Constraints on aspherical core structure from PKiKP-PcP differential travel times

Keith D. Koper, Moira L. Pyle, and Jill M. Franks
Department of Earth and Atmospheric Sciences, Saint Louis University, St. Louis, Missouri, USA

Received 27 May 2002; revised 25 August 2002; accepted 6 November 2002; published 25 March 2003.

We have assembled a data set of 302 differential PKiKP-PcP travel times that were recorded by seismic array stations of the International Monitoring System. This is an order of magnitude more data than all previous PKiKP-PcP observations combined. The differential times were measured using a beam-forming and cross-correlation algorithm that has a precision of 1–2 sampling intervals (<0.1 s). The use of differential times mitigates the effects of hypocentral errors and heterogeneities in the crust and upper mantle, leaving the data sensitive to deep Earth structure. The PKiKP-PcP times are especially sensitive to variations in the thickness of the liquid outer core. The data are well fit by PREM, with residuals having a mean of −0.29 s and a standard deviation of 0.44 s, and rule out lateral variations in outer core thickness of more than 4 km. Most of the geographically coherent residuals can be explained by lower mantle heterogeneities and ellipticity corrections however there remain at least three regions that seem to require aspherical core structure. The anomalous data are best explained by long-wavelength topography on the core-mantle boundary (CMB) with peak-to-peak amplitude of 3.5 km. The CMB topography is most likely maintained by isostatic compensation of thickness variations in a distinct layer at the base of the mantle.

INDEX TERMS: 7203 Seismology: Body wave propagation; 7207 Seismology: Core and mantle; 8115 Tectonophysics: Core processes (1507);
KEYWORDS: PKiKP, inner core, CMB topography, travel time residuals, beam forming, IMS data


1. Introduction

The radial isotropic P wave structure of Earth’s core is well constrained from the travel times of refracted and reflected body waves, and the eigenfrequencies of Earth’s normal modes [Stark et al., 1986; Masters and Shearer, 1990]. Recent seismic efforts have focused on detecting and quantifying the aspherical structure of Earth’s core. Just as observations of aspherical mantle structure have led to leaps of understanding about plate tectonics, mantle convection, and mantle chemistry, we can expect observations of aspherical core structure to lead to important advances in understanding the geodynamo, the chemistry of the core, and the balance of heat transfer within the Earth. Unfortunately, seismic signals from aspherical core structure are subtle and tend to have nonunique explanations that are often controversial.

The two most frequently studied topics have been the distribution of cylindrical anisotropy in the inner core (for reviews, see Song [1997], Creager [2000], and Tromp [2001]) and topography on the core-mantle boundary at long wavelengths [e.g., Morelli and Dziewonski, 1987] and short wavelengths [e.g., Earle and Shearer, 1997]; however, there exist other less studied possibilities for core heterogeneity including topography on the inner core-out core boundary [Souriau and Souriau, 1989], zones of finite rigidity in the outer core [Rost and Revenaugh, 2001], small-scale scatterers within the inner core [Vidale and Earle, 2000], and perhaps most controversially, long-wavelength lateral wave speed variations in the liquid outer core [Souriau and Poupinet, 1990; Vasco and Johnson, 1998; Romanowicz and Breger, 2000; Boschi and Dziewonski, 2000].

Body waves give the most precision in determining deep earth structure and a common tactic in body wave studies of the core has been the use of differential travel times of various core-sensitive phases. The differing greatly mitigates the effects of hypocentral uncertainties and lateral variations in crustal and upper mantle structure. The inner core has most often been studied with the PKPAB-PKPDF and PKPBC-PKPDF combinations [e.g., Creager, 1999, and references therein] and more recently with PKPCD-PKPDF [Kaneshima et al., 1994; Niu and Wen, 2001; Ouzounis and Creager, 2001], while the outer core has been studied with SKS-SKKS differential times [Souriau and Poupinet, 1990, 1991; Lay and Young, 1990]. A less commonly observed phase pair that is strongly sensitive to core structure is PKiKP-PcP, the direct reflections from the
inner and outer core (Figure 1). Specifically, differential PKiKP-PcP travel times are most sensitive to the thickness of the liquid core as well as any lateral wave speed anomalies that may exist in the liquid core itself [Engdahl et al., 1974; Souriau and Souriau, 1989; Bina and Silver, 1997].

We have recently constructed a database of over 300 PKiKP-PcP differential travel times from short period array stations of the International Monitoring System (IMS) seismic network. This is 12 times more data than all previous PKiKP-PcP reports combined. The geographical bias of the source-receiver pairs prohibits the development of a global model, however several specific geographic regions are densely sampled. The regions span a wide range in latitude and longitude and include areas where PcP-PKiKP has previously been observed (beneath Alaska and Australia) as well as previously unsampled areas (beneath eastern Asia and central America). We analyze the variations in these PKiKP-PcP travel time residuals within and between the well sampled regions to constrain aspherical core structure.

2. Sensitivity of PKiKP-PcP Differential Times

As with other phase pair methods, the differential PKiKP-PcP travel times are insensitive to hypocentral uncertainties. This is especially true for depth uncertainties in the case of PKiKP-PcP. Within a distance range of $0^\circ - 60^\circ$ and a focal depth range of $0 - 700$ km the partial derivative of the differential PKiKP-PcP travel time with respect to focal depth varies between 0.0 and 0.01 s/km (using PREM). Considering an event with a nominal depth of 150 km and nominal distance of $30^\circ$, the partial has a value of $-0.003$ s/km and so a depth mislocation as large as 20 km will generate an apparent travel time anomaly of only 0.06 s. The sensitivity to distance is substantially larger with the partial reaching a maximum of $-0.02$ s/km, however with respect to the previous nominal location, a 10 km mislocation in range would generate an apparent travel time anomaly of only 0.15 s.

The sensitivity of PKiKP-PcP differential times to topography on the core-mantle boundary (CMB) and the inner core-outer core boundary (ICB) can be estimated using first-order perturbation theory. For instance, the relationships connecting perturbations in radius, $\delta r$, and travel time, $\delta t$, for waves transmitted and reflected at a boundary are given by Dziewonski and Gilbert [1976]. Assuming a specific velocity model it is straightforward to compute the expected differential travel time anomaly as a function of ray parameter (Figure 2). Topography on the ICB and CMB creates nearly the same magnitude of anomaly though the signs are reversed. Hence it is the thickness of the liquid core that is resolved. Figure 2 also shows that topography greater than 1–2 km on either boundary should be resolvable. Alternatively, if constraints on CMB and ICB topography can be made using theoretical arguments or previous seismological models, then the PKiKP-PcP differential times can be interpreted in terms of variations in the vertically averaged profile of $P$ wave speed in the outer core.

The main drawback to inferring core structure with PKiKP-PcP times is that as source receiver distance increases.
increases, the points where PKiKP pierces the CMB become increasingly distant from the PcP reflection point on the CMB. For example, an earthquake at 30° has PKiKP pierce points that are \(~900\) km distant from the PcP reflection point. CMB topography at wavelengths smaller than this will influence the data but not in a consistent way. Only long-wavelength CMB topography is resolvable with PKiKP-PcP times. Furthermore, the lowerrmost mantle is known to have significant small-wavelength velocity anomalies that could influence the PKiKP-PcP times. Although these waves are traveling nearly vertically just above the CMB and so would spend little time sampling D$_0$, the extreme anomalies observed in Ultralow-velocity zone (ULVZ) regions [Garnero et al., 1998] could have significant effects on PKiKP-PcP times. Few of the data presented here are likely to have been significantly affected by ULVZ's, however we will return to this point when interpreting the IMS data and quantify the potential impact.

3. Observing PKiKP-PcP Phase Pairs

[9] A search of the seismological literature revealed 25 previous PKiKP-PcP travel time reports (Table 1). This scarcity is mainly related to the difficulty in observing PKiKP at precritical distances. While observations of PKiKP at postcritical distances (\(\Delta \geq 110°\)) are fairly common, at precritical distances nearly all of the PK energy incident on the ICB is transmitted as PK and PKJ energy. In fact, it took nearly 35 years after the discovery of the inner core for the first unambiguous observations of precritical PKiKP to be made [Engdahl et al., 1970]. The difficulty in observing precritical PKiKP is perhaps best appreciated in the work of Shearer and Masters [1990], who have searched over 4900 seismograms in the distance range of 20° < \(\Delta < 90°\) and found just two unambiguous PKiKP arrivals.

[10] One factor associated with nearly all precritical PKiKP observations is the use of a small-aperture array of vertical component seismometers as a receiver. Standard signal processing techniques can be applied to such data that dramatically increase signal-to-noise ratios. Array data also provide slowness observations which help confirm the identity of prospective PKiKP arrivals. A second factor that contributes to PKiKP observability is source impulsivity. Virtually all previous PKiKP reports are associated with either explosions or nonshallow earthquakes, both of which are known to have more compact and impulsive energy release than shallow earthquakes. Nonshallow earthquakes also generate far less surface wave energy than shallow earthquakes, and so given a fixed energy release, nonshallow earthquakes preferentially excite body waves and hence core reflections.

[11] A third important factor for observing PKiKP is the radiation pattern of the source. There must be a significant amount of compressional energy directed at very small takeoff angles for PKiKP to be excited. Since nuclear explosions have strong isotropic components, they excel at producing PKiKP observations, while earthquakes must have favorable dip-slip focal mechanisms to produce PKiKP. The fact that nuclear explosions from geographically diverse test sites (Nevada, Aleutians, Semipalatinsk, Novaya Zemlya, China) have generated observations of precritical PKiKP implies that energy focusing by small-wavelength ICB topography or lower mantle velocity anomalies is not a prerequisite for observing PKiKP.

[12] A final complication in observing PKiKP-PcP phase pairs is that at the smallest distances (<10°), where theoretical precritical PKiKP amplitudes are largest, the theoretical PcP amplitudes become quite low (Figure 3). At these ranges it can be the case that PKiKP is observed and that PcP is not. For example, seismograms of nuclear explosions detonated at the Russian test site in Semipalatinsk and recorded at a distance of 6° show clear PKiKP phases but no PcP phases [Adushkin et al., 2000]. We have found a handful of these examples in the IMS data set as well.

4. Data From IMS Stations

[13] The seismic component of the International Monitoring System differs from other global networks in that a large fraction of the seismograph stations are arrays of short-period seismometers. As noted above, the existence of such arrays is a key requirement for observing precritical PKiKP. Seismic data from IMS stations are currently archived at the International Data Center (IDC) in Vienna. Although access to the most recent data is restricted to researchers with specific government affiliations, complete IMS data from the time period 1 January 1995 to 20 February 2000 are freely available from the prototype International Data Center (pIDC) in Washington, D. C.

[14] As part of its normal operations the pIDC produced an earthquake catalog known as the Reviewed Event Bulletin (REB) that includes phase arrival times and hypocentral parameters. We were able to obtain the REB for the entire open period with the exception of the months of September 1999, December 1999, and February 2000. A search of the REB revealed 571 joint reports of PcP and PKiKP at distances of 10.8°–79.6°. The vast majority of times arose from a handful of array stations located relatively close to subduction zones (Table 2). Only a single three-component station, STKA in southeastern Australia, reported a substantial number of PKiKP-PcP observations.

[15] The distributions of the REB differential travel time residuals with respect to three standard radial models, PREM [Dziewonski and Anderson, 1981], IASP91 [Kennett and Engdahl, 1991], and AK135 [Kennett et al., 1995], show similar scatter with standard deviations of 1.37 s; however, PREM gives a much smaller mean (0.16 s) than either IASP91 (−1.35 s) or AK135 (−1.24 s). Note that since these are differential residuals, the effect of the reference model used to locate the events is mitigated and the baseline difference is significant in terms of Earth structure. The practical reason for the difference is that PKiKP-PcP observations from Engdahl et al. [1974] were included in the construction of PREM, while IASP91 and AK135 were apparently constructed without PKiKP-PcP travel time constraints.

5. Validation of the PKiKP-PcP Times

[16] By differencing the PKiKP and PcP arrival times the effect of systematic clock errors at seismograph stations is eliminated. However, there are other potential sources of systematic error that may not be mitigated as well by simply
differencing the reported arrival times. For example, a study of delay times at stations that report to the International Seismological Centre (ISC) found cases where apparent anomalies are created by seasonal changes in the background seismic noise and possibly by changes in the analyst.

Table 1. Previous Differential PKiKP-PcP Travel Time Observations

<table>
<thead>
<tr>
<th>Date</th>
<th>Origin Time, UTC</th>
<th>Event Lat, °N</th>
<th>Event Long, °E</th>
<th>Event Depth, km</th>
<th>Station Lat, °N</th>
<th>Station Long, °E</th>
<th>Δt, deg</th>
<th>Δt, s</th>
<th>Ref*</th>
</tr>
</thead>
<tbody>
<tr>
<td>19 Jan. 68</td>
<td>1815:00.1</td>
<td>38.634</td>
<td>−116.215</td>
<td>0.0</td>
<td>46.689</td>
<td>−106.222</td>
<td>10.90</td>
<td>477.50</td>
<td>Q73</td>
</tr>
<tr>
<td>16 Nov. 69</td>
<td>1030:01.7</td>
<td>13.351</td>
<td>−89.650</td>
<td>79.0</td>
<td>46.689</td>
<td>−106.222</td>
<td>36.04</td>
<td>438.75</td>
<td>EFR70</td>
</tr>
<tr>
<td>20 June 69</td>
<td>0237:51.5</td>
<td>53.172</td>
<td>−162.435</td>
<td>44.0</td>
<td>45.689</td>
<td>−106.222</td>
<td>35.94</td>
<td>438.35</td>
<td>EFR70</td>
</tr>
<tr>
<td>14 Oct. 70</td>
<td>0559:57.1</td>
<td>73.315</td>
<td>55.146</td>
<td>0.0</td>
<td>82.483</td>
<td>−62.400</td>
<td>21.34</td>
<td>464.90</td>
<td>BWP73</td>
</tr>
<tr>
<td>14 Oct. 70</td>
<td>0559:57.1</td>
<td>73.315</td>
<td>55.146</td>
<td>0.0</td>
<td>74.687</td>
<td>−94.900</td>
<td>31.08</td>
<td>448.20</td>
<td>BWP73</td>
</tr>
<tr>
<td>14 Oct. 70</td>
<td>0559:57.1</td>
<td>73.315</td>
<td>55.146</td>
<td>0.0</td>
<td>63.733</td>
<td>−68.467</td>
<td>38.17</td>
<td>433.50</td>
<td>BWP73</td>
</tr>
<tr>
<td>6 Nov. 71</td>
<td>2200:00.1</td>
<td>51.472</td>
<td>179.107</td>
<td>1.7</td>
<td>46.376</td>
<td>−135.098</td>
<td>26.64</td>
<td>457.40</td>
<td>BWP73</td>
</tr>
<tr>
<td>19 Jan. 68</td>
<td>1815:00.1</td>
<td>38.634</td>
<td>−116.215</td>
<td>0.0</td>
<td>46.689</td>
<td>−106.222</td>
<td>10.90</td>
<td>477.50</td>
<td>Q73</td>
</tr>
<tr>
<td>23 Aug. 72</td>
<td>0847:16.0</td>
<td>58.250</td>
<td>−153.578</td>
<td>61.0</td>
<td>47.424</td>
<td>−106.741</td>
<td>29.69</td>
<td>451.15</td>
<td>EFM74</td>
</tr>
<tr>
<td>20 Feb. 73</td>
<td>0740:34.7</td>
<td>58.307</td>
<td>−149.810</td>
<td>12.0</td>
<td>47.424</td>
<td>−106.741</td>
<td>27.71</td>
<td>454.80</td>
<td>EFM74</td>
</tr>
<tr>
<td>6 June 73</td>
<td>1300:00.8</td>
<td>37.245</td>
<td>−116.346</td>
<td>0.0</td>
<td>45.888</td>
<td>−105.728</td>
<td>11.73</td>
<td>477.20</td>
<td>EFM74</td>
</tr>
<tr>
<td>13 Nov. 79</td>
<td>2043:38.8</td>
<td>−23.576</td>
<td>−174.858</td>
<td>32.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>47.12</td>
<td>416.10</td>
<td>SS89</td>
</tr>
<tr>
<td>19 Feb. 81</td>
<td>0823:02.2</td>
<td>−21.541</td>
<td>169.462</td>
<td>33.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>32.83</td>
<td>445.00</td>
<td>SS89</td>
</tr>
<tr>
<td>23 Aug. 81</td>
<td>0159:50.1</td>
<td>−22.069</td>
<td>170.955</td>
<td>100.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>34.19</td>
<td>442.20</td>
<td>SS89</td>
</tr>
<tr>
<td>7 June 81</td>
<td>2140:35.5</td>
<td>16.592</td>
<td>145.471</td>
<td>311.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>37.92</td>
<td>433.40</td>
<td>SS89</td>
</tr>
<tr>
<td>17 April 82</td>
<td>0920:57.8</td>
<td>19.871</td>
<td>120.526</td>
<td>10.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>41.83</td>
<td>424.30</td>
<td>SS89</td>
</tr>
<tr>
<td>3 Dec. 82</td>
<td>2229:59.7</td>
<td>−13.323</td>
<td>167.205</td>
<td>257.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>32.12</td>
<td>447.10</td>
<td>SS89</td>
</tr>
<tr>
<td>26 Oct. 84</td>
<td>0849:24.8</td>
<td>1.623</td>
<td>126.291</td>
<td>56.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>22.85</td>
<td>465.30</td>
<td>SS89</td>
</tr>
<tr>
<td>3 July 85</td>
<td>0436:51.7</td>
<td>−4.439</td>
<td>152.828</td>
<td>33.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>23.70</td>
<td>461.60</td>
<td>SS89</td>
</tr>
<tr>
<td>7 Nov. 85</td>
<td>1912:31.0</td>
<td>−35.257</td>
<td>−179.347</td>
<td>44.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>43.38</td>
<td>421.60</td>
<td>SS89</td>
</tr>
<tr>
<td>16 June 86</td>
<td>1048:25.7</td>
<td>−22.037</td>
<td>−178.925</td>
<td>547.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>43.54</td>
<td>418.10</td>
<td>SS89</td>
</tr>
<tr>
<td>14 Oct. 86</td>
<td>1653:08.1</td>
<td>5.030</td>
<td>153.616</td>
<td>41.0</td>
<td>−9.194</td>
<td>134.339</td>
<td>23.90</td>
<td>463.70</td>
<td>SS89</td>
</tr>
<tr>
<td>21 May 92</td>
<td>0459:57.6</td>
<td>41.604</td>
<td>88.813</td>
<td>0.0</td>
<td>31.024</td>
<td>91.7</td>
<td>10.80</td>
<td>476.80</td>
<td>BWP73</td>
</tr>
</tbody>
</table>

*References are Q73, Qamar [1973], EFR70, Engdahl et al. [1970]; BWP73, Buchbinder et al. [1973]; EFM74, Engdahl et al. [1974]; SS89, Souriau and Souriau [1989], and BS97, Bona and Silver [1997].

Figure 3. Reflection coefficients for PcP at the CMB and PKiKP at the ICB for a spherical Earth [Zhao and Dahlen, 1993]. A surface focus and PREM were used for the calculations. The transmission coefficients for PKiKP at the CMB are near unity. The theoretical relative amplitudes are also affected by differences in geometrical spreading, which increase PcP/PKiKP ratios by factors of 2–4, and anelastic attenuation in the outer core and lower mantle.
We inspected the zero slowness beams and defined small time windows, generally 5–8 s long, that bracketed the peaks of the \( P_{cP} \) and \( P_{KiKP} \) pulses. We used a grid search to find the horizontal slowness vector that maximized the peak amplitude in each time window and then formed optimal slowness \( P_{cP} \) and \( P_{KiKP} \) beams. Next we cross-correlated the optimal \( P_{cP} \) pulse with the optimal \( P_{KiKP} \) pulse to determine a differential time between the two phases. In practice, the observed slownesses of \( P_{cP} \) and \( P_{KiKP} \) were such that the optimal beams differed only slightly from the zero slowness beams. Therefore the differential times taken from the zero slowness beams were nearly always the same as those taken from the optimal slowness beams. The differential times from linear beams were usually within 1–2 sampling intervals of the times from the PWS beams.

As an example, we consider a 5.3 \( m_b \) earthquake that occurred in Central America (3 March 1998, 0224:47.4 UTC, 14.493\(^\circ\)N 91.384\(^\circ\)W, 67.2 km) and was recorded 32.3\(^\circ\) away at the IMS array in Pinedale, Wyoming (PDAR). This event produced an exceptional sequence of core reflections including \( P_{cP} \), \( pP_{cP} \), \( ScP \), \( sS_{cP} \), and \( P_{KiKP} \) (Figure 4). The \( P_{cP} \) and \( P_{KiKP} \) pulses are quite similar, and cross-correlation using the zero slowness PWS beam gives a differential time of 447.250 s. Cross-correlation of the two pulses using the zero slowness linear beam gives a differential time that is just one sampling interval different, 447.225 s.

Performing a grid search over horizontal slowness \( (s_x, s_y) \), we find that values of \( s_x = 3.44 \) s/deg and \( s_y = 2.50 \) s/deg maximize the peak energy in the \( P_{KiKP} \) time window. This corresponds to a backazimuth of 126\(^\circ\) and slowness magnitude of 4.24 s/deg. A similar search using the \( P_{cP} \) window gives \( s_x = 4.76 \) s/deg and \( s_y = 4.06 \) s/deg, which is equivalent to a backazimuth of 130\(^\circ\) and magnitude of 5.18 s/deg.

To analyze this data, we developed a grid of azimuths and slownesses, which we refer to as a horizontal slowness vector (HSV). The HSV is a 3D vector that represents the slowness of a particular wavefront. Each HSV has a unique backazimuth and slowness, and by searching the grid, we can find the optimal HSV for each phase. The optimal HSV for \( P_{cP} \) is 126\(^\circ\) with a slowness magnitude of 4.24 s/deg, and the optimal HSV for \( P_{KiKP} \) is 130\(^\circ\) with a slowness magnitude of 4.18 s/deg. These values are very close, suggesting that the two phases are generated by the same source or a nearby source.

### Table 2: IMS Stations Used in This Study

<table>
<thead>
<tr>
<th>Station</th>
<th>Lat, N</th>
<th>Long, E</th>
<th>Aperture, km</th>
<th>Aspect Ratio</th>
<th>Number of Elements</th>
<th>Number of Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASAR</td>
<td>-23.666</td>
<td>133.904</td>
<td>9.8</td>
<td>0.011</td>
<td>19</td>
<td>110</td>
</tr>
<tr>
<td>BRAR</td>
<td>39.725</td>
<td>33.639</td>
<td>3.7</td>
<td>0.040</td>
<td>6</td>
<td>1</td>
</tr>
<tr>
<td>CMAR</td>
<td>18.457</td>
<td>98.943</td>
<td>11.0</td>
<td>0.006</td>
<td>18</td>
<td>58</td>
</tr>
<tr>
<td>FINES</td>
<td>61.444</td>
<td>26.077</td>
<td>2.1</td>
<td>0.020</td>
<td>16</td>
<td>1</td>
</tr>
<tr>
<td>KSAR</td>
<td>37.442</td>
<td>127.884</td>
<td>10.7</td>
<td>0.028</td>
<td>19</td>
<td>22</td>
</tr>
<tr>
<td>ILAR</td>
<td>64.771</td>
<td>-146.887</td>
<td>9.6</td>
<td>0.038</td>
<td>19</td>
<td>4</td>
</tr>
<tr>
<td>NVAR</td>
<td>38.429</td>
<td>-118.303</td>
<td>9.9</td>
<td>0.069</td>
<td>11</td>
<td>2</td>
</tr>
<tr>
<td>PDAR</td>
<td>42.767</td>
<td>-109.558</td>
<td>3.4</td>
<td>0.046</td>
<td>13</td>
<td>12</td>
</tr>
<tr>
<td>STKA</td>
<td>-31.877</td>
<td>141.595</td>
<td>–</td>
<td>–</td>
<td>1</td>
<td>8</td>
</tr>
<tr>
<td>TXAR</td>
<td>29.334</td>
<td>-103.667</td>
<td>4.7</td>
<td>0.018</td>
<td>9</td>
<td>6</td>
</tr>
<tr>
<td>WRA</td>
<td>-19.943</td>
<td>134.139</td>
<td>23.9</td>
<td>0.002</td>
<td>20</td>
<td>67</td>
</tr>
<tr>
<td>YKA</td>
<td>62.493</td>
<td>-114.605</td>
<td>24.1</td>
<td>0.002</td>
<td>18</td>
<td>11</td>
</tr>
</tbody>
</table>

*We define aperture as the diameter of the circle that has the same area as the smallest rectangle that includes all the elements.

*bWe define aspect ratio as the change in elevation between the highest and lowest elements, divided by the aperture.

Figure 4. Seismograms recorded at PDAR from a 5.3 \( m_b \) earthquake located 32.3\(^\circ\) away (3 March 1998, 0224:47.4 UTC, 14.493\(^\circ\)N 91.384\(^\circ\)W, 67.2 km). We show three time windows with the zero time set as the origin time: (a) the entire waveform, (b) the \( P_{cP} \) time window, and (c) the \( P_{KiKP} \) time window. In each case the top trace is from a single array element, the middle trace is linear stack at zero slowness, and the bottom trace is a phase weighted stack at zero slowness. All the data have been high passed at 1.0 Hz.
6.25 s/deg. The results of these searches are illustrated in Figure 5. A cross correlation of these two optimal PWS beams gives a differential PKiKP-PcP travel time of 447.250 s, precisely the same value obtained with the zero slowness beam. This is not surprising since the optimal beams differ very little from the zero slowness beam. Owing to the broadness of the slowness peaks in Figure 5, there is a large range of slownesses which give identical differential PKiKP-PcP travel times.

The horizontal slownesses derived above differ substantially from theoretical expectations. For instance PREM predicts slowness magnitudes of 2.7 s/deg for PcP and 0.71 s/deg for PKiKP, and the expected backazimuth is 145°. Slowness anomalies at arrays are not uncommon and can arise because of deep earth velocity heterogeneities [Kanasewich et al., 1972; Tibuleac and Herrin, 1999], near receiver site effects [Lin and Roecker, 1996; Tibuleac and Herrin, 1997], or intra-array velocity variations, including topography [Bokelmann, 1995]. Accounting for such anomalies is critical when slownesses are used in event location, and a large amount of effort has gone into calibrating various array stations and generating corrective slowness perturbations [Koch and Kradolfer, 1997; Bondar and North, 1999; Bondar et al., 1999]. However, since the differential PKiKP-PcP times are insensitive to modest slowness variations, our data are unaffected by these slowness anomalies.

We reinforce this idea with two experiments that illustrate the robustness of the PKiKP-PcP times. First, we modify the beam-forming algorithm to take into account intraarray topography. Following the work of Bokelmann [1995], we include vertical slowness as a free parameter in the grid search for the optimal PcP and PKiKP beams. For the PcP time window of the previous example the optimal three-dimensional (3-D) slowness vector \((s_x, s_y, s_z)\) becomes \((2.89, -3.51, 18.3)\) s/deg, corresponding to a backazimuth of 140°, a horizontal slowness magnitude of 4.54 s/deg, and a mean velocity of 5.9 km/s beneath the array. For the PKiKP time window the optimal slowness vector is \((1.71, 2.81, 25.7)\), corresponding to a backazimuth of 148°, a horizontal slowness magnitude of 3.3 s/deg and a mean velocity of 4.3 km/s beneath the array. A comparison of the optimal 3-D beams, the optimal 2-D beams, and the zero slowness beams is presented in Figure 6. Cross correlation of the optimal 3-D PcP and PKiKP pulses gives the same differential time as before, 447.250 s. So although accounting for topography substantially reduces the bias in horizontal slowness estimation, it has no effect on the quality of the differential times. Note also that the differential slownesses we observe are normal and so whatever structure is causing the abnormally high slownesses is affecting PcP and PKiKP in the same way.

For the second test we determine PKiKP-PcP times on a trace by trace basis. For almost all of the events in our database, PKiKP is not recognizable on single traces and so this experiment can be done just for a very few high-quality events. This was part of the motivation for our choice of the example event. Using PcP and PKiKP time windows previously defined from inspection of the zero slowness beams we find that individual cross correlations give differential times that vary by at most ±2 sampling intervals, 447.200 s to 447.300 s. The average of these 13 values is

![Figure 5](image_url)
reasons for the large number of REB PKiKP insensitive to minor variations in the slowness used to form times determined from cross correlation of array beams are similar in all cases, indicating that differential times have very small changes (0.0–0.10 s) for variations in intraarray topography. The pulse shape is similar in all cases, indicating that differential PKiKP-PcP times determined from cross correlation of array beams are insensitive to minor variations in the slowness used to form the beams.

447.242 s, only 0.008 s different than the optimal beam derived value. Therefore any intraarray velocity anomalies at PDAR have a negligible effect on beam-derived PKiKP-PcP differential times.

Another reason for our use of this particular event as an example in the robustness tests is that PDAR is well known as an array with significant slowness anomalies and it also has a relatively large amount of intraarray topography (Table 2). If PKiKP-PcP times are robust with respect to slowness anomalies at PDAR, it can be expected that they will be robust at the other array stations as well. At larger aperture stations, such as YKA and WRA, we generally find narrower peaks in slowness space corresponding to higher slowness resolution; however, grid searches around the optimal slowness values again show that PKiKP-PcP differential times have very small changes (0.0–0.10 s) for slowness values that change by as much as 2–3 s/deg. At least part of the reason for the general robustness of the PKiKP-PcP times is the fact that the cross-correlation process tends to align the largest peaks in the PcP and PKiKP pulses, and the locations of these peaks are stable features of the beams, much more so than the locations of the first breaks for instance.

6. Results

We were successful in observing a total of 302 PKiKP-PcP differential travel times. There were several reasons for the large number of REB PKiKP-PcP reports (over 250) that could not be verified. In some cases we were either not able to access the waveforms for the appropriate time period or the data that were obtained were obviously bad. Most commonly, data were rejected because PKiKP could not be reliably observed above the noise, even in the PWS beams. In three cases, data were rejected because of an unusually large slowness for the apparent PKiKP arrival (>10 s/deg), implying phase misidentification. There were also 5–10 data in which the PcP and PKiKP arrivals had substantially different pulse shapes and cross correlation could not give a reasonable alignment. This was true for the optimal slowness beams and the zero slowness beams (plain sums). In these cases we handpicked the differential times if the data were otherwise acceptable. We estimate the observational error of these handpicked differential times to be 0.25 s. Just 8 of the 56 PKiKP-PcP reports from the three-component station STKA were found to be viable, supporting previous results on the difficulty of observing precritical PKiKP on three-component stations from earthquake sources.

The distance range at which PcP-PKiKP reports were verified was 5.86° < ∆ < 56.40°, with 75% at distances of 20°–40°. The lack of observations at distances greater than 56.40° is consistent with the low theoretical PKiKP amplitude for 60° < ∆ < 90° (Figure 3). Approximately 80% of the data came from earthquakes having a focal depth greater than 40 km, again consistent with expectations. In contrast, many of the earthquakes were surprisingly small, with 41% having mb < 5.0. This implies that it is the impulsivity and radiation pattern of the source that are paramount in generating precritical PKiKP observations and that overall source magnitude is a secondary consideration.

6.1. Global Properties of the PKiKP-PcP Data Set

The 302 raw PKiKP-PcP travel time residuals have a mean of –0.07 s, a standard deviation of 0.49 s, and a maximum amplitude of 1.85 s with respect to PREM (Figure 7). Much of the scatter seen in the original REB data set has been eliminated; however, the AK135 and IASP91 models retain significant nonzero means, –1.50 s and –1.62 s, respectively. Considering the kernels shown in Figure 2, it could be that the outer core is 7–8 km too thin in AK135 and IASP91 when compared to the real Earth. However, IASP91 and AK135 actually have thicker outer cores than PREM (by 6.5 and 3.5 km, respectively). This implies that the average velocities in the outer cores of AK135 and IASP91 need to be reduced to be reconciled with the PKiKP-PcP data. In any case, for the remainder of the paper we compute residuals using PREM as the reference model.

We investigate two possibilities for reducing the scatter in the global PKiKP-PcP differential times. First, we consider corrections for Earth’s ellipticity [Kennett and Gudmundsson, 1996]. These corrections implicitly include the effect of the variation in the radii of the CMB and ICB induced by the rotation of the Earth. Note that these variations are nontrivial as the difference between the equatorial and polar radii goes from 21.4 km at Earth’s surface to 8.9 km at the CMB to 3.0 km at the ICB (see Appendix A for details). Residuals that remain after the ellipticity corrections have been applied are then due either
to nonhydrostatic core structure or to hydrostatic core structure induced by density anomalies unrelated to Earth rotation, such as those created during mantle convection. Since most of the $PKiKP$-$PcP$ ray paths are near equatorial and we are computing differential values the theoretical travel time corrections are relatively small (Figure 8a), but they do reduce the standard deviation of the data from 0.49 s to 0.44 s, primarily by reducing the strongly negative residuals at the high-latitude stations YKA and ILAR. The overall mean shifts from $-0.07$ s to $-0.29$ s.

A second possible source of bias is the different sensitivity of $PcP$ and $PKiKP$ ray paths to 3-D mantle structure, especially in the deep mantle. We estimate corrections to $PKiKP$-$PcP$ times using two recent tomographic models of long-wavelength $P$ wave variation in the mantle [Boschi and Dziewonski, 2000; Zhao, 2001]. In both cases we generate corrections by integrating velocity anomalies along 1-D mantle ray paths and then differencing the theoretical $PKiKP$ and $PcP$ residuals. The corrections are generally small (Figures 8b and 8c), and the two models show an impressively strong correlation (Figure 8d) given that different data sets, different parameterizations, and different reference models were used in their development. Applying corrections from either of the two models to either the raw times or the times corrected for ellipticity does not appreciably change the standard deviation of the residual distribution.

The strong geographical bias in our source-receiver paths, with 68% of our data arising from the three Australia stations STKA, WRA, and ASAR, complicates the interpretation. Such biases can introduce systematic errors in the results.

Figure 7. Differential $PKiKP$-$PcP$ travel time residuals for the 302 confirmed data. The residuals are uncorrected for ellipticity and 3-D mantle structure and are computed with respect to PREM. The stars indicate earthquake epicenters, and the colored circles indicate the ICB bounce points. The surface projections of the ray paths are shown as lines connecting each source-receiver pair. The arrows indicate the four spots where robust anomalies remain after corrections for ellipticity and mantle structure are applied; however, we do not interpret area 4 in terms of aspherical core structure because of the possibility of contamination by anomalous features in the lowermost mantle.

Figure 8. Theoretical corrections to the 302 differential $PKiKP$-$PcP$ travel times for (a) ellipticity, (b) 3-D mantle heterogeneities using the model of Boschi and Dziewonski [2000] (BD00), and (c) 3-D mantle heterogeneities using the model of Zhao [2001] (Z01). Application of the corrections is discussed in the text. (d) Correlation between the corrections for the two theoretical mantle models. The correlation coefficient is 0.49.
interpretation of the effects of the travel time corrections on the global data set. For example, it could be the case that the 3-D mantle corrections make modest but significant improvements to the Alaska residuals (5% of the data) but that the global variance is not reduced because the theoretical corrections have no effect on the Australia residuals. For this reason we prefer to interpret the \( \text{PKiKP-PcP} \) residuals, with and without corrections, on a region by region basis. We examine the data from four well-sampled geographical regions which together account for 99% of our confirmed \( \text{PKiKP-PcP} \) differential travel time observations.

6.2. \( \text{PKiKP-PcP} \) Times Beneath Australia

[31] The three Australia stations are the most prolific recorders of precritical \( \text{PKiKP} \) in the world due to their position in a stable continental interior and the proximity of a large number of deep earthquakes. In particular, the unusual sensitivity of the three-component station STKA is probably related to its location in the Gawler Archean craton which presumably has a deep continental root and hence high \( Q \) in the upper mantle. Residuals recorded jointly at WRA and ASAR show a high degree of correlation (Figure 9), implying that the Earth structure signal in the data is much larger than that created by observational errors.

[32] The residuals have a mean of \(-0.26\) s and standard deviation of \(0.44\) s after ellipticity corrections and show a spatially complicated pattern (Figure 10). However, the variations tend to be coherent geographically, with neighboring residuals having similar values. The exception to this is the cluster of residuals beneath the northeastern tip of Australia. The poor coherence of these data prevents an interpretation in terms of Earth structure. A possible explanation for this complexity is that source side ray paths overlap with a well documented ULVZ region \cite{Garnero et al., 1998}. Such regions are known to vary on short scales and could induce direct travel time effects not predicted by long-wavelength mantle models, as well as inducing indirect travel time effects by distorting the \( \text{PcP} \) and/or \( \text{PKiKP} \) pulses and making the cross-correlation technique less robust.

[33] The residuals with bounce points east of 150° are matched exceptionally well by the predictions of the Zhao [2001] 3-D mantle model. When corrected by this model, the residuals have significantly reduced variance and little geographical coherence with a slightly negative mean near

![Figure 9](image_url) \( \text{PKiKP-PcP} \) differential residuals that were simultaneously recorded at WRA and ASAR from the same earthquake. The two array stations are separated by \(\sim 400\) km and have significantly different geometries. The strong positive correlation indicates that observational errors are much lower than the Earth structure signal.

![Figure 10](image_url) The differential \( \text{PKiKP-PcP} \) travel time residuals for the three Australia stations WRA, ASAR, and STKA. The stars are epicenters, the squares are receiver locations, and the colored circles are ICB bounce points. We show (a) the raw, uncorrected data and (b) the theoretical corrections for the 3-D mantle model of Zhao [2001]. Much of the spatial variation in the residuals can be accounted for by the mantle corrections. The arrow shows the location of the data that cannot be easily explained without aspherical core structure.
the global average of $-0.29\,\text{s}$. The northernmost residual cluster (centered at 140°E) can also be well accounted for by ellipticity and 3-D mantle corrections. The only residuals that remain significantly anomalous and coherent after the standard corrections are those that cluster along the 130° meridian. The mean here is about 0.3 $\text{s}$ too large compared to the global mean and would be consistent with a thickening of the outer core by 1.0–1.5 km.

6.3. PKiKP-PcP Times Beneath Southeast Asia

[34] We observed a total of 80 PKiKP-PcP differential travel times at the two array stations, KSAR and CMAR, located in southeastern Asia. The ellipticity corrected residuals have a mean of $-0.42\,\text{s}$ and a standard deviation of 0.32 $\text{s}$. The residuals clustered beneath the South China Sea show particularly complicated behavior and account for most of the scatter. The 3-D mantle corrections of Zhao [2001] (and to a lesser extent of Boschi and Dziewonski [2000]) can explain some of this variation, however the sharp differences among the oval shaped cluster immediately to the east of Vietnam, and the evidence for no ULVZs in this region [Garnero et al., 1998], indicate higher than normal observational error.

[35] The remaining residuals, north of $\sim 18^\circ$, are consistently negative and have no obvious regional trend. It is likely that most of the small variations can be explained by 3-D mantle structure and observational error. The exception to this is the patch of residuals to the southeast of Japan that are produced by Marianas earthquakes recorded at KSAR. These residuals are the most negative in the region, and the theoretical corrections for 3-D mantle structure and ellipticity cannot account for their large magnitude. Furthermore, the residuals are coherent geographically, implying low observational error. These residuals can be accommodated by 1–2 $\text{km}$ thinning of the liquid core.

6.4. PKiKP-PcP Times Beneath Southern North America

[36] The bulk of the data validated at PDAR and TXAR arise from a series of intermediate depth events beneath Central America and result in two tight geographical clusters. The raw anomalies within each cluster are coherent, but there is a significant difference between the two clusters. The 10 residuals of the northern cluster have a mean of 0.56 $\text{s}$, while the 5 residuals from the southern cluster have a mean of $-0.24\,\text{s}$. Corrections for ellipticity and both 3-D mantle models are mild (magnitudes less than 0.2 $\text{s}$) for all the data. More importantly, all of the predicted corrections are similar for each cluster and so cannot account for the 0.75 $\text{s}$ change in mean residual between the two clusters. The mean of the southern cluster is similar to the global mean and so it is the northern cluster that has a substantial unexplained anomaly (mean of 0.44 $\text{s}$ after ellipticity corrections). Such a dramatic change occurring over a small spatial region implies that a ULVZ at the base of mantle may be playing a role. Evidence for ULVZs in this region is somewhat contradictory with Persh et al. [2001] favoring a simple CMB and Havens and Revenaugh [2001] reporting evidence for a ULVZ with significant lateral variations on a scale of 200 km. This discrepancy may be resolved if the top of the ULVZ is diffuse and so less visible to the short-period data used by Persh et al. [2001].

[37] Matching the strong positive anomaly in PKiKP-PcP times would require the PcP bounce points to sample normal CMB, while either one or both of the PKiKP penetration points sample a ULVZ. A typical ULVZ model from Havens and Revenaugh [2001] consists of a 10% decrease in $P$ wave velocity over a 15 km interval, which gives a one-way travel time perturbation of approximately 0.11 $\text{s}$ at vertical incidence. Because of trade-offs among model parameters, it is possible that the ULVZ could create a larger travel time anomaly than our example and still be consistent with the waveform modeling of Havens and Revenaugh [2001]. However, even if we consider a ULVZ twice as thick and assume that both PKiKP penetration points are affected and the PcP reflection point is unaffected, the travel time correction becomes 0.45 $\text{s}$, which is not quite large enough to account for the observed deviation of $\sim 0.7\,\text{s}$ (0.44 $\text{s}$ minus global mean of $-0.27\,\text{s}$). The remainder of the anomaly could be explained by a 1–2 km thickening of the outer core.

6.5. PKiKP-PcP Times Beneath Northern North America

[38] The raw residuals recorded at YKA and ILAR are the most strongly negative of the global data set. The single positive residual (0.62 $\text{s}$) arises from a Mexico earthquake which samples the core beneath the central United States. The remaining 14 residuals, which sample the core beneath Alaska and Siberia, range from $-0.6\,\text{s}$ to $-1.8\,\text{s}$ with a mean of $-1.1\,\text{s}$. The anomaly is robust geographically with the magnitude of residuals at YKA (mean of $-1.2\,\text{s}$) similar to that at ILAR (mean of $-0.9\,\text{s}$). The residuals are also uncorrelated with focal depth, implying that shallow subduction zone structure is not influencing the data.

[39] The most obvious explanation for the residuals is the thinning of the liquid core owing to Earth’s ellipticity. However, the travel time perturbations for this factor average only $-0.53\,\text{s}$, and the mean residual after correction remains anomalous at $-0.55\,\text{s}$. Supporting the idea that some other deep earth structure is relevant is the fact that four PKiKP-PcP times sampling the core at significantly higher latitudes, where the ellipticity effect should be even stronger, have a mean residual of only $-0.27\,\text{s}$ [Buchbinder et al., 1973]. Applying 3-D mantle corrections reduces the scatter slightly but still leaves the mean residuals below $-0.5\,\text{s}$ in both cases. Since the mean residual of the ellipticity corrected global data set is about $-0.3\,\text{s}$, there is $-0.2\,\text{s}$ to $-0.3\,\text{s}$ of statistically significant, unexplained, travel time anomaly.

[40] Most of the YKA and ILAR data come from high-quality waveforms recorded at small distances at which the predicted anomalies due to mantle heterogeneity are much too small to contribute to the observed anomalies. Ultralow-velocity zones have been observed at the CMB in this area [Garnero et al., 1998] and may contribute to some of the scatter in the data, but it is unlikely that they could create such a pronounced, consistently negative anomaly. This would require the PcP bounce points to be within the ULVZ and the PKiKP penetration points to be outside the ULVZ for all seven of the strongly negative data. This seems an unlikely coincidence and contradicts the geographical extent of the mapped ULVZs as shown by Garnero et al.
[1998]. Alternatively, the data can be explained by a 1–2 km thinning of the outer core.

7. Implications for Core Asphericity

The vast majority of the PKiKP-PcP travel time residuals can be adequately explained by the combination of effects from mantle heterogeneity, ellipticity, and observational uncertainties. There are only four well sampled locations that show significant and coherent travel time residuals after standard corrections have been made (Figure 7). Of these, the abnormally positive residuals for Central American events recorded at PDAR are most anomalous. However, these data interact with a particularly complicated portion of the CMB, and it is possible that a large fraction, perhaps all, of the anomaly can be explained by the existence of a small yet intense ULVZ that is unimaged by the long-wavelength $P$ wave models we have used to compensate for the effects of mantle heterogeneities. Furthermore, these data occur at distances of $32^\circ–35^\circ$, where inaccuracies caused by using 1-D ray paths in 3-D tomographic models could be significant. It is expected that 1-D ray paths will more closely match true 3-D ray paths when there is little ray bending, as is the case for $PcP$ and PKiKP at smaller distances. For these reasons we prefer not to interpret the PDAR data in terms of anomalous core structure.

It is more difficult to explain the three remaining anomalous areas without appealing to aspherical core structure since they occur at small distances where PKiKP-PcP times are less sensitive to mantle structure and where the assumed 1-D ray paths should more closely describe the actual ray paths. These data consist of (1) five Alaska earthquakes recorded at YKA that have a mean negative residual of about $-0.25$ s, (2) nine earthquakes from the Marianas and Japan subduction zones recorded at KSAR that have a mean negative residual of about $-0.50$ s, and (3) five Flores Sea earthquakes recorded at ASAR that show a mean positive residual of about 0.4 s. These residuals include corrections for ellipticity and are calculated with respect to the global mean of $-0.29$ s.

There are several empirically based arguments which also suggest that these PKiKP-PcP residuals, although subtle, do reflect aspherical core structure. First, the global set of PKiKP-PcP travel times are dramatically better fit by PREM than by AK135, yet the two reference models have liquid cores that differ in thickness by just 3 km and mean liquid core velocities that differ by only 0.2% (and these variations work in opposite directions). This implies that the PKiKP-PcP differential times are more sensitive to subtle variations in core structure than to mantle heterogeneities or observational uncertainties. Second, the high correlation of PKiKP-PcP residuals between two stations that are separated by 400 km (Figure 9) implies that the signal from deep Earth structure is dominant over observational uncertainties. Third, the rotationally induced thinning of the liquid core is clearly seen in the uncorrected data, even though the maximum theoretical value for this is only $\sim 0.5$ s.

It is less clear what type of aspherical core structure is responsible for the PKiKP-PcP residuals. The seismic data cannot discriminate among the various possibilities by themselves and it is necessary to consider other geophysical and geodynamical arguments when deciding which type of core asphericity is most likely. We briefly examine the plausibility of each of the three main candidates: (1) topography on the CMB, (2) topography on the ICB, and (3) nonhydrostatic lateral variations in $P$ wave speed in the outer core.

7.1. Topography on the CMB

Although a number of global models of long-wavelength CMB topography have been developed [Creager and Jordan, 1986; Morelli and Dziewonski, 1987; Doornbos and Hilton, 1989; Rodgers and Wahr, 1993; Obayashi and Fukao, 1997; Boschi and Dziewonski, 2000], the validity of such models is controversial because of biases induced by smooth model parameterizations, uneven data sampling, and the difficulty in separating signals from CMB topography with those from lower mantle heterogeneities [Pulliam and Stark, 1993; Stark and Hengartner, 1993; Morelli and Dziewonski, 1995; Garcia and Souriau, 2000]. We select a recent model of CMB topography developed with a cellular parameterization and extensive geographical coverage [Boschi and Dziewonski, 2000] to examine the potential impact on our PKiKP-PcP data. We correct both PKiKP pierce points and the $PcP$ reflection points using first-order perturbation relations and find that the variance of the residuals increases slightly. For example, the standard deviation goes from 0.44 s to 0.51 s when topographic corrections are added to a data set already corrected for ellipticity and 3-D mantle structure.

Considering the PKiKP-PcP kernel (Figure 2), the three anomalous regions could be explained by CMB topography varying from a 2 km depression beneath Japan to a 1 km depression beneath western Canada to a 1.5 km elevation beneath northwestern Australia. The preferred CMB topography model of Boschi and Dziewonski [2000] is fairly consistent with the Alaska and Japan depressions; however, it is inconsistent with the CMB elevation north-west of Australia. Furthermore, this model predicts substantial negative residuals for data recorded at TXAR and PDAR, and large portions of the CMAR data, which are not observed.

Owing to the limited geographical coverage of the PKiKP-PcP data, and the controversy over the accuracy of global models of CMB topography, it is perhaps more useful to consider whether the amplitude of relief on the CMB required by the PKiKP-PcP data (3.5 km) is reasonable if this phenomenon is the sole cause of the observed anomalies. At very long wavelengths, CMB topography is constrained to be $\sim 500$ m from geodetic observations of the free core nutation [Gwinn et al., 1986]. At short wavelengths, similar constraints on CMB topography (300–500 m) arise from the study of seismic energy scattered on the underside of the CMB [Earle and Shearer, 1997]. However, there are few strict observational constraints on intermediate wavelength topography, and current models based on seismic travel time tomography have peak-to-peak amplitudes as high as 15 km [Boschi and Dziewonski, 2000]. Thus the requirement of 3.5 km of CMB relief, assuming all of the signal is created by CMB topography, does not violate any observational constraints.

The nonelliptical portion of the equipotential surface at the CMB created by mass anomalies in the mantle is
thought to have peak-to-peak variation on the order of 1 km [Forte and Peltier, 1991] and so cannot account for substantial CMB topography. It is also unlikely the dynamic topography created by mantle convection could generate significant CMB topography [S. Zatman, personal communication, 2002]. The best explanation for maintaining CMB topography of a few kilometers is via isostatic compensation of a distinct layer at the base of the mantle (D∗). On the basis of observations of a seismic discontinuity in the lowermost mantle, it is reasonable to assume that D∗ has a density 1% larger than the overlying mantle and has lateral thickness variations of ~200 km [Ryssev et al., 1998]. This corresponds to CMB topography variations of ~2.5 km and so is the right order of magnitude to explain the PKiKP-PcP residuals.

7.2. Topography on the ICB

[51] There are essentially no observational constraints, seismic or otherwise, on ICB topography. Though the density contrast at the ICB is small, mass anomalies in the mantle can only account for about 300 m of peak to peak topography [Forte and Peltier, 1991]. Furthermore, the probable low viscosity of the inner core [Buffet, 1997] implies that any nonhydrostatic topography at the ICB would be difficult to maintain. However, if a suitable mechanism could be found for generating and sustaining ICB topography, the PKiKP-PcP data presented here would limit peak-to-peak topography to be <4 km (assuming CMB topography is not positively correlated).

7.3. Lateral Variations in Outer Core Wave Speed

[50] An upper bound on lateral variations in the radially averaged P wave structure of the outer core can be estimated by assuming ICB and CMB topography are negligible. Approximating the outer core ray paths as vertical gives a δt in seconds of about ~481 × δVV/V, and so δVV/V of about 0.05%, 0.10%, and −0.08% could accommodate the three anomalous regions beneath Alaska, Japan, and Australia. However, these values are 3 orders of magnitude greater than the theoretical, geodynamical upper limit for a fully convecting outer core [Stevenson, 1987]. Theoretical limits on lateral heterogeneity for a stably stratified region in the outer core, perhaps just below the CMB, are more flexible but are still much smaller than the PKiKP-PcP observational constraints. For example, mapping the travel time anomalies to the outermost 200 km of the liquid core gives δVV/V of about 1.0%, which is still 2 orders of magnitude larger than the highest possible theoretical bound [Stevenson, 1987].

[51] It appears that the only way to interpret PKiKP-PcP residuals in terms of nonhydrostatic outer core structure, and remain consistent with the standard geodynamical arguments, is to decouple velocity perturbations from density perturbations. For an isotropic material this would require an exceptionally bizarre rheology [Widmer et al., 1992], and so an outer core with significant anisotropy in P wave velocity is required. However, Romanowicz and Breger [2000] have argued that the amount of outer core anisotropy required to fit anomalously split normal modes is unreasonably high and instead speculate that an outer core with regions of small but finite rigidity is a more viable possibility. Either of these models seems overly exotic with respect to the subtlety of the signal observed in the PKiKP-PcP data.

8. Conclusions

[52] The 302 differential PKiKP-PcP travel times assembled in the present study are an order of magnitude larger than all previous PKiKP-PcP reports combined. The fact that many of our observations came from earthquakes that were relatively small (41% had a magnitude smaller than 5.0 mb) and from many different geographical regions implies that the impulsivity and radiation pattern of the source are the key factors in observing precritical PKiKP. Unusually large earthquakes and/or energy focusing by deep Earth heterogeneities are not prerequisites for observing precritical PKiKP. In the past some observations of PKiKP-PcP amplitude ratios, which constrain the density jump across the ICB, have been dismissed based on the assumption that precritical PKiKP is observed only when it has an anomalously high amplitude. We intend to revisit this issue in a future study using the IMS data.

[53] Our observations of near-vertical PKiKP phases at frequencies of 1 Hz place a strong constraint on the thickness of the ICB, requiring that the solid-liquid transition take place within an interval of less than about 5 km [Cummins and Johnson, 1988]. The absence of validated precritical PKiKP arrivals in the distance range of 57° < Δ < 90°, the so-called transparent zone, is also consistent with the ICB being extremely sharp, having a thickness less than 3 km [Cummins and Johnson, 1988]. However, to confirm this point a systematic search for PkiKP at these distances, using earthquakes with proper focal mechanisms, will need to be carried out. In general, our results reinforce the conventional wisdom that the ICB is simple and sharp over large portions of the Earth. The existence of substantial partial melt in the inner core [Singh et al., 2000] is not excluded by these observations, especially if any “mushy” dendritic zone between solid and liquid is overlain by a thin, hardened “crust”. This can be accommodated by models of inner core growth that include the mechanical effects of sedimentary compaction of the solid iron precipitates in addition to the thermodynamic effects of the multicomponent, liquid-to-solid phase transition [Sumita et al., 1996].

[54] The PKiKP-PcP travel times are well fit by PREM and the majority of the residuals can be explained by mantle heterogeneities, the effect of ellipticity, and observational uncertainties. There are just three geographical locations where a robust signal remains after standard corrections are applied, constraining the maximum unexplained variation in PKiKP-PcP residuals to be about 1.0 s. Assuming a homogeneous outer core this corresponds to a maximum peak-to-peak variation in outer core thickness of 3–4 km in the regions sampled. Hence topography on the CMB is limited to a similar range unless significant correlated topography exists at the ICB as well. However, owing to the low viscosity of the inner core it is unlikely significant nonhydrostatic topography can be maintained. We prefer to interpret the anomalous PKiKP-PcP residuals solely in terms of relatively long wavelength CMB topography with a peak-to-peak amplitude of 3–4 km. This type of topography is best explained by isostatic compensation of
defined as the equatorial and polar radii. Denoting the equatorial rotation shown as (a) ellipticity and (b) difference between the equatorial and polar thickness variations in a distinct D″ layer at the base of the mantle.

Appendix A: Effect of the Reference Density Model on Travel Time Corrections for Ellipticity

The theoretical travel time corrections for Earth’s ellipticity that are commonly used by seismologists are based on first-order perturbations to a 1-D reference velocity model [Dziewonski and Gilbert, 1976; Kennett and Gudmundsson, 1996]. Previous work has shown that these corrections are relatively insensitive to the particular velocity model used; however, they do depend strongly on the function used to describe the variation of Earth’s ellipticity, ϵ, with radius, r. This function, ϵ(r), in turn depends on the radial density model, ρ(r), the ellipticity at Earth’s surface, ϵ₀, and the assumption that Earth behaves hydrostatically [e.g., Jeffreys, 1970]. It is expected that ϵ(r) is relatively insensitive to ρ(r) and therefore that the ellipticity travel time corrections are insensitive to not just the reference velocity model but also to the reference density model.

We have double checked this assumption by calculating ϵ(r) using PREM’s density model and comparing it to the ϵ(r) used by Kennett and Gudmundsson [1996] to generate the ellipticity corrections we have used in this paper. The ellipticity model used by Kennett and Gudmundsson [1996] is presented by Bullen and Haddon [1973] and derived from the density model B₂ [Bullen and Haddon, 1967] (B. L. N. Kennett, written communication, 2002). The calculation of ϵ(r) can be accomplished by numerically integrating the second-order differential equation known as Clairaut’s equation [Jeffreys, 1970, p. 185] with the additional constraint that dϵ/dr goes to zero at r = 0. Alternatively, Clairaut’s equation can be transformed into an easily integrable first-order differential equation involving the Radau variable,

$$
\eta = \frac{d \log \epsilon}{d \log r}, \quad (A1)
$$

that is valid as long as η(r) stays below ~0.6 for the given ρ(r). We find the curves are virtually identical (Figure A1) confirming the conventional wisdom that the ellipticity corrections are insensitive to ρ(r). We further find that η stays below 0.587 and so the Radau approximation is valid for PREM’s density distribution.

Acknowledgments. We thank D. Zhao and L. Boschi for making their 3-D mantle models available and L. Boschi for additionally making his CMB topography model available. The following free software packages were useful: GMT [Wessel and Smith, 1991] and the TauP Toolkit [Crotwell et al., 1999]. Comments from J. Tromp, S. Zatman, G. Smith, and J. Vidale were very helpful. Critical comments from the two referees, P. Richards and S. Rost, and the Associate Editor, J. Revenaugh, significantly improved the quality of the paper. Finally, we thank the prototype International Data Center for making the IMS data accessible. This work was supported by the National Science Foundation under contract EAR-0296078 (see EAR-0087330).

References


J. M. Franks and K. D. Koper, Department of Earth and Atmospheric Sciences, Macelwane Hall, Saint Louis University, St. Louis, MO 63103, USA. (franksjm@slu.edu; koper@cas.slu.edu)

M. L. Pyle, Department of Earth and Planetary Sciences, Washington University, St. Louis, MO 63130, USA. (mpyle@levec.wustl.edu)