggestion was published alongside another paper [Woodhouse et al., 1986] which also suggested that the inner core was anisotropic using evidence from free oscillations of the Earth.

Early studies postulated that the axis of anisotropy may be at an angle of 6° to the Earth’s rotation axis [Shearer and Toy, 1991; Creager, 1992]. A tilted symmetry axis tilted has also been suggested by Su and Dziewonski [1995], Song and Richards [1996], McSweeney et al. [1997], and Isse and Nakanishi [2002], with the tilt angle varying from 4° to 10.5°.

Later studies suggested instead that anisotropy is not uniform across the lateral extent of the inner core. Tanaka and Hamaguchi [1997] were the first to show that velocity...
anisotropy in the inner core is present only in some regions. Fast and slow equatorial residuals are geographically separated with a slow ‘western hemisphere’ and a fast ‘eastern hemisphere’. Observations of polar rays caused the authors to propose that anisotropy in the upper inner core only exists in the western hemisphere. This hemispherical pattern has subsequently been confirmed by other authors [Creager, 1999; Garcia and Souriau, 2000; Garcia, 2002; Niu and Wen, 2001; Wen and Niu, 2002; Oreshin and Vinnik, 2004; Yu and Wen, 2006].

[7] The hemispherical variation in the inner core is not confined to anisotropy only. Studies of the uppermost 100 km of the inner core do not find anisotropy, but instead observe that the eastern hemisphere has a larger isotropic velocity than the western hemisphere [Niu and Wen, 2001; Wen and Niu, 2002; Yu and Wen, 2006]. In the model by Sun and Song [2008], the upper parts of the inner core showed very weak anisotropy in the eastern hemisphere and stronger anisotropy in the western hemisphere below an isotropic layer of 100–200 km. The hemispherical pattern is also seen in attenuation, with Cao and Romanowicz [2004] finding that in the top 85 km of the inner core the eastern hemisphere is more attenuating than the western hemisphere.

[8] Even though many studies find the hemispherical pattern, there is not a common definition of the location of the boundaries of these two hemispheres; a wide range of boundaries have been suggested (Table 1). While the word ‘hemispherical’ is used to describe the variation, the inner core is not divided into two equal halves. Rather, the inner core is divided into two segments, by two lines of constant longitude which are not necessarily separated by 180°: suggestions of the separation of the two longitudes range from 120° to 170°. The definition of the edges of the hemispheres offered by Niu and Wen [2001] has been used by several other authors [Yu and Wen, 2006; Stroujkova and Cormier, 2004; Cao and Romanowicz, 2004]. The most extensive body wave model of inner core anisotropy, which allowed the cylindrical anisotropy to vary with longitude as well as depth, was published by Sun and Song [2008]. Unfortunately, the parametrization used prevented the authors from constraining the precise location of the boundaries between the two hemispheres.

[9] The depth to which hemispherical differences are thought to persist has also not been agreed upon. Most studies focus on the difference in anisotropy and cite a depth of at least 400 km [Souriau, 2007; Garcia and Souriau, 2000; Creager, 1999], but the difference may persist down to 600–700 km below the ICB [Creager, 1999]. In the model by Sun and Song [2008], the eastern and western hemispheres displayed different seismic properties up to 600–700 km below the ICB. This study also suggested a difference in isotropic velocities between the eastern and western hemisphere, but the differences found were too small to draw any firm conclusions.

[10] The other problem is that previous studies have often combined earlier data sets together with new data to produce updated models of inner core structure; for example the model published by Sun and Song [2008] used data from Poupinet et al. [1993], Song and Helmberger [1993], Vinnik et al. [1994], Song and Helmberger [1995], Song [1996], Tanaka and Hamaguchi [1997], Creager [1999], Niu and Wen [2001], and Sun and Song [2002]. Some studies use cross correlation to determine the differential traveltimes [e.g., McSweeney et al., 1997; Creager, 1999], while others hand pick the data [e.g., Heffrich et al., 2002] or picked at the maximum amplitudes of each phase [e.g., Yu and Wen, 2006] leading to inconsistencies when combining data sets. We have gathered a new data set which is independent of those used in previous studies, and where all of the measurements have been made using waveform cross correlation. By making an independent data set which is internally consistent, we are able to both verify earlier findings as well as more precisely determine the boundary locations and the depth extent of the hemispherical pattern.

[11] We explore the properties of the inner core using this data set with regards to the average anisotropic structure in the inner core and the possibility of a tilt to the inner core axis. Instead of using visual inspection, we find the locations of the boundaries between the two hemispheres using variance minimization and investigate the differences between the eastern and western hemispheres in the inner core, and the depths to which the hemispherical differences extend. We compare the results of our work to previous studies which have used other data sets and discuss the implications of our findings.

2. Data and Methods

[12] We use compressional PKP body waves to study the anisotropic structure of the inner core. PKP waves which turn in the middle portion of the outer core are termed PKPab, those sampling the lower portion of the outer core are termed PKPdf. PKPdf fit the inner core.

Table 1. Definitions of the Locations of the Edges of the Two Inner Core Hemispheres

<table>
<thead>
<tr>
<th>Reference</th>
<th>Longitude First Boundary</th>
<th>Longitude Second Boundary</th>
<th>Technique Used</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tanaka and Hamaguchi [1997]</td>
<td>43°E</td>
<td>177°E</td>
<td>Fitting spherical harmonics to PKPbc–PKPdf equatorial data.</td>
</tr>
<tr>
<td>Garcia and Souriau [2000]</td>
<td>40°E</td>
<td>160°E</td>
<td>Analysis of the variance reduction for PKPdf and PKPbc–PKPdf times when an isotropic layer of varying thickness is imposed at the top of the inner core.</td>
</tr>
<tr>
<td>Oreshin and Vinnik [2004]</td>
<td>50°E</td>
<td>120°W</td>
<td>PKPdf/PKb and PKPdf/PKb spectral ratios.</td>
</tr>
<tr>
<td>This study</td>
<td>14°E</td>
<td>151°W</td>
<td>Variance minimization for polar and equatorial PKPbc–PKPdf and PKPab–PKPdf data.</td>
</tr>
</tbody>
</table>
PKPbc and those which travel through the Earth for PKPdf phases sensitive to the outer and inner core. PKPbc and PKPdf are termed PKPdf or PKIKP (Figure 1). As the rays have similar paths through the crust and upper mantle, diverging only in the core, the outer core phases are often used as reference phases. Differential traveltimes, and not absolute traveltimes, between PKPdf and either PKPbc or PKPdf are then measured to minimize the effect of earthquake mislocation and velocity heterogeneity in the crust and mantle. The difference between the arrival times is assumed to be caused by structure in the core. As the outer core is relatively homogeneous [Souria et al., 2003; Ishii and Dziewonski, 2005], anomalous differential traveltimes can be ascribed to the inner core.

[13] A new, high-quality data set has been collected from earthquakes recorded between 1996 and 2005. We use events with body wave magnitude, $m_{th}$, greater than 5.5 and recorded at stations with an epicentral distance between 146.5° and 180°. The locations and magnitudes of the events are taken from the PDE results in the global Centroid Moment Tensor catalog (CMT, www.globalcmt.org). Only events deeper than 15 km are used to avoid the complications of the PKPdf phase or other crustal multiples arriving in the time window of the main PKPdf phase. Station responses were removed from the seismograms, and a two-pole Butterworth filter with corners at 0.5Hz and 2Hz was applied. Visual inspection of the seismograms is carried out to ensure that only data with low noise levels are used. From over 28,000 seismograms collected, 620 were of sufficiently high quality to measure differential traveltimes for the PKPbc–PKPdf pair and 735 to for the PKPab–PKPdf pair. The high rejection rate of seismograms reflects the need for high-quality data to study the inner core in this manner. The PKPbc–PKPdf differential traveltimes are measured for epicentral distances between 146.5° and 155.5° and the PKPab–PKPdf differential traveltimes between 150° and 177°. The PKPbc–PKPdf data are sensitive only to the uppermost 350 km of the inner core, while the PKPab–PKPdf data also sample deeper regions of the inner core (up to 1100 km below the ICB) as PKPab is observed at greater epicentral distances than PKPbc. The distribution of both data sets is nonuniform due to the limited source and receiver locations.

[14] To find the differential traveltimes, the PKPdf and the PKPbc waveforms are cross correlated. A similar procedure is carried out to measure the PKPab–PKPdf differential traveltimes, using a Hilbert transform on the PKPab waveform before cross correlation. The Hilbert transform is required because the PKPab phase is a maximum travelt ime phase which has traveled through a caustic. The differential travelt ime residual, $\delta t$, is found for each seismogram:

$$\delta t = (t_{PKPbc} - t_{PKPdf})_{data} - (t_{PKPbc} - t_{PKPdf})_{model} \tag{1}$$

A PKPdf wave traveling faster than predicted will result in a positive differential travelt ime residual $\delta t$. Model predictions are provided for IASP91 [Kennett and Engdahl, 1991] using the 'times' software [Kennett et al., 1995; Kennett and Gudmundsson, 1996], and are corrected for the ellipticity of the Earth [Dziewonski and Gilbert, 1976].

[15] The direction of raypaths in the inner core is characterized by the parameter $\zeta$, which corresponds to the angle between the PKPdf raypath in the inner core and the Earth's rotation axis. As the raypath is symmetrical in the inner core, this angle can be approximated by

$$\cos(\zeta) = \frac{\cos(\theta_i) - \cos(\theta_o)}{\sqrt{2 - 2 \cos(\theta_i) \cos(\theta_o) - 2 \sin(\theta_i) \sin(\theta_o) \cos(\phi_i - \phi_o)}} \tag{2}$$

where $\theta_i$ and $\theta_o$ are the colatitudes of the ray as it enters and exits the inner core and $\phi_i$ and $\phi_o$ are the longitudes of the ray as it enters and exits the inner core. We define $\zeta$ using the entrance and exit points of the ray in the inner core, because earthquakes which occur at different depths do not have symmetrical raypaths. This prevents the possibility of earthquake depth affecting the value we calculate for $\zeta$. Polar paths are defined as those with $\zeta < 35°$ and equatorial paths are those with $\zeta > 35°$.

3. Results

3.1. Uniform Anisotropy

[16] We first investigate uniform inner core anisotropy as seen in early body wave studies [Morelli et al., 1986; Shearer and Toy, 1991]. Figure 2 shows two example seismograms, one for an equatorial path and one for a polar path. The equatorial path (Figure 2a) has a differential travelt ime residual of $-1.1$ s, which means that PKPdf arrives later than predicted, indicating that this ray has encountered a lower
velocity in this direction than predicted by IASP91. The polar path (Figure 2b) has a differential traveltime residual of 1.1 s; the PKPdf arrival is earlier than predicted, indicating a larger velocity in the polar direction.

The observed changes in differential traveltime residual with epicentral distance are shown in Figure 3. In the PKPbc–PKPdf data set (Figure 3a) it can be seen that there are polar paths with large positive (greater than 1.7 s) differential traveltime residuals, corresponding to high velocity through the inner core. In agreement with previous studies, we interpret the anomalous polar paths as being affected by inner core anisotropy. The anomalous polar paths are distinct from the bulk of the data, which is made up of the remaining polar paths and all of the equatorial paths, with differential traveltime residuals between 1.3 s and −3.7 s.

A slight trend of increasing differential traveltime residual with increasing epicentral distance can be seen in the anomalous polar paths. This trend is due to the anisotropy in the inner core; as rays travel over a greater epicentral distance in the inner core they sample the inner core for a greater time and are more strongly affected by inner core anisotropy. The bulk of the data which do not display anomalously high traveltimes instead show a trend of decreasing differential traveltime with increasing epicentral distance. The velocity structure represented by the 1D earth model IASP91 does not completely represent either the equatorial or polar data; this is because the model is trying to average out the effects of anisotropic structure on rays traveling through the inner core.

Figure 3b shows that the distinction between polar rays and equatorial rays is much less obvious in the PKPab–PKPdf data set. The differential traveltime residuals form a much broader cluster around zero, with the polar paths again providing outliers with large positive residuals. The scatter of the data points increases as the epicentral distance increases; this is to be expected as the PKPdf and PKPab raypaths diverge with increasing epicentral distance.

When the fractional differential traveltime residuals are compared with $\zeta$ (Figure 4), we see a trend of increasing differential traveltime residual with decreasing $\zeta$. This trend confirms that the inner core is anisotropic: waves travel faster through the inner core when parallel to the rotation axis (small $\zeta$) than when perpendicular to it (large $\zeta$). The fractional traveltime (or velocity) anomalies created by a weakly anisotropic inner core, which is uniform and not depth dependent, can be modeled by a quartic function in $\cos \zeta$:

$$\frac{\delta t}{t} = \frac{\delta v}{v} = a + b \cos^2 \zeta + c \cos^4 \zeta$$

where $a$ is the difference in equatorial velocity between the observed inner core and the spherically symmetric model; $b$ and $c$ can be described in terms of the Love coefficients [Love, 1927] of the anisotropic material [Creager, 1999]. $\delta t/t$ is the fractional differential traveltime residual: the differential traveltime residual divided by the total time a PKPdf ray spends in the inner core. This is equal to the fractional velocity anomaly $\delta v/v$ when the anisotropy is weak. The error
in the percentage of anisotropy is found using bootstrap resampling [Efron and Tibshirani, 1991].

We fit curves of the form shown in equation (3) to the fractional differential traveltime residuals \(\frac{d}{t}\) of the PKPbc-PKPdf and the PKPab-PKPdf data sets and find that both exhibit anisotropy in the inner core (Table 2). Our new data supports the claims first made by Morelli et al. [1986] that there is \(P\) wave anisotropy in the inner core. There is a small difference in the anisotropy percentage found using the two data sets; slightly stronger anisotropy is found in the PKPbc-PKPdf data. Combining both types of differential traveltime residuals, the anisotropy in the inner core is 3.5% ± 0.1%. This value is the same as that found by Creager [1992], although the data sets used are independent.

The Voigt average isotropic inner core velocity (relative to the IASP91 velocity model used) is found by averaging the anisotropic velocities over all spherical angles [Creager, 1999], so that

\[
\delta v_{iso} = a + \frac{b}{3} + \frac{c}{5}
\]

For the PKPbc-PKPdf data set, which samples deeper regions of inner core, the observed isotropic velocity is 0.26% ± 0.03% greater. IASP91 describes the deeper parts of the inner core less well than the shallow parts, in particular the deeper inner core requires slightly higher velocities. The Voigt velocity of the inner core when the two data sets are combined is 0.17% ± 0.02% greater than that in the model IASP91, justifying our use of this model. The velocity difference between IASP91 and PREM [Dziewonski and Anderson, 1981] at the inner core boundary (ICB) is 0.57%. Our observed isotropic velocity variations are therefore smaller than the differences between existing velocity models of the inner core.

3.2. Longitudinal Variation of Traveltime Residuals

Uniform anisotropy does not explain the differential traveltime residuals completely, as they also vary as a function of the turning longitude of the PKPdf ray (Figure 5). The equatorial paths display a small sinusoidal trend, with smaller differential traveltime residuals between 180°W and 0°E and slightly higher differential traveltime residuals between 0°E and 180°W, similar to those seen by Creager [1999] and Garcia et al. [2006]. The polar paths show large positive traveltime residuals for some turning longitudes, and smaller

![Figure 3.](image-url)
traveltime residuals for other turning longitudes. For the PKPbc-PKPdf data set (Figure 5a), the polar paths between 180°W and 0°E (which are concentrated between 90°W and 60°W) show large positive traveltime residuals, while those polar paths with turning longitudes between 0°E and 180°W show small traveltime residuals. The same pattern can be seen in the PKPab-PKPdf data set (Figure 5b), although the broader range of turning longitudes for polar paths present in this data set shows that the region with large positive traveltime residuals extends to roughly 30°E.

The differential traveltime residuals are plotted on a map in Figure 6, which again shows that the distribution and values are not uniform. The rays which follow equatorial paths in general have small positive or negative traveltime residuals (blue triangles). There is a greater scatter in the PKPab-PKPdf differential traveltime residuals (Figure 6b) than in the PKPbc-PKPdf data (Figure 6a), because there is a greater separation between the PKPdf and PKPab raypaths than between the PKPdf and PKPbc raypaths. In particular, there are more equatorial PKPab-PKPdf data with small, positive differential traveltime residuals. For example those PKPab-PKPdf equatorial paths which turn under the central Pacific Ocean have slightly higher differential traveltimes than the PKPbc-PKPdf path in the same region.

The longitudinal pattern becomes more apparent when only the polar paths are shown (Figure 7). Paths under the Americas and the western part of the Pacific have large, positive differential traveltime residuals (red triangles); paths under the Indian Ocean, Australia and the eastern Pacific do not exhibit large positive differential traveltime residuals. Instead they have very small differential traveltime residuals, as would be expected if the inner core velocity was isotropic. The regional variation in polar path traveltime residuals suggest that anisotropy is not uniformly distributed through the inner core. The small differential traveltime residuals under Australia agree with the findings of Tanaka and Hamaguchi [1997], but are inconsistent with those of Isse and Nakanishi [2002], who reported positive differential traveltime residuals under Australia.

3.3. Tilted Symmetry Axis

Initially, the longitudinal variation in the polar paths was interpreted as evidence that the axis of cylindrical anisotropy is not aligned with the axis of Earth’s rotation [Shearer and Toy, 1991; Su and Dziewonski, 1995; Song and Richards, 1996; McSweeney et al., 1997; Isse and Nakanishi, 2002]. The motivations for suggesting a tilted anisotropy axis are twofold. First, a tilt in the axis of anisotropy would mean that equation (2) does not provide the angle $\zeta$ between a ray and the axis of anisotropy. This would mean that those polar paths which do not show anomalous fast traveltimes are not ‘polar’, and a model of uniform inner core anisotropy with a tilted axis might provide a better fit. Secondly, a tilted axis is often used in body wave studies of differential rotation of the inner core with respect to the outer core and mantle [e.g., Song and Richards, 1996; Su et al., 1996; Isse and Nakanishi, 2002]. If the axis of anisotropy is tilted and the inner core is rotating then, over time, the effective value of $\zeta$ for a particular raypath would vary so that the differential traveltime measurements would also vary over several years. We therefore investigate whether our data support the hypothesis that the inner core axis of anisotropy is not coincident with Earth’s axis of rotation.

<table>
<thead>
<tr>
<th>Data Set</th>
<th>a</th>
<th>b</th>
<th>c</th>
<th>Anisotropy (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PKPbc-PKPdf</td>
<td>−0.0057</td>
<td>−0.0125</td>
<td>0.0508</td>
<td>3.8% ± 0.1%</td>
</tr>
<tr>
<td>PKPab-PKPdf</td>
<td>−0.0013</td>
<td>−0.0196</td>
<td>0.0523</td>
<td>3.3% ± 0.2%</td>
</tr>
<tr>
<td>Both data sets</td>
<td>−0.0028</td>
<td>−0.0185</td>
<td>0.0537</td>
<td>3.5% ± 0.1%</td>
</tr>
</tbody>
</table>
Figure 5. Differential traveltime residuals for (a) $PKP_{bc}-PKP_{df}$ and (b) $PKP_{ab}-PKP_{df}$ measurements with changing turning longitude for polar paths ($\zeta < 35^\circ$) in blue and intermediate and equatorial paths ($\zeta > 35^\circ$) in yellow.

Figure 6. Differential traveltime residuals for (a) $PKP_{bc}-PKP_{df}$ and (b) $PKP_{ab}-PKP_{df}$ measurements. A positive differential traveltime residual (red) means that the $PKP_{df}$ phase has traveled faster than expected through the inner core. The triangle location corresponds to the turning location of the $PKP_{df}$ ray in the inner core; the black line is the $PKP_{df}$ raypath through the inner core.
The variance between the residuals of the combined data sets and a uniform inner core anisotropy model with a tilted axis is calculated for a range of anisotropy axis locations. We define the variance as

\[ \text{variance} = \frac{1}{N} \sum_{\text{all data}} \left( \frac{\delta t}{t} - \frac{\delta t_{\text{obs}}}{t} \right)^2 \]

where \( N \) is the total number of data points present and \( \xi \) is the angle between the raypath and the tilted axis of anisotropy. The sum of the residuals for all the data points is used as a measure of how well a tilt of the axis of anisotropy fits the data: small residuals indicate a good fit, and large residuals a poorer fit. The best fit of the data to a model of cylindrical anisotropy (i.e., the variance minimum) is gained when the axis is at 88°N, 225°E (Figure 8). The variance for this model is \( 0.335 \times 10^{-4} \). A secondary minimum, around 15°N, 305°E corresponds the anisotropy axis being the slow direction, and the fast paths under the Americas lying on the fast ‘equatorial’ plain. This secondary minimum is caused by the uneven polar path distribution and does not correspond to a realistic anisotropy model.

To calculate the uncertainty in the axis location, bootstrapping of the variance was carried out. The axis locations which have a variance within one sigma of the minimum are shown by the solid line in Figure 8. The axis locations suggested by Su and Dziewonski [1995] and Song and Richards [1996] fall within the error boundary, while the locations suggested by Shearer and Toy [1991] and McSweeney et al. [1997] do not agree with our findings. Earth’s rotation axis is located well within the error boundary and has a variance of \( 0.336 \times 10^{-4} \) which is not significantly different from the tilted axis variance. Thus, the data support either a very small deviation of the anisotropy axis or alignment of the axis of anisotropy with Earth’s rotation axis. Souriau et al.

Figure 7. Differential traveltime residuals for (a) polar PKPbc–PKPdf and (b) polar PKPab–PKPdf measurements. A positive differential traveltime residual (red) means that the PKPdf phase has traveled faster than expected through the inner core. The black line indicates the path of the PKPdf phase through the inner core. The triangle location corresponds to the turning location of the PKPdf ray in the inner core.

Figure 8. Variance of fractional differential traveltime residuals as a function of anisotropy axis location. Dark colors correspond to large variances (and show a poor fit between the data and an anisotropy axis), and light colors correspond to small variances (and therefore a good fit). The northern hemisphere above 15°N is shown.

[27]
Figure 9

(a) Variance of combined datasets

(b) Cross-sections through minima

One boundary at 14E

One boundary at 151W
[1997] also cast doubt on the analysis of inner core rotation using the tilted axis technique; we support their finding. The absence of evidence for a tilt of the axis of inner core anisotropy found in our study suggests that a different mechanism for time varying data reported by inner core super-rotation studies must be considered.

3.4. Hemispheres Boundaries

[29] As a tilt in the axis of inner core anisotropy has been ruled out to explain the longitudinal variation in the polar data, the possibility of lateral variation in the inner core must be entertained. By visually inspecting the polar data (the method used by Niu and Wen [2001]), lines of longitude separating the two hemispheres can be drawn. The edges of the two hemispheres appear to be at 175°W and 55°E, very similar to values found in previous studies (Table 1). This technique makes use of only the turning points of the polar rays, and discards any information which could be revealed by equatorial paths; moreover it relies upon a subjective analysis of the data.

[30] In order to methodically find the edges of the two hemispheres we carry out variance minimization using a two hemisphere model, where the hemispheres may be of unequal widths. To extract the imprint of hemispherical structure from all of the data, we use the equatorial paths in addition to the polar paths. The longitude of each of the boundaries between the two hemispheres is allowed to vary at 1° intervals. For each pair of boundary locations, we then calculate the best fitting anisotropy curve (equation (3)) and the corresponding variance of the data (as in equation (5)). Figure 9a shows the residuals for the combined data sets as a function of the longitudes of the boundaries between the two hemispheres. The best fitting locations of the hemisphere boundaries are at 14°E and 151°W, with 4.4% anisotropy in the western hemisphere, and 1.0% in the eastern hemisphere. The variance for this model is 0.251 × 10⁻⁴, which is significantly smaller than the variance for the uniform anisotropy models with or without tilted symmetry axis. The apparent secondary minimum in the misfit corresponds simply to switching the labels of the eastern and western boundaries; it is not a different set of possible boundary locations.

[31] Cross sections along the best fitting hemisphere boundaries are shown in Figure 9b. These are cross sections of the variance plot shown in Figure 9a which summarize what happens if one boundary is kept fixed and the other boundary is moved. The upper cross section shows the variance when one boundary is kept fixed at 14°E while the lower cross section shows the variance when one boundary is fixed at 151°W. To provide a confidence interval for the edges of the two hemispheres, bootstrap resampling was carried out. The boundaries which have variances within one sigma of the minimum variance are shaded in the profiles in Figure 9b.

[32] The confidence interval of the variance minimization technique for the eastern boundary is 18°W to 50°E, which does not include the longitude of this boundary suggested by Garcia [2002]; and the value suggested by Oreshin and Vinnik [2004] is at the limit of this boundary. The confidence interval of the western boundary is 161°E-89°W, which is wider than the eastern boundary, most likely due to the relative lack of polar paths turning under the eastern Pacific Ocean. In previous studies the location of this boundary has been suggested to be between 160°E [Creager, 1999; Garcia and Souriau, 2000] and 120°W [Oreshin and Vinnik, 2004]. The shape of the variance profile when the location of this second boundary is allowed to move (Figure 9b, top) shows that although the confidence interval extends to 89°W, the boundary is not likely to be between 120°W and 89°W. Even though none of the previous suggestions of the location of this boundary shown in Table 1 can be discounted, we find the best fitting location of this boundary using the variance minimization technique (151°W) is around 30° to the east of the values suggested by Tanaka and Hamaguchi, [1997], Garcia [2002], and Niu and Wen [2001], but 30° to the west of the longitude suggested by Oreshin and Vinnik [2004].

[33] When PKPbc-PKPdf and PKPab-PKPdf data sets are independently used to find the hemisphere locations, the PKPbc-PKPdf data is best fit by 2°W and 166°W, even though there are relatively few polar data near these longitudes to provide firm constraints on the location of the boundaries. The PKPab-PKPdf data is best fit by boundaries at 14°E and 151°W. As the global distribution of the PKPab-PKPdf data is better than the PKPbc-PKPdf data, it is not surprising that the boundaries preferred by the PKPbc-PKPdf data are the same as the boundaries preferred by the whole data set.

3.5. Hemispherical Isotropy and Anisotropy

[34] Once the locations of the best fitting boundaries between the eastern and western hemispheres have been found, we investigate the properties of the two inner core hemispheres. We divide both data sets into two subsets: rays which turn in the western hemisphere and rays which turn in the eastern hemisphere. The resulting best fitting boundaries for each individual data set, as found in section 3.4 are used. The best fitting anisotropy curves (equation (3)) are shown in Figure 10 and the best fitting parameters are shown in Table 3. For the PKPbc-PKPdf data set, the western hemisphere shows anisotropy of 4.8%, while the eastern hemisphere is only 0.5% anisotropic (Figure 10a). There are very similar numbers of data points in both hemispheres (308 data points in the western hemisphere and 312 in the eastern hemisphere) so that there is no bias due the number of data points affecting the measurement of anisotropy in these two hemispheres. The PKPab-PKPdf data set produces very similar results, with 4.5% anisotropy in the western hemisphere, and 1.4% anisotropy in the eastern hemisphere. As with the PKPbc-PKPdf results, there are similar numbers of data points in the two hemispheres.

[35] We determine the depth extent of the hemispherical structure, by separating the differential traveltime residual data into different epicentral distance ranges. The variations

Figure 9. (a) Variance when fitting both the PKPbc-PKPdf and the PKPab-PKPdf measurements to a two-hemisphere model as a function of boundary locations. The minimum is marked by the circle and is for boundaries at 14°E and 151°W. (b) Cross sections of the residuals along each of the best fitting boundaries. The dotted line shows the preferred boundary locations, green indicates the turning longitudes with a variance within 1 σ variance of the minimum.
Figure 10
of the fractional differential traveltime residual with $\zeta$ for $PKdf$ rays which reach different depths in the inner core are shown in Figures 10c and 10d. There are insufficient $PKPbc$-$PKPdf$ data to describe the anisotropy experienced by paths in the inner core which turn above 180 km below the ICB in the western hemisphere, and above 208 km in the eastern hemisphere. For deeper rays, there is consistently a small amount of anisotropy in the eastern hemisphere (0.7–1.0%) and a larger amount in the western hemisphere (4.4–5.0%). There is no clear indication in $PKPbc$-$PKPdf$ that the anisotropy increases (or decreases) with depth in either hemisphere. At those depth ranges where there are sufficient (polar) $PKPab$-$PKPdf$ paths to measure the anisotropy in both hemispheres of the inner core (220–340 km and 490–660 km below the ICB), we again find the same pattern of a more anisotropic hemisphere in the west and a less anisotropic hemisphere in the east. Thus, hemispherical differences persist to at least 660 km below the ICB.

[36] Considering the $PKPab$-$PKPdf$ data corresponding to the eastern hemisphere, the anisotropy seems to increase with depth. For epicentral distances of less than 155.5°, corresponding to rays which turn in the uppermost 350 km of the inner core, both $PKPbc$-$PKPdf$ and $PKPab$-$PKPdf$ data show that there is around 1% anisotropy in the eastern hemisphere. However, the $PKPab$-$PKPdf$ data corresponding to epicentral distances of 165–170° (660–840 km depth) are fitted by a model which contains 3.1% anisotropy throughout the eastern hemisphere of the inner core. As we find that the eastern hemisphere above 660 km depth displays only 1.2% anisotropy, to account for the increased traveltime anomaly observed by these deep polar rays it is likely that anisotropy in the inner core between 660 km and 840 km below the ICB is much higher: of the order of 6%. This measurement is based on only a few data points, but it appears that below 660 km the eastern hemisphere displays the same strong anisotropy as the western hemisphere. The western hemisphere consistently shows anisotropy of 4.5–5.0% whichever data set and depth interval are considered. There is insufficient data to measure anisotropy in the very deepest reaches of the inner core (840–1220 km below the ICB), to which only the $PKPab$-$PKPdf$ data are sensitive (Figure 10d, bottom).

[37] The Voigt average isotropic velocities (equation (4)) of the two hemispheres are shown in Table 4. The eastern hemisphere is has a lower velocity than the western hemisphere, but the differences are much smaller than the difference in anisotropic velocity between the two hemispheres.

### Table 3. Coefficients of Equation (3) for the $PKPbc$-$PKPdf$ and $PKPab$-$PKPdf$ Data Sets

<table>
<thead>
<tr>
<th>Data Set</th>
<th>a</th>
<th>b</th>
<th>c</th>
<th>Anisotropy (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$PKPbc$-$PKPdf$ - East</td>
<td>-0.0056</td>
<td>0.0144</td>
<td>-0.0090</td>
<td>0.5% ± 0.2%</td>
</tr>
<tr>
<td>$PKPbc$-$PKPdf$ - West</td>
<td>-0.0071</td>
<td>-0.0196</td>
<td>0.0676</td>
<td>4.8% ± 0.2%</td>
</tr>
<tr>
<td>$PKPab$-$PKPdf$ - East</td>
<td>0.0002</td>
<td>-0.0145</td>
<td>0.0282</td>
<td>1.4% ± 0.3%</td>
</tr>
<tr>
<td>$PKPab$-$PKPdf$ - West</td>
<td>-0.0048</td>
<td>-0.0075</td>
<td>0.0527</td>
<td>4.5% ± 0.4%</td>
</tr>
</tbody>
</table>

### Table 4. Isotropic Velocity Perturbations in the Eastern and Western Hemispheres Calculated Using Equation (4) Relative to IASP91

<table>
<thead>
<tr>
<th>Data Set</th>
<th>Eastern Hemisphere</th>
<th>Western Hemisphere</th>
</tr>
</thead>
<tbody>
<tr>
<td>$PKPbc$-$PKPdf$</td>
<td>0.01% ± 0.4%</td>
<td>0.01% ± 0.4%</td>
</tr>
<tr>
<td>$PKPab$-$PKPdf$</td>
<td>0.10% ± 0.5%</td>
<td>0.32% ± 0.5%</td>
</tr>
<tr>
<td>Both data sets combined</td>
<td>-0.03% ± 0.3%</td>
<td>0.17% ± 0.3%</td>
</tr>
</tbody>
</table>

Previous studies found similar differences in isotropic velocity, but were not able to draw firm conclusions as their error boundaries were too large [Creager, 1999; Sun and Song, 2008]. Our new data set is internally consistent, reducing the scatter in the data, and resulting in bootstrap error boundaries which are smaller than the difference between the isotropic velocities in the two hemispheres. Thus, we conclude that there is a robust small difference between the average velocities of the eastern and western hemispheres. Differences in isotropic velocity between the two hemispheres has been seen before in the top 100–200 km of the inner core [Niu and Wen, 2001; Wen and Niu, 2002; Yu and Wen, 2006], but we provide evidence that these differences persist to at least 660 km depth.

### 4. Possible Effect of CMB Structure

[38] The technique of using differential traveltime residuals should minimize the problem of mantle structure creating signals which could be interpreted as inner core anisotropy. However, Tkaleč et al. [2002] and Tkaleč [2010] suggest that much of the signal that is usually interpreted as inner core anisotropy might be due to core–mantle boundary (CMB) and mantle structures. The possible effects of CMB structure are most likely for the $PKPbc$-$PKPdf$ data set, as the $PKPbc$ and $PKPab$ rays pierce the CMB in different locations. Figure 11a shows the polar differential traveltime residuals together with the points where they pierce the CMB. Many of the polar paths with positive $PKPbc$-$PKPdf$ differential traveltime residuals travel through the CMB below the southern Atlantic Ocean, around the South Sandwich islands. The residuals for polar rays in this region vary between 0.2s and 5.3s; so that this area of the CMB does not appear to be consistently biasing the differential traveltime residuals. A number of the polar paths leave the core and travel upward through the CMB under the Aleutian Islands (around 50°N and between 172°E and 163°W). These paths also exhibit both large and small differential traveltime residuals, again indicating that this area of the CMB is not responsible for variation in traveltime residuals.

[39] Figure 11b shows a map of P wave velocity anomalies (relative to PREM) at the CMB from model P20RTS [Ritsema and van Heijst, 2002], which are of the order of 1%. To accrue a differential traveltime residual of around 4 s (as is the case for many of the polar paths), the $PKPdf$ phase would have to travel through a high–velocity anomaly (or the $PKPbc$ or $PKPab$ phase through a low–velocity anomaly) of this
magnitude for around 4800 km. As the raypaths of the PKPdf, PKPbc and PKPab phases are relatively close together in the mantle, which has a thickness of around 2850 km, it is not feasible that this type of P wave velocity structure in the mantle could be responsible for the PKPbc-PKPdf or PKPab-PKPdf differential traveltime residuals.

Our new data set supports the presence of anisotropy in the inner core, in agreement with studies of the inner core using free oscillations of the Earth [Woodhouse et al., 1986; Tromp, 1993; Durek and Romanowicz, 1999; Beghein and Trampert, 2003; Deuss et al., 2010] and previous body wave studies [Morelli et al., 1986; Shearer and Toy, 1991; Creager, 1992; Sun and Song, 2008]. A model of uniform cylindrical anisotropy in the inner core results in 3.5% ± 0.1% anisotropy in the inner core; that is, waves which travel parallel to the Earth’s rotation axis have a velocity 3.5% faster than those traveling parallel to the equatorial plane. These values are similar to those found by other authors, i.e., 3% by Morelli et al. [1986], 3.5% by Creager [1992], and 3% by Song and Helmberger [1993]. The scatter of the PKPab-PKPdf data is greater than that of the PKPbc-PKPdf data, this difference is caused by the greater separation between the PKPab and PKPdf paths in the heterogeneous lowermost mantle.

Although several authors [Shearer and Toy, 1991; Su and Dziewonski, 1995; McSweeny et al., 1997; Isse and Nakanishi, 2002] have suggested that the anisotropy axis is tilted with respect to the Earth’s rotation axis, we find that tilting the axis of inner core anisotropy relative to Earth’s rotation axis does not significantly improve the fit of the data to the anisotropy model. Without any other physical motivation to support the presence of a tilted anisotropy axis, there is no compelling evidence that the axis of anisotropy in the inner core is not aligned with the Earth’s rotation axis. If the inner core anisotropy axis is aligned with the rotation axis, then the method used by Song and Richards [1996] and others to investigate inner core rotation using the tilt of the axis of inner core anisotropy will be invalid.

We also find evidence for large-scale, regional variation in inner core anisotropy. The inner core can be divided into two hemispheres by two lines of constant longitude. The western hemisphere is anisotropic and the eastern hemisphere is nearly isotropic; this is in agreement with the findings of Tanaka and Hamaguchi [1997], Creager [1999], Garcia [2002], Wen and Niu [2002], Oreshin and Vinnik [2004], Yu and Wen [2006], and Sun and Song [2008]. Our data is
fitted best if the edges of these two hemispheres are at 14°E and 151°W. The longitude of the eastern boundary is nearly 30° west of the location used by several other authors (for example that of Niu and Wen [2001]), while the longitude of the western boundary has a wider confidence interval but falls between the wide range of values proposed by various authors.

[44] When the edges of the hemispheres are at these locations, the western hemisphere is 4.4% anisotropic, while the eastern hemisphere is nearly isotropic (exhibiting only 1.0% anisotropy). These values are close to those found by others; Creager [1999] finds 4.1% anisotropy in the western hemisphere and 0.6% in the eastern hemisphere. The similarity is encouraging as our measurements have been made using a data set independent of those carried out in previous studies. Our new data set also allow us, for the first time, to robustly determine that the eastern hemisphere has an isotropic velocity 0.2% lower than the western hemisphere. The hemispherical isotropic and anisotropic pattern in the inner core exists to at least 660 km below the inner core boundary. Below this depth there is very little data, but our data suggests that the hemispherical pattern in the inner core is no longer present and that anisotropy in the eastern hemisphere increases to approximately the same as found in the rest of the western hemisphere.

5.2. Inner Core Dynamics

[45] Hemispherical variation in the inner core anisotropy has an impact on our understanding of the processes at work in the core. Many different mechanisms have been proposed to explain the formation and continued existence of inner core anisotropy. The hemispherical structure allows us to disregard those mechanisms which cannot support longitudinal variability. For example, given the low viscosity and rapid convection of the outer core, it is unlikely that compositional heterogeneity on such a large scale has continued to exist at the inner core boundary while hundreds of kilometers of anisotropic inner core have solidified from the outer core. Understanding of the Earth’s geodynamo is developing rapidly and several mechanisms for inner core anisotropy rely on the magnetic field to create such anisotropy. It is possible that variations in heat flow at the core-mantle boundary, due to thermal and chemical processes and structures there, may affect heat flow at the inner core boundary. This variation in heat flow may be enough to allow anisotropy to form in some parts of the inner core and not in others [Sumita and Olson, 1999; Aubert et al., 2008].

[46] Figure 12a shows the inner core sensitive raypaths of our observations for both PKPbc-PKPdf and PKPab-PKPdf, together with the normal mode splitting function for the coupled mode pair $16S_{5}^{-}17S_{4}$ (Figure 12b) taken from Deuss et al. [2010]. Normal mode splitting functions represent the depth-averaged structure of the Earth, as seen by one or a pair of normal modes. Mode pair $16S_{5}^{-}17S_{4}$ is only sensitive to

Figure 12. (a) PKPbc-PKPdf and PKPab-PKPdf differential traveltime data. The thin lines indicate the paths of the PKPdf rays through the inner core, and the triangles indicate the turning point of the PKPdf rays in the inner core. (b) Cross-coupled splitting function for normal mode pair $16S_{5}^{-}17S_{4}$ [Deuss et al., 2010]. (c) Magnetic field at the core-mantle boundary [Kelly and Gubbins, 1997]. The dominant dipole field is not plotted. Lines of longitude which separate the eastern and western hemispheres (14°E and 151°W) are shown on Figures 12a–12c.
asymmetric structure in the inner core and displays a hemispherical pattern of splitting. In the western hemisphere there are red regions above North America and West Antarctica, corresponding to regions where the mode pair is split due to inner core anisotropy. The blue regions below Siberia and to the south of Australia, which correspond to the absence of splitting, because of the isotropic properties of the eastern hemisphere. The hemisphere boundaries, determined using the body wave data set described here, are shown with the splitting function and it is clear that the body wave hemispherical boundary locations agree well with the normal mode observations.

[47] The agreement between the splitting functions reported by Deuss et al. [2010] and the body wave observations we report here allows us to lay to rest a long-lived concern about interpretations of body wave data as evidence of hemispherical structure in the inner core. A significant proportion of the polar paths which provide information about the properties of the western hemisphere parallel to Earth’s rotation axis comes from South Sandwich Islands to Alaska raypaths. Some studies have suggested that only these paths, and not the whole western hemisphere, may be anomalous. However, the normal mode data confirm that there is a large-scale ‘hemispherical’ pattern in the inner core; the large number of South Sandwich Islands-Alaska raypaths fortuitously correspond to the region where normal mode data suggest that inner core anisotropy is strongest.

[48] If the proposed mechanism for inner core anisotropy is controlled by the magnetic field in the deep Earth (for example as suggested by Buffett and Wenk [2001]), then it is to be expected that there is some form of hemispherical pattern observable in the nondipole component of the magnetic field. Figure 2c shows the nondipole terms of the magnetic field at the core–mantle boundary from Kelly and Gubbins [1997]. The boundaries between the two inner core hemispheres correspond to the division between a more strongly varying and a less strongly varying magnetic field at core–mantle boundary. This correspondence of the hemispherical pattern in anisotropy with a ‘hemispherical’ signal in the magnetic field in the core adds credence to magnetic alignment mechanisms for inner core anisotropy [Karato, 1993a, 1993b; Buffett and Wenk, 2001].

[49] Two recent papers, Aloussiere et al. [2010] and Monnereau et al. [2010], have suggested a different mechanism for the hemispherical variation: inner core translation. This translation is induced by the inner core crystallizing, or growing, on the western side and melting on the eastern side, causing a translation of individual crystals in the inner core while leaving the center of mass of the inner core stationary. Aloussiere et al. [2010] support their hypothesis using laboratory experiments and modeling, and suggest that although the inner core started to form approximately one billion years ago [Labrosse et al., 2001], the oldest parts of the current inner core date from only 100 million years ago. Monnereau et al. [2010] use PKiKP–PKPdf differential traveltime and amplitude data to motivate their study and make inferences about the different grain sizes in the two hemispheres of the inner core, with larger grains corresponding to the older eastern side of the inner core.

[50] While the proposed inner core translation may explain the hemispherical differences which have been observed in the uppermost 100 km of the inner core, by both Monnereau et al. [2010] and others [e.g., Niu and Wen, 2001; Wen and Niu, 2002; Cao and Romanowicz, 2004; Yu and Wen, 2006], the mechanism can only explain the isotropic differences between the eastern and western hemispheres, and not the large differences in anisotropy which we observe here. The translational mechanisms result only in a difference in crystal size, and not alignment, which would be required to generate seismic anisotropy. Furthermore these mechanisms do not appear to be compatible with the presence of a seismically distinct ‘innermost inner core’ which has been suggested by several authors [Ishii and Dziewonski, 2002; Cormier and Stroujkova, 2005; Calvert et al., 2006; Cao and Romanowicz, 2007; Niu and Chen, 2008; Sun and Song, 2008].

5.3. Inner Core Superrotation

[51] If the inner core anisotropy shows hemispherical variation, then the superrotation of the inner core at 0.1–1.0°/year [Song and Richards, 1996; Vidale and Earle, 2005; Cao et al., 2007; Zhang et al., 2008] would mean that these hemispheres would be under opposite parts of the mantle every 1,800 years. Assuming that the anisotropic texture in the inner core is dependent on interactions between the inner core and the rest of the Earth, any mechanism which could introduce anisotropy in only one hemisphere of the inner core would have to take place over time scales which are much shorter than the rotation rate of the inner core. A growth rate of 0.15–0.8 mm/year [Buffett, 2000] would indicate that the core would only have grown 0.5 m–3 m in the time taken for the inner core to complete a full revolution with respect to the mantle. The hemispherical structure we observe in the inner core exists to at least 660 km below the inner core boundary. Estimates of the age of the inner core are between 0.4 and 1.9 Gy [Nimmo, 2007]; for the hemispherical pattern to form as the inner core was growing would require that the anisotropy mechanism has been acting in exactly the same manner for at least 200 million years and probably much longer. It is therefore difficult to conceive how this type of longitudinally varying structure could be introduced to the inner core if differential rotation was taking place. Hemispherical variation is a strong indicator that there is no steady inner core superrotation. The absence of relative rotation can be used to restrict the current geodynamical models to those which require no such rotation.

[52] Some recent models of the inner core’s motion [e.g., Song and Dai, 2008] have suggested that the inner core may be oscillating, and not steadily rotating, with respect to the mantle. The amplitude of such inner core oscillations may be small [Dumberry, 2007] or significantly larger than any steady rotation [Aubert and Dumberry, 2010]. This type of oscillation is not contraindicated by the seismological evidence. The previously observed traveltime variations [Song and Richards, 1996; Vidale and Earle, 2005; Cao et al., 2007; Zhang et al., 2008] might be part of an oscillating inner core, superimposed on a much slower steady superrotation. The inner core oscillation would blur the boundaries between the eastern and western hemisphere, explaining part of our difficulties in accurately locating the edges of the boundaries between the two inner core hemispheres and the wide confidence intervals. The idea of ‘blurred’ boundaries would also be consistent with normal mode observations of
the hemispherical structure [Deuss et al., 2010] which show wide transitional regions between the two hemispheres.

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