The structure of the base of the outer core inferred from seismic waves diffracted around the inner core

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[1] We systematically searched for seismograms of waves diffracted around the inner core (PKPCdiff) from all the temporary seismic arrays with data currently available at the IRIS DMC, as well as some permanent regional seismic arrays including F-NET in Japan and GRF in Germany, to assemble the largest high-quality PKPCdiff database ever created. PKPCdiff waves preferentially sample the base of the outer core and so contain important clues about Earth structure in this region. We measured PKPDF/PKPCdiff differential traveltimes and PKPCdiff/PKPDF amplitude ratios in the distance range of 154°–160° and modeled the observations using grid searches and full wave theory synthetic seismograms. The optimum model found by fitting the differential traveltimes has relatively low velocity at the base of the outer core as in AK135, which is consistent with many previous traveltime studies. However, the optimum model found by fitting the amplitude ratios (PKPCdiff/PKPDF) does not exhibit this feature, and instead is closer to PREM. The discrepancy may be explained by two likely causes. One is that small-scale topography or roughness on the ICB tends to scatter energy away from PKPCdiff waves by generating trailing coda waves. The other is that there exists a thin layer with relatively low Q at the base of the outer core. This might be expected if there are suspended solid particles at the base of the outer core, as proposed decades ago. Both mechanisms could generate smaller PKPCdiff amplitudes without significantly affecting PKPCdiff traveltimes.

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1. Introduction

[2] A large number of seismological studies have suggested that the region just above the inner core boundary (ICB) is distinct from the rest of the outer core. The layer about 400 km above the ICB was originally termed the F-layer and was characterized by a strong low velocity zone [Jeffreys, 1939]. Later Earth models, constructed with more accurate traveltime data, instead defined this as a region of increased velocity, and often included one or more first order discontinuities above the ICB [Bolt, 1962, 1964; Adams and Randall, 1964; Sacks and Saa, 1969]. Ultimately though, these model types were also discarded, as PKP precursors were reinterpreted as energy scattered from mantle heterogeneities near the core mantle boundary (CMB) [Doornbos and Husebye, 1972; Cleary and Haddon, 1972; Haddon and Cleary, 1974]. Hence modern reference Earth models universally have smoothly increasing velocities at the base of the outer core.

[3] There is still some uncertainty, however, about the steepness of the velocity gradient at the base of the outer core.

This is important because the velocity gradient strongly constrains the density gradient in the Earth’s core. If we assume the pressure derivative of the bulk modulus is constant (~3.5) in the outer core [Anderson and Ahrens, 1994], then the Bullen parameter \( \eta \) [Bullen, 1963], which is the ratio between the actual density gradient and the gradient corresponding to uniform composition, only depends on the velocity gradient. The widely used reference Earth model PREM [Dziewonski and Anderson, 1981] has the Bullen parameter \( \eta \sim 1 \) through the outer core, which implies a homogeneous, adiabatic medium.

[4] More recent reference Earth models, such as PREM2 [Song and Helmberger, 1995] and AK135 [Kennett et al., 1995], have significantly lower velocity gradients at the base of the outer core, corresponding to Bullen parameters significantly higher than 1 (Figure 1). In other words, these models imply that near the base of the outer core density increases too quickly to be explained solely by compression, and some sort of change in chemistry or phase must occur. Interestingly, core dynamical studies suggested a slurry layer occupied by free-floating broken dendrites just above the ICB several decades ago [Loper and Roberts, 1978, 1981; Loper, 1978]. However, more recent studies consider a slurry of suspended particles unlikely to exist in the Earth and instead suggest that a thin mushy zone is more probable [Bergman, 2003; Shimizu et al., 2005].

[5] Several other seismic studies have also reported velocity gradients at the base of the outer core that are

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more similar to AK135 than to PREM. Qamar [1973] obtained a P wave velocity model named KOR5 with a very small gradient in the F-layer, and derived an $\eta$ of 3–4. Souriau and Poupinert [1991] examined the PKP($BC + Cdiff$) traveltime data collected by the International Seismological Centre (ISC) and suggested the velocity in the lowermost 150 km of the outer core is significantly lower than that in PREM and that the gradient is nearly flat (zero). Song and Helmberger [1992] suggested a smaller velocity gradient about 400 km above the ICB than PREM by fitting the amplitude of long-period PKP phases recorded at WWSSN stations. Recently, the largest ever data set of PKP waveforms was assembled and the authors found that AK135 provided the best overall fit to the observations [Garcia et al., 2006].

But not all seismic studies support a relatively low velocity gradient at the base of the outer core. Choy and Cormier [1983] found that KOR5 predicts too large PKP$_{Cdiff}$ amplitudes compared to the observations and ruled out the low velocity gradient zone above the ICB, which is consistent with their previous results [Rial and Cormier, 1980; Cormier, 1981]. Both Huang [1996] and Kaneshima et al. [1994] investigated the ICB velocity structure beneath North America’s Pacific seashore and they obtained a slightly smaller P wave velocity than that in PREM, but significantly larger than that in AK135 at the base of the outer core. Most recently Yu et al. [2005] re-examined the outer core velocity structure by analyzing differential traveltimes, amplitude ratios and waveforms of various PKP waves recorded at Global Seismograph Network (GSN) and several regional networks and they found out that the velocity structure at the base of the outer core exhibited a strong hemispherical difference. The data sampling the quasi-eastern hemisphere (40°W–180°E) can be explained by a PREM-type outer core velocity structure, while those sampling the quasi-western hemisphere (180°W–40°E) can be best explained by their preferred model OW, which has velocity closer to PREM2 than PREM at the base of the outer core.

In this study, we focus on studying the structure of the base of the outer core by analyzing PKP$_{Cdiff}$ waves, which are waves diffracted around the ICB and are more sensitive to the base of the outer core than any other phase. Although ray theory predicts zero amplitude after the C-cusp (Figure 2), which occurs at a distance of ~152.5° in PREM or ~155.5° in AK135 for a source at 100 km depth, both theoretical calculations [Cormier and Richards, 1977; Cormier, 1981] and observations [Huang, 1996] indicate that PKP$_{Cdiff}$ has significant energy for several degrees into the inner core shadow zone. At these distances the amplitude decay of PKP$_{Cdiff}$ is mainly controlled by diffraction and is extremely sensitive to the velocity gradient just above the ICB and the period of PKP$_{Cdiff}$.

Systematic study of PKP$_{Cdiff}$ at distances greater than 154° has been somewhat neglected because it is difficult to observe. Small earthquakes do not generate PKP$_{Cdiff}$ waves large enough to be observed above the background noise, and large earthquakes usually have long source durations, which makes PKP$_{DF}$ interfere with PKP$_{Cdiff}$ (the time separation between them is less than 10 s). Similarly, shallow earthquakes sometimes generate PKP$_{DF}$ depth phases that interfere with PKP$_{Cdiff}$, though with careful modeling of the source time functions this problem can often be overcome.

The proliferation of regional broadband arrays over the last decade has made the observations of PKP$_{Cdiff}$ waves less difficult, and has enabled us to assemble a large, high-quality set of PKP$_{Cdiff}$ waveforms. In order to reduce shallow structure effects, we use PKP$_{DF}$ as a reference phase since it has a raypath very similar to PKP$_{Cdiff}$ in the crust and mantle. Although the top of the inner core always has an effect on the traveltime and amplitude of PKP$_{Cdiff}$, our synthetic tests show that this effect is small compared with the effect of the base of the outer core on PKP$_{Cdiff}$. Therefore the differential traveltimes and amplitude ratios between PKP$_{DF}$ and PKP$_{Cdiff}$ measured from the high-quality record sections provide an excellent opportunity to explore the structure at the base of the outer core.

2. Assembly of PKP$_{Cdiff}$ Waveform Database

Because PKP$_{Cdiff}$ is only observable over a very limited range of source-receiver distances and has relatively small amplitude, a closely-spaced seismic array is important to identify the phase and to generate a high-quality PKP$_{Cdiff}$ data set. An array allows small coherent phases to be identified and isolated, and slowness estimates can be made to verify the identity of prospective phases. Another advantage is that the differential traveltimes (PKP$_{Cdiff}$–PKP$_{DF}$) can be measured more accurately by cross-correlating waveforms in the PKP$_{Cdiff}$ time window and the PKP$_{DF}$ time window respectively, instead of cross-correlating the PKP$_{Cdiff}$ window with the PKP$_{DF}$ window for a single station. This is especially important because diffracted waves undergo some shape change as they arrive at larger distances, and so they have shapes dissimilar with PKP$_{DF}$.

We systematically examined the waveforms of earthquakes with depth greater than 60 km and magnitude greater than 5.5 $M_w$ from all the temporary PASSCAL
networks with data open to public researchers at the IRIS DMC, as well as some permanent regional seismic arrays such as F-NET (Full Range Seismograph Network of Japan) and GRF (Gräfenberg) array in Germany. Deep earthquakes have relatively short source durations, which enables us to separate PKPDF and PKPCdiff more easily since the differential time between these two phases is about 10 s. Also, deep sources avoid the potential interference of the depth phase pPKPDF with PKPCdiff.

[12] All the record sections in the distance range $154^\circ$–$160^\circ$ were checked by eye for quality. We picked those events that have high signal-to-noise ratios (SNRs) records and show clear PKPDF and PKPCdiff phases (Figure 3). The amplitudes of PKPCdiff waves decay with distance and are almost at the noise level after a few degrees. In our data selection criteria, we choose all the traces with high SNRs after $154^\circ$ until the one at which the PKPCdiff phase is no longer identifiable. We discuss the possible bias caused by choosing only the “best” data in a later section of this manuscript.

[13] As expected, the PKPCdiff waves are very difficult to observe. We examined 111 record sections from F-net, 110 record sections from GRF, and more than 300 record sections from the temporary networks at IRIS DMC in the distance range $154^\circ$–$160^\circ$. The total number of seismograms inspected was about 11,000. We obtained only 21 record sections, corresponding to 370 individual seismograms, which show high-quality PKPCdiff waves. Table 1 lists all the events and the corresponding seismic arrays used in this study. Among the 21 events in Table 1, 10 events were recorded at F-NET, 3 events were recorded at GRF, 2 events were recorded by the BANJO/SEDA (Broadband Andean Joint and Seismic Exploration of the Deep Altiplano) experiment, 2 events were recorded from INDEPTH-II (International Deep Profiling of Tibet and Himalayas, Phase II) experiment and 4 events were recorded by the INDEPTH-III experiment. Because of the small number of high quality seismograms, the geographical sampling of the core by the PKPCdiff data set is limited (Figure 4). Most of our data sample the quasi-western hemisphere ($180^\circ$–$40^\circ$E) with limited sampling points in the quasi-eastern hemisphere ($40^\circ$–$180^\circ$E).

3. Analysis of Differential Traveltimes (PKPCdiff–PKPDF)

3.1. Data Processing

[14] Differential PKPCdiff–PKPDF traveltimes from our data set are measured as follows. First for each record section, we filter velocity seismograms with a 3-pole Butterworth bandpass filter around 3 s with a one octave band. Although in theory filtering the seismograms at different frequency bands can provide us a separate constraint about the velocity structure at the base of the outer core, unfortunately, we found that it is difficult to get accurate measurements of PKPCdiff–PKPDF traveltimes at higher or lower frequencies. At longer periods, the short time interval between PKPDF and PKPCdiff makes them interfere with each other [Souriau and Roudil, 1995]. At higher frequencies, noise and complicated waveforms prohibit cross-correlation from working properly to get accurate traveltime estimates. Around 3 s, PKPCdiff and PKPDF are well separated, and we can obtain the most accurate measurements for the differential times and the amplitude ratios.

[15] After filtering, we use the multichannel cross-correlation method (mcct) [van Decar and Crosson, 1990] to align the record section within the PKPDF time window to get relative time shifts. Then we stack the aligned record section to get the time of the peak amplitude. By using those relative time shifts and the time of the peak amplitude, we can get the arrival times of PKPDF as described by van Decar and Crosson [1990]. Repeating the same procedure...
for PKP$_{\text{diff}}$, we get the traveltime of PKP$_{\text{diff}}$ and then the differential times PKP$_{\text{diff}}$ – PKP$_{\text{DF}}$. The differential times measured by the mccc method were compared with those obtained with the adaptive stacking technique [Rawlinson and Kennett, 2004], which was proposed specifically for aligning waveforms with somewhat dissimilar shapes. The methods give almost identical results for the PKP$_{\text{diff}}$ – PKP$_{\text{DF}}$ differential times. For each record section, we also invert the arrival times yielded by mccc for the slowness and back azimuth of PKP$_{\text{DF}}$ and PKP$_{\text{diff}}$ phases. We find that they are generally very close to the predicted values in AK135 for most of the events, which ensures that we are indeed analyzing the proper phases. For a few events the deviations are somewhat big, but in these cases the standard

**Table 1.** Events and Seismic Arrays Used in This Study

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<th>Event ID</th>
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<th>Lon, °E</th>
<th>Depth, km</th>
<th>Mag, Mw</th>
<th>Array* Code</th>
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<th>φc</th>
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*For the array code, FN stands for F-net, DI stands for INDEPTH-II, DIII stands for INDEPTH-III, BS stands for BANJO/SEDA, and GRF stands for Gräfenberg.

*b# is the number of seismograms used in the record section.

cφ is the angle of the turning direction of PKP$_{\text{DF}}$ in the inner core with respect to the Earth’s rotation axis.

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**Figure 3.** (a) An example of vertical seismic record sections which show good PKP$_{\text{DF}}$ and PKP$_{\text{diff}}$ waveforms from an event recorded at F-net array with magnitude 6.1 Mw and depth 274 km (Event No. 4 in Table 1). Seismograms are filtered around 3-s using band-pass filtering. (b) The corresponding synthetic seismograms computed from full-wave theory method for an explosive source at depth of 274 km based on PREM model.
errors are also large because of poor source-receiver geometry. As an example, we have poor backazimuth resolution for some events recorded at INDEPTH-III because the strike of the array matches the expected backazimuth [see Zou et al., 2007, Figure 1].

[16] We apply this same method to predict $PKP_{\text{diff}} - PKP_{DF}$ differential times for arbitrary velocity models using synthetic seismograms calculated with the full-wave theory technique [Cormier and Richards, 1977]. This method works with spherically symmetric, anelastic Earth models parametrized as polynomial functions of normalized radius. Full-wave theory correctly accounts for the diffraction and tunneling of body waves at and near the grazing incidence to a boundary by incorporating the Langer approximation [Cormier and Richards, 1988]. This technique is computationally fast and because it operates in spherical geometry it avoids the earth-flattening approximation. It has shown good agreement with the reflectivity method [Choy et al., 1980] and the generalized ray method [Song and Helmberger, 1992] for generating synthetic core phases. An example record section is shown in Figure 3. We compared full wave theory synthetics to those calculated using a frequency-wave number ($f$-$k$) integration method [Zhu and Rivera, 2002; Herrmann and Mandal, 1986] and found slightly different results; however, the $f$-$k$ synthetics contained some artifacts that were presumably related to layering in a flattened Earth model.

[17] Our observed $PKP_{\text{diff}} - PKP_{DF}$ differential traveltimes are plotted relative to PREM in Figure 5. The effect of different source depth is accounted for as follows. For each event’s depth and distance, we find the $PKP_{DF}$ ray parameter predicted by PREM. For that ray parameter, we find the corresponding distance for an event at 100 km depth, which is the distance plotted in Figure 5. All the synthetics are computed with a source depth at 100 km. Although the data are somewhat scattered, it appears that AK135 fits the data better than PREM. We use a statistical method called an $F$ test [Menke, 1989] to verify this quantitatively. First, we calculate the variances of the misfits between the data and the predicted values from the two models, which are 0.1524 for AK135 and 0.2117 for PREM, respectively. The $F$ value is then formed from the ratio of the two variances, which is 1.3891. Using an $F$-calculator (http://faculty.vassar.edu/lowry/tabs.html), we find that there is only a 0.16% probability that the two models are equal, which implies that AK135 is better than PREM at a 99.8% probability in terms of fitting the data. However, we do not observe obvious quasi-hemispherical differences in the data as seen for instance by Yu and Wen [2006]. Although the average differential traveltime residual for the eastern hemisphere (crosses in Figure 5) is a little bigger than for the western hemisphere (gray dots in Figure 5), the scattering ranges of the data overlap. The reason for the lack of a clear hemispherical signal may be that our data set is limited in the eastern hemisphere, or because the hemispherical pattern of the inner core velocity observed by some studies [Niu and Wen, 2001; Wen and Niu, 2002; García, 2002; Stroujkova and Cormier, 2004; Yu et al., 2005; Yu and Wen, 2006] disappears at larger distances (such as the range considered here) as reported by other authors [García et al., 2006; García, 2002; Cao and Romanowicz, 2004]. This would be expected if the quasi-hemispherical pattern in the inner core exists only to relatively shallow depths beneath the ICB.

### 3.2. Data Modeling

[18] In modeling the traveltime data we only vary the velocity structure at the base of the outer core and fix the
velocity in the remainder of the Earth according to PREM. There are several reasons why this type of limited parametrization is justified. First, because the raypaths of $PKP_{DF}$ and $PKPC_{diff}$ are very close in the crust and most of the mantle, the differential times can be attributed mainly to the velocity structure in the lower outer core and at most the upper 500 km of the inner core. For our distance range of $154^\circ - 160^\circ$ the reference phase $PKP_{DF}$ turns at 300–500 km below the ICB, while $PKC_{diff}$ has maximum sensitivity just above the ICB. Second, the difference in inner core and the lowermost mantle velocity structure between standard reference models has very small effects on $PKC_{diff} - PKP_{DF}$ differential times, while the difference in the velocity at the base of outer core between standard reference models has a large effect (Figure 6). Third, we note that the angle of $PKP_{DF}$ with respect to Earth’s rotation axis for all our data is greater than 45° (Table 1). On the basis of the reported inner core anisotropy models [e.g., Creager, 1992; Shearer, 1994; Vinnik et al., 1994; Song and Richards, 1996], the traveltime difference of $PKP_{DF}$ caused by anisotropy is very small (less than 0.2 s). So the observed differential times $PKC_{diff} - PKP_{DF}$ would not be affected significantly by inner core anisotropy. Fourth, we do not observe a strong quasi-hemispherical signal in the data, and so separate modeling for each hemisphere is not warranted. [19] Among velocity models at the base of the outer core reported by different studies [e.g., Qamar, 1973; Dziewonski and Anderson, 1981; Choy and Cormier, 1983; Souriau and Poupinet, 1991; Song and Helmberger, 1995; Kennett et al., 1995; Yu et al., 2005], the main difference is the structure of the velocity and its gradient at the bottom 400 km (or less) of the outer core. In PREM [Dziewonski and Anderson, 1981], the velocity increases with a nearly constant gradient around $0.6 \times 10^{-3} \text{s}^{-1}$. In PREM2 [Song and Helmberger, 1995] and AK135 [Kennett et al., 1995], the velocity gradient decreases from about $0.6 \times 10^{-3} \text{s}^{-1}$ at 400 km above the ICB to nearly zero at the ICB, and the velocity profile with depth is more flat than that in PREM. Therefore we choose 400 km above the ICB as the minimum “pinning depth” at which the models we evaluate are constrained to agree with PREM in value and gradient. [20] We use a systematic grid search to find a model that best fits the data. We let the pinning depth start at 400 km above the ICB and increase it in 50 km intervals until it is 50 km above the ICB. For each pinning depth, we keep the velocity and its gradient the same as in PREM, and let the
gradient decrease linearly with different slopes to satisfy the velocity gradient at the ICB between \(0.6 \times 10^{-3}\) s\(^{-1}\) (the value in PREM) and zero (the value in AK135) with an increment of \(0.01 \times 10^{-3}\) s\(^{-1}\). We use 2nd-order polynomials (3 polynomial coefficients) to describe the velocity structure and have two constraints at the pinning depth and one constraint at the ICB. Therefore we have exact analytical solutions for the polynomial coefficients. In this way, we generate 488 different velocity models at the base of the outer core (8 different pinning depth \times 61 different slopes of the gradient for each different pinning depth).

For each trial velocity model, we compute the synthetic seismograms and get the \(PKP_{\text{diff}} – PKP_{DF}\) differential times with the method described above. Using an L2...
norm misfit function between the synthetics and the data, we find a best fitting model that is very close to AK135 (Figure 7). This is consistent with the results of Garcia et al. [2006] who worked with a much larger data set of PKP arrivals at different distances. The uncertainty of our results is estimated using a bootstrap resampling algorithm [Tichelaar and Ruff, 1989], which randomly resamples the data with replacement to generate a pseudo-data set with the same number of elements as the true data set. An optimal solution is obtained by fitting the pseudo-data set. By repeating the procedure 200 times, we obtain a population of best fitting models and therefore the mean and standard deviation can be estimated (Figure 7).

4. Analysis of Amplitude Ratios (PKP_{Cdiff}/PKP_{DF})

PKP_{Cdiff} amplitudes are very sensitive to the velocity gradient at the bottom of the outer core, just as P_{diff} amplitudes are very sensitive to the gradient at the base of the mantle [Alexander and Phinney, 1966; Phinney and Alexander, 1966; Valenzuela and Wysession, 1998]. Different velocity gradients focus or defocus the seismic energy in different ways. A larger positive gradient bends energy back toward the surface, causing smaller amplitudes of the diffracted waves; a negative or smaller positive gradient traps more energy near the ICB, thus causing larger amplitude of the diffracted wave into the shadow zone.

4.1. Data Processing

Similar to the traveltime analyses, in order to mitigate the effects of shallow structure on PKP_{Cdiff} amplitude, we use the amplitude ratios between PKP_{Cdiff} and PKP_{DF} to constrain the velocity structure at the base of the outer core. The procedure we use to obtain the amplitude ratios is similar to the one used to get the differential times. We measure the peak-to-peak amplitudes from the filtered seismograms for both PKPDF and PKPCdiff. Synthetic amplitude ratios are obtained in the same way from the synthetic seismograms computed with full-wave theory for an explosive source. In order to compare the observed amplitude ratios with the synthetic ones, we apply the source corrections to the data by using radiation coefficients of PKP_{DF} and PKP_{Cdiff} according to the Harvard CMT focal mechanism. Since the take-off angles of PKP_{DF} and PKP_{Cdiff} are very close, the corrections are quite small. The distance of each measurement is adjusted for focal depth using the same approach as in the traveltime analysis. The corrected measurements are presented in Figure 8.
observed amplitude ratios do not exhibit clear hemispherical differences, which is consistent with the amplitude observations of [Yu and Wen, 2006] (see their Figure 5), as well as the traveltime observations presented here (Figure 5). However, the amplitude ratios observed by [Souriau and Roudil, 1995] (see their Figure 8) are consistently larger than our data. They also filtered the seismograms around 3 s to get the measurements, and so the differences must be due to different Earth sampling or data quality. One possibility for the discrepancy is that by using arrays to identify PKP waves we are able to confidently include smaller phases, while the data set of [Souriau and Roudil, 1995] only includes phases that were large enough to be confidently identified on individual seismograms.

### 4.2. Data Modeling

[24] Before we evaluate the different trial models, we perform sensitivity tests to see what Earth properties most affect PKP/DP amplitude ratios. As expected, we find that the difference in the inner core and the lowermost mantle velocity between different reference models does not affect the amplitude ratios significantly, but that the difference in the velocity at the base of the outer core has a very prominent effect on the amplitude ratios (Figure 9). A smaller velocity gradient at the bottom of the outer core tends to generate larger PKP amplitudes, thus larger amplitude ratios (PKP/DP). Therefore AK135 predicts much larger PKP/DP amplitudes than PREM. We also confirmed that the difference in the velocity at the base of the outer core between PREM and AK135 has nearly no effect on the PKP/DP amplitude. Thus the large amplitude ratios from AK135 are totally due to the large PKP/DP amplitudes. Unlike the differential traveltimes, however, the effect of inner core P wave attenuation (Q) on amplitude ratios is not negligible (Figure 9c). Smaller Q in the inner core decreases the PKP/DP amplitude, and thus increases the PKP/DP/DP amplitude ratio. Therefore there is always a trade-off between inner core attenuation and the velocity structure at the base of the outer core.

[25] We perform a similar grid search over the velocity structure at the base of the outer core as we did for the traveltimes. To account for the effects of inner core attenuation, we assign different inner core Q values (300, 400, 500, 600) for each of the velocity models. This range is defined based on the results of several recent investigations [e.g., Dziewonski and Anderson, 1981; Bhattacharyya et al., 1992; Wen and Niu, 2002; Cao and Romanowicz, 2004; Yu and Wen, 2006; Garcia et al., 2006]. Therefore the total number of models we evaluate is four times those evaluated in the traveltime analysis. By minimizing the L2-norm misfit function between the observed and synthetic amplitude ratios in log-space, we find the best fitting model.

[26] Not surprisingly, for different values of inner core Q, we obtain different best fitting models. However, all the models are significantly different from the best model found by fitting the traveltimes. Even when we let the inner core Q be 600, the best model with respect to the amplitude ratios still has significantly faster velocity at the base of the outer core than the best model with respect to the differential times. For example, the dotted line with shadowed area in Figure 7 shows the best fitting model and its uncertainty obtained from bootstrap resampling algorithm with a reasonable inner core Q value of 400. The corresponding predicted amplitude ratio curve is presented in Figure 8 as the dotted curve. A smaller inner core Q value decreases the PKP/DP amplitude, and thus requires a smaller PKP/DP amplitude, which corresponds to a steeper velocity structure. The best model from the traveltimes gives too big PKP/DP amplitudes that are inconsistent with our observations.

[27] We check the robustness of this result with several additional tests. Allowing for a depth-dependent inner core Q, by using different combinations of Q in the upper 200 km of the inner core (Q from 100 to 1000) and Q for the rest of the inner core (Q from 600 to 1000), we find PKP/DP/DP amplitude ratios for the best traveltime derived velocity model are consistently too large compared to the observations. We find the same result for several other high-quality velocity models determined in section 3.2. Considering a frequency-dependent inner core Q model suggested previously [e.g., Li and Cormier, 2002; Cormier and Li, 2002], we experiment with lower corner frequency of 0.1–0.5 Hz and find that they do not significantly change the amplitude ratios as well. We also find that changes to the
density jump and shear velocity jump across the ICB have relatively small changes on the theoretical amplitude ratios, and do not allow the traveltime data to be reconciled with the amplitude data.

5. Discussion

[28] Our $PKP_{\text{Cdiff}}$–$PKP_{\text{DF}}$ differential times give a flatter velocity structure at the base of the outer core than PREM [Dziewonski and Anderson, 1981], which may imply a different chemical composition at the base of the outer core, as suggested by previous studies [Souriau and Poupinet, 1991; Song and Helberger, 1995]. However, our $PKP_{\text{Cdiff}}$ $PKP_{\text{DF}}$ amplitude ratios do not support this feature (Figure 7), instead they prefer a PREM-type velocity structure at the base of the outer core, regardless of inner core Q values. This apparent discrepancy is indirectly supported by many previous studies. The result from the differential times is especially consistent with those studies based on traveltimes [e.g., Souriau and Poupinet, 1991; Song and Helberger, 1995; Kennett et al., 1995; Yu et al., 2005]. Meanwhile, our result from the amplitudes ratios is consistent with previous $PKP_{\text{Cdiff}}$ amplitudes studies [Choy and Cormier, 1983; Cormier, 1981]. Note that it is also consistent with the results of PREM2 [Song and Helberger, 1995]. In that study the authors found that for either constant Q or constant $\tau^*$ in the inner core, the $PKP_{\text{Cdiff}}/PKP_{\text{DF}}$ amplitude ratios predicted by PREM2 were larger than the observations [see Song and Helberger, 1995, Figure 8].

[29] It is also important to point out that any potential bias in our data selection magnifies the discrepancy. By choosing only waveforms with high quality $PKP$ phases, it is possible that our data set is skewed toward abnormally large $PKP_{\text{Cdiff}}$ waves. However, the best fitting traveltimes model would be inconsistent with $PKP_{\text{Cdiff}}$ waves even larger than what we observe. In other words, if a hypothetical perfectly averaged $PKP_{\text{Cdiff}}$ data set has lower amplitudes than our data set, the discrepancy with the traveltimes would be even bigger than what we observe.

[30] To summarize, the observed $PKP_{\text{Cdiff}}$ amplitudes are smaller than what is expected. This is analogous to earlier seismological studies that found unexpectedly small amplitudes for $P_{\text{diff}}$ amplitudes, suggesting some sort of complexity in the lowermost mantle [e.g., Ruff and Helberger, 1982]. In that case, the preferred interpretation of the authors was of a jagged, but positive velocity gradient in the lowermost mantle, implying that $D^+$ is more complicated than a simple, thermal boundary layer. However, in our case the velocity structure just above the ICB cannot be altered to match the observed amplitudes because the predicted traveltimes would then become inaccurate. Instead a non-standard mechanism must exist that acts to reduce $PKP_{\text{Cdiff}}$ amplitudes while not significantly affecting the corresponding traveltimes. It appears that the small-scale heterogeneity inside the inner core observed by many authors [e.g., Vidale and Earle, 2000; Cormier and Li, 2002; Koper et al., 2004; Leyton and Koper, 2007] would not be able to explain this discrepancy because the inner core scattering would decrease the $PKP_{\text{DF}}$ amplitude and thus increase the $PKP_{\text{Cdiff}}/PKP_{\text{DF}}$ ratios. We believe the two primary candidates for explaining the observed discrepancy are (1) small-scale topography or roughness on the ICB that scatters energy from the main $PKP_{\text{Cdiff}}$ phase into trailing coda waves, and (2) a layer at the base of the outer core that possesses low intrinsic Q owing to anelastic mechanisms.

5.1. Hypothesis #1: A Rough ICB

[31] Although strong coda waves following $PKP_{\text{Cdiff}}$ have been observed, they are generally not interpreted as being caused by irregularities at the ICB [Nakanishi, 1990; Tanaka, 2005; Zou et al., 2007]. Slowness information derived from array and polarization analysis tends instead to support scatterer locations in the mantle. However, these studies do not include a contribution to $PKP_{\text{Cdiff}}$ coda waves from complexities at the ICB, and several recent studies have found independent evidence favoring a rough ICB. Poupinet and Kennett [2004] proposed inner core boundary scattering to explain the complexity of the PKiKP coda recorded at Warramunga seismic array and suggested the ICB has some short-scale heterogeneities with lowered wave speed. Koper and Dombrovskaya [2005] found large variation in PKiKP/P amplitude ratios and suggested the presence of heterogeneities at, or very near the ICB. Significant variability of PKiKP amplitudes from explosion sources also led to the suggestion of a mosaic structure at the inner core’s surface [Krasnoshechekov et al., 2005]. Most recently, high-quality earthquake doublet data recorded at different locations suggested that small-wavelength, irregular topography is present at the ICB [Wen, 2006; Cao et al., 2007].

[32] To evaluate this hypothesis numerical simulations of short-period $PKP_{\text{Cdiff}}$ waves for rough ICB models need to be carried out to determine if geodynamically reasonable structures have the desired effect. This is a difficult problem that is outside the scope of this paper, but it is potentially feasible using modern computation approaches such as the axi-symmetric finite difference method that has recently been used to simulate the effect of complicated CMB models on ScS and core grazing S waves [Lay et al., 2006], or the pseudo-spectral method recently applied to inner core scattered waves [Cormier, 2007]. A numerical evaluation is also important because of the counter-intuitive possibility that a rough ICB would lead to enhanced $PKP_{\text{Cdiff}}$ amplitudes, as in Biot scattering associated with seafloor topography [Menke, 1982].

5.2. Hypothesis #2: A Slurry Zone at the Base of the Outer Core

[33] In this scenario, we hypothesize that intrinsic attenuation in the lowermost outer core is much higher compared to the rest of the outer core. Thus the amplitude of $PKP_{\text{Cdiff}}$ would be significantly reduced while traveling horizontally in the low-Q layer, but $PKP_{\text{DF}}$ amplitudes would be less affected because of the steeper raypaths through the layer. This type of model has been suggested on geodynamical grounds and termed a slurry zone [Loper and Roberts, 1978, 1981; Loper, 1978]. Such a feature could be formed by the nucleation and gradual precipitation of solid particles from super-cooled fluid at the base of the outer core. Alternatively, the super-cooling instability could lead to the radial growth of dendritic arms of solid material from the inner core into the outer core, forming a thin, mushy zone; presumably, dendrites from the mushy zone could be broken and distributed throughout the base of the outer core by
large scale convective motions [Copley et al., 1970; Loper and Roberts, 1978]. Internal friction between suspended solid particles and the surrounding fluid is expected for seismic waves traversing the region, and therefore the slurry layer might be expected to have relatively low Q compared to the rest of the outer core.

[34] To our knowledge, there has been only one seismic study in which such a model has been quantitatively evaluated [Cormier, 1981]. In this model (designated as QIC3) Q increased smoothly from 200 to 10,000 as the radius increased from the ICB to about 250 km shallower. QIC3 was rejected because it underpredicted absolute short-period PKP$_{DF}$ amplitudes measured at WWSSN stations [Buchbinder, 1971]. However, we note that QIC3 also possessed an extremely low-Q layer at the top of the inner core, with a minimum Q value of 125 just beneath the ICB tapering to a Q value of 1000 about 250 km beneath the ICB.

To evaluate the possibility of a low-Q layer at the base of the outer core we carry out a systematic grid search similar to the ones described earlier. We fix the velocity structure to the model preferred by the PKP$_{DF}$$-$$PKP_{Caiff}$ differential traveltimes (Figure 7) and search over a range of Q values (from 100 to 600 with 50 as the interval) and thicknesses (from 50 km to 400 km with 50 km as the interval) to find the parameters that best fit the observed PKP$_{DF}$/PKP$_{Caiff}$ amplitude ratios. Assuming an inner core Q of 400, the best model has a Q of 300 over the bottom 350 km of the outer core. In this case, the predicted amplitude ratio curve is very close to the synthetic curve (the dotted curve in Figure 8) from the best model preferred by the amplitude ratios. Unfortunately, the problem is underdetermined as there are significant trade-offs among the model parameters: between inner core Q and outer core Q, and between outer core Q and layer thickness. However, we do find that extremely thin (<100 km) outer core low-Q layers are incompatible with the amplitude ratios because they lead to overly steep decay rates.

[36] The best fitting models found above significantly reduce absolute PKP$_{DF}$ amplitudes, although not as much as QIC3. For instance, with a Q of 300 at the bottom 350 km outer core, PKP$_{DF}$ amplitude decreases by 24%, and PKP$_{Caiff}$ amplitude decreases by 52% at distance of 158°. This effect would be expected to be larger at smaller distances, as PKP$_{DF}$ travel less steeply through the base of the outer core; however, the variation in absolute PKP$_{DF}$ amplitude is on the same order predicted for models with extremely low-Q layers at the top of the inner core and a standard outer core model.

6. Conclusions

[37] We systematically searched the high-quality PKP$_{Caiff}$ waveforms from all the temporary networks with data currently available at IRIS DMC and some regional seismic arrays to assemble the largest PKP$_{Caiff}$ data set ever. Our best model derived by fitting the differential times (PKP$_{Caiff}$$-$$PKP_{DF}$) indicates that the velocity and its gradient at the base of the outer core are significantly lower than those in PREM, which is consistent with many previous studies [e.g., Souriau and Poupinet, 1991; Song and Helmberger, 1995; Kennett et al., 1995; Yu et al., 2005]. The relatively low gradient at the base of the outer core implies that the Bullen parameter is significantly greater than 1, which in turn implies the existence of an abnormally dense layer at the base of the outer core [Souriau and Poupinet, 1991; Song and Helmberger, 1995]. This may explain why recent body wave estimates of the density jump across the ICB are smaller than estimates made from broadly sensitive normal mode observations [Koper and Dombrovskaya, 2005].

[38] However, our best fitting model derived from the amplitude ratios lacks the flat velocity profile at the base of the outer core required by the traveltimes, and instead it is closer to the velocity profile in PREM. This result is also consistent with some previous PKP$_{Caiff}$ amplitude studies [Choy and Cormier, 1983; Cormier, 1981]. We believe the velocity profile at the base of the outer core is in fact significantly flatter than PREM, as revealed by our traveltime analysis, and that observed PKP$_{Caiff}$ amplitudes are reduced by a non-standard mechanism not normally included in synthetic seismograms.

[39] One way to reconcile our traveltime and amplitude observations is with the existence of rough topography on the ICB that acts to scatter energy away from PKP$_{Caiff}$ as it propagates horizontally above the ICB. A complicated or rough ICB has recently been suggested by several authors using independent data sets of waves reflected from the ICB [Poupinet and Kennett, 2004; Koper and Dombrovskaya, 2005; Krasnoshechekov et al., 2005; Wen, 2006; Cao et al., 2007]. This would be an elastic process in which energy is partitioned away from the main phases into later arriving coda waves. Several researchers have examined unusual looking PKP$_{Caiff}$ coda waves, but to date all of the analyses have preferred mantle locations for the scatterers. Nevertheless, the evidence from the PKiKP studies warrants the future numerical experimentation to determine if reasonable models of ICB complexity can produce the needed reduction in PKP$_{Caiff}$ amplitudes, while leaving PKP$_{DF}$ traveltimes relatively unchanged.

[40] A second way to reconcile our observations is with the existence of a relatively low-Q layer at the base of the outer core. For example, a model with flat velocities at the base of the outer core (similar to AK135), an inner core $Q_p$ of 400, and a 350 km thick layer with $Q_p$ of 300 at the bottom of the outer core fits both the traveltime and amplitude constraints. Therefore besides being unusually dense, the base of the outer core may also be unusually attenuating at body wave periods (1–10 s). A physical model for such properties was suggested by Loper and Roberts [1978] in the form of a slurry layer of suspended solid Fe particles in a fluid matrix. Extrapolating results of ultrasonic attenuation experiments with metal powdervicous liquid suspensions [Schultz et al., 1998] suggests that the viscosity of such a slurry would be on the order of $10^{11}$ Pa s to produce the observed body wave attenuation. Recent metallurgical studies, however, suggest that such a slurry layer is unlikely to exist, because of the dendritic growth patterns observed in solidifying metals [Bergman, 1997; Bergman et al., 2005]. However, a recent study (D. Gubbins, G. Masters, and F. Nimmo, A thermo-chemical boundary layer at the base of Earth’s outer core and independent estimate of core heat flux, submitted to Geophysical Journal International, 2007) re-investigated this scenario and proposed a core model in which part of the
light components forms a solid upon freezing of the inner core, and then floats upwards and remelts at some point. To fit the seismic properties and heat fluxes, their preferred model has a 200 km thick layer at the base of the outer core which has less light elements concentration than the rest of the outer core. This provides a physical explanation for the larger density gradient and smaller velocity gradient for the bottom 200 km of the outer core, and the possible low Q there as well.

[41] Although bulk Q in the upper 1000 km of the outer core has been measured to be nearly infinity from high frequency body waves [e.g., Cormier and Ricards, 1976], it is possible there may be some deeper bulk attenuation. This idea of a low-Q layer at the base of the fluid core can be further evaluated in two ways. First, the optimal models reported here can be tested against the large databases of PKD/25P/AB/BC amplitude ratios that exist at distances both smaller and larger than the range considered here [Garcia et al., 2006]. Second, the low-Q models can be tested against normal mode observations. A recent study that used a global search algorithm to fit a large database of normal mode observations found that existing radial \( Q_n \) models of the outer are too large by factors of 2–10 [Resovsky et al., 2005]. Those authors preferred a value of \( Q_n \sim 6000 \) for the outer core, a value much higher than suggested here for the base of the outer core; however, the outer core was treated as a single layer and no allowance was made for variation within this layer. Because the mode data averages significantly over radius, it is possible a low-Q layer is consistent with the data if a more flexible parametrization is used.

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