Antipodal waveform observations of seismic waves diffracting and refracting at the base of Earth's outer core

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ABSTRACT

We approach this analysis of the inner-outter core boundary (IOCB) with antipodal waveform data in the distance range 179.0°–180° to test the hypothesis whether propagation at the base of the outer core is commensurate with diffraction and/or refraction. The propagation paths observed cover about two-thirds of the IOCB surface. Seismic data from seven diameters are examined—Tonga to Algeria (station code TAM), Sulawesi to Amazon (PTGA), northern Chile to Hainan Island (QIZ), two between central Chile and the mainland China (XAN, ENH), and two diameters from New Zealand to both Portugal (PTO) and Spain (ECAL). The adequacy of global Earth models—both 1-D and 1-D core +3D mantle—in fitting the antipodal observations is found to be deficient. Lateral heterogeneity across the propagation paths are mapped and projected to the Earth’s surface for context. We stack data to increase signal-to-noise, and model waveform data via the 3D spectral element method on the EarthSimulator4. With the exception of one path to ENH China, none of the global Earth models match amplitudes of diffracted waves or the stacked amplitudes relative to PKIKP, which travels along the antipodal diameter. Energetic arrivals observed in the waveform data set are modeled as a combination of refraction within a low velocity zone (LVZ) at the base of the outer core, and diffusion around this structure. The observations at the base of the outer core may be further subdivided between paths with a thin LVZ zone (QIZ, XAN, PTGA), and those with a thicker, slower LVZ (TAM). The former paths are characterized by a thin LVZ (20–50 km thick, layer–gradient, respectively) with a velocity of approximately 10.0 km/s. The second group of paths shows large amplitude, complicated waveforms for TAM, corroborated by data from PTO and ECAL. TAM waveforms are modeled with a negative gradient from 100 to 50–75 km thickness above the IOCB, to a basal velocity of 8.8 km/s at IOCB. We also modeled the prospect of non-zero rigidity at the bottom of the outer core, but found no evidence of sensitivity in the core data. The structures fitting the data correlate across the inner–outer core boundary, with a velocity anomaly in PKPbc data juxtaposed immediately above the TAM path beneath SE Asia. More generally, the TAM IOCB path when projected up to the core-mantle boundary (CMB) apparently aligns with regions between the large low shear velocity provinces (LLSVP) at the CMB. Other paths with thinner or negligible LVZ at the IOCB partially underlie the LLSVP when projected to the CMB.

1. Introduction

The fluid nature of Earth’s outer core was determined by Jeffreys (1926), followed in 1936 with Lehmann’s (1936) discovery of the solid inner core. Birch (1952) summarized available data and concluded the core is an iron alloy with a small component of lighter elements. This inner-outter core boundary (IOCB) separates the freezing, growing inner core from the convecting outer core which drives the dynamo generating Earth’s magnetic field. Verhoogen (1961) invoked latent heat of freezing of iron in the core as the main source of energy, whereas Braginsky (1963) proposed compositional convection driven by separation of the lighter component(s) of the outer core alloy by freezing. Chemical heterogeneity at the base of the outer core is directly coupled to density heterogeneity, which induces convection in the outer core (Gubbins, 1977; Loper, 1978). For an Earth model based upon inversion of body waves and normal mode data corrected for Q, Butler and Anderson (1978) determined that the top and bottom of the outer core appear inhomogeneous or non-adiabatic or both. Stevenson (1987) proposed that there are negligible lateral variations in the outer core” due to its low viscosity. More recently, Cormier et al. (2011) reviewed melting
and freezing of the inner core at the IOCB in models proposed by Monnereau et al. (2010) and Alboussiére et al. (2010), who hypothesize that to explain observed, east-west hemispherical differences ascribed to inner core structure the western hemisphere of the IOCB may be crystallizing, while the eastern hemisphere may be melting. This process also predicts the development of a stably stratified dense layer at the bottom of the outer core, depleted in light elements relative to the bulk of the outer core (Cormier et al., 2011; Hirose et al., 2013). However, from a geodynamical perspective it is difficult to sustain large lateral density variations in the outer core, unless the viscosity in this region is significantly larger than usually thought (Cormier, 2009).

Global 1-D Earth models—PREM (Dziewonski and Anderson, 1981) and AK135 (Kennett et al., 1995)—differ at the bottom of the outer core: PREM has a positive gradient whereas AK135 is essentially flat. Other studies show PREM-like or AK135-like gradients (Fig. S1), and this seismic velocity variability suggests contrasts in chemical composition and AK135 (Kennett et al., 1995). Global studies show PREM-like or AK135-like gradients (Fig. S1), and this PREM has a positive gradient whereas AK135 is essentially flat. Other variations in the outer core, unless the viscosity in this region is significantly larger than usually thought (Cormier, 2009)

In this paper we approach the study of the IOCB at antipodal distances (179.0°–180°), which maximizes the path coverage of the IOCB. Prior antipodal study of P wave diffraction (Pdiff) sampling over 99% of Earth’s core-mantle boundary (CMB) is reported in Butler and Tsuboi (2020). We approach this analysis of the IOCB with antipodal waveform data to test the hypotheses whether propagation at the base of the outer core is commensurate with diffraction and/or refraction. The adequacy of global 1-D Earth models—or 1-D core +3D mantle—in fitting the antipodal observations is considered. Lateral heterogeneity among C propagation paths is mapped and projected to the Earth’s surface for context. We compare and contrast antipodal wave propagation above and below the IOCB, and review an apparent correlation with large low shear velocity provinces (LLSVP) at the CMB.

2. Antipodal framework

At the antipode (180°) from an earthquake, the seismic energy converges from all azimuths, thereby amplifying the antipodal phase. The seismic energy from all azimuths about the earthquake source converges from all azimuths, thereby amplifying the antipodal phase. The seismic energy from all azimuths about the earthquake source converges from all azimuths, thereby amplifying the antipodal phase.

The corresponding Cdiff propagation path of PKICP is ~27° referenced to PKIKP, which corresponds to ~582 km at the IOCB; furthermore, the antipodal path samples all azimuths between the earthquake and seismic station. The corresponding Cdiff propagation angle from the prior diffractions studies is obtained subtracting 152.7° (PREM-C-cusp) from the maximum range: e.g., (Ohtaki et al., 2018): 157° - 152.7° = 4.3°

In this paper we approach the study of the IOCB at antipodal distances (179.0°–180°), which maximizes the path coverage of the IOCB. Prior antipodal study of P wave diffraction (Pdiff) sampling over 99% of Earth’s core-mantle boundary (CMB) is reported in Butler and Tsuboi (2020). We approach this analysis of the IOCB with antipodal waveform data to test the hypotheses whether propagation at the base of the outer core is commensurate with diffraction and/or refraction. The adequacy of global 1-D Earth models—or 1-D core +3D mantle—in fitting the antipodal observations is considered. Lateral heterogeneity among C propagation paths is mapped and projected to the Earth’s surface for context. We compare and contrast antipodal wave propagation above and below the IOCB, and review an apparent correlation with large low shear velocity provinces (LLSVP) at the CMB.

2. Antipodal framework

At the antipode (180°) from an earthquake, the seismic energy converges from all azimuths, thereby amplifying the antipodal phase. The seismic energy from all azimuths about the earthquake source coalesces together, and the individual ray paths merge into a ray surface. The ray surfaces comprising the antipodal propagation circumscribe the Earth about the diameter between earthquake source and receiving antipodal station. Although only PKIKP propagates along this diameter (Fig. 1) and is not amplified at the antipode, we find it convenient to review and discuss antipodal data diametrically as designated by their unique diameters. This reconnaissance seeks to understand the seismic energy propagating along the IOCB associated as Cdiff. This energy follows PKIKP by about 9 s. PKIKP includes its whispering gallery PKIKP + PKIIIKP + … propagating at the top of the inner core which arrives within 2.2 s of PKIKP.

The propagation path of C from source to antipode traverses an annular surface at the IOCB (Fig. S3). Antipodal Cdiff (>179°) has been observed and modeled by Rial and Cormier (1980) and Butler and Tsuboi (2010). Butler and Tsuboi (2010) recognized the significant lateral inhomogeneity in the propagation of both PKIKP and Cdiff over paths between Tongan earthquakes and TAM Tamanrasset, Algeria, as
Seismic data from five diameters are examined (Fig. 1) — Tonga to Amazon, northern Chile to Hainan Island, and two between central Chile and the mainland China. Each of these diameters have been previously examined for PKIKP and PKIIKP (Butler and Tsuboi, 2021; Tsuboi and Butler, 2020; Butler and Tsuboi, 2010). In addition, two diameters are included from New Zealand to both Portugal and Spain, of which the former was explored by Rial and Cormier (1980). See Table S1 in the supplement for station–event antipodal pairs, earthquake locations and origin times, and source mechanisms (GCMT, Ekstrom et al., 2012 and NZ earthquake, Anderson et al., 1993).

The key seismic stations studied include TAM.G, PTGA.IU, QIZ.IC, CD, ENH.IC, XAN.IC, PTO.WWSSN, and ECALE.ES. Antipodal data were limited to angular distances >179.0°. The 20 samples/s, vertical data were converted to displacement (m) and high-pass filtered using a two-pass, elliptical filter (high-pass corner = 1/50 Hz, 3 poles, dBStop = 50, Ripple = 2). Fig. 2 plots selected waveforms from Butler and Tsuboi (2021) which are modeled herein. Two additional stations antipodal to New Zealand are reviewed in the context of overlapping IOCB propagation with TAM — PTO, Portugal is an analog, World Wide Standardized Seismograph Network (WWSSN) station studied by Rial and Cormier (1980); and ECAL, Spain is a recent broadband station (established in 2001). An additional Chinese network station considered by Usoltseva et al. (2023) was not included due to site mislocation by 70 km reported by ISCNEWS (2021).

The large arrivals following the Cdiff time at TAM are also observed in the PTO and ECAL diameters (Fig. 3), which differ substantially from the other sites (PTGA, QIZ, ENH, XAN) in Fig. 2. The angular separations of ~22° between TAM and both PTO and ECAL diameters to New Zealand conform with similar wave propagation paths (Fig. 4). Propagation paths for PTO, ECAL, and TAM substantially overlap at the IOCB, and therefore may share common outer core C propagation structures. This suggests that the IOCB structure associated with TAM, PTO, and ECAL differs substantially from the other sites (PTGA, QIZ, ENH, XAN) in Fig. 2.

2.2. Stacking the data

To improve signal-to-noise ratios (SNR), we employed data stacking for each of the five diameters in Fig. 1. Each of these digital, antipodal
stations and their respective earthquake source (Table S1) are used in synthetically modeling features highlighted from stacking the antipodal data (Fig. 5). Wave propagation paths for the several diameters do not share common paths in the crust and mantle. Nonetheless, nearby diameters (e.g., to China) do show similarities indicating common features at the base of the outer core.

We employed phase-weighted stacking (Schimmel and Paulssen, 1997) in conjunction with slant stacking (Niu and Chen, 2008) for a $C_{\text{eff}}$ ray parameter, $p = 2.07$ s/° (AK135). The stacked data are aligned with the nominal, $C_{\text{eff}}$ arrival time at 1253.9 s (average of PREM and AK135). Most of the antipodal events are thrust earthquakes (e.g., Table S1). In stacking both normal and thrust events, the normal events are reoriented (switched sign). Data are independently stacked for each station, employing various subsets of the data based upon distance from the antipode. Stacked data for TAM, PTGA, QIZ, ENH, XAN are shown in Fig. 5. The single antipodal observations at ECAL and PTO were not included in the stacking.

The phase-weighted, slant stacked data reinforce the features shown in Fig. 2 for individual station observations. TAM shows large amplitudes, not commensurate with the other stations. Table 1 summarizes the apparent, stacked C amplitudes (relative to PKIKP) and arrival times (relative to PREM) measured at the first substantial peak for each station. Whereas the other stations show amplitudes declining to back ground within about 20 s following the peak arrivals, the TAM stack continues to grow in amplitude and complexity. Furthermore, in Fig. 2 the relative amplitude of C to PKIKP at TAM is about 0.7 for the 1992 event, as well for the 2001 and 2011 Tongan earthquakes, which collectively are nearly identical triplets (Tsuboi and Butler, 2020).

In a study of inner core structure with antipodal data, both TAM and PTGA stacks showed substantial arrivals in common between PKIKP and PKIIKP—modeled as reflective regions at 100 km and more beneath the ICB (Butler and Tsuboi, 2021). QIZ, ENH, and XAN did not show such energy, which led to a proposed dichotomy within the inner core, wherein TAM and PTGA sample significant inner core structure, but QIZ, ENH, and XAN are much simpler and "PREM-like". This dichotomy in turn requires large-scale heterogeneous structure within the inner core.

In contrast at the base of the outer core, TAM differs from PTGA. This is shown in Fig. 6 For $C_{\text{eff}}$ there is no correspondence between TAM and PTGA stacks after PKIIKP. Therefore, whereas TAM and PTGA share common inner core structure, at the base of the outer core PTGA is better characterized like the China–Chile diameters which differ from TAM, PTO, and ECAL. Hence, only TAM shows inhomogeneous structure both above and below the ICB, though further analysis may include PTO and ECAL in the group.

3. Simulations: 3D spectral element modeling

In modeling the antipodal data set, we approached the problem synthetically using the 3D spectral element method (SEM—Komatitsch and Vilotte, 1998; Komatitsch et al., 2002; Tsuboi et al., 2003; Komatitsch et al., 2005; Tsuboi et al., 2016) as previously applied (Butler and Tsuboi, 2010; Tsuboi and Butler, 2020; Butler and Tsuboi, 2020; Butler and Tsuboi, 2021). The initial model used incorporates a simple PREM (elastic and anelastic) model for the core, a 3D tomographic model—s362wmani—for Earth's Mantle (Kustowski et al., 2008), crustal model CRUST2.0 (Bassin et al., 2000), and ellipticity. We also considered a more recent 3D mantle structure SGL0Eran (Chang et al., 2015), and found essentially same results—see ENH in Fig. 8. For consistency, we used s362wmani where a 3D mantle was included in the SEM model.

In the calculation of the spectral element method, the framing of the earth model divides the entire earth into six quadrangular pyramids, and each quadrangular pyramid is divided into finer quadrangular pyramids and assigned to individual CPUs of the supercomputer to perform the calculation. Initially, we employed SEM synthetics with a resolution of 3.5 s. For the synthesis of narrow low velocity zones at the IOCB, we increased the resolution to 1.6 s. Identical filtering with low-pass corners at 1/4 and 1/2 Hz, respectively, permit direct comparison of the SEM and data. Filter characteristics are two-pass elliptical with 2 poles, dbstop = 100, ripple = 1/2.

In the SEM calculation, the theoretical seismic waveform propagating globally with an accuracy of 1.6 s was calculated by dividing it into 244.7 billion grid points. The total number of cores used in the calculation is 41,334 and the ES4 vector engine (VE) is S168. The grid point spacing in this mesh is 0.94 km on average. For this scale of calculation, it took about 30 min CPU time to calculate the mesh and 4 h 40 min CPU time to calculate a theoretical seismic waveform of 23 min duration—from earthquake origin through PKPab. The size of the mesh is about 41 Thyte.

In synthesizing the antipodal observations, global CMT mechanisms
and locations (Ekström et al., 2012) are used in modeling the earthquake sources. Incorporating a 3D mantle and crust, we include within the SEM synthetics the energy scattered from structure above the core (e.g., crust, upper mantle, D′′ and the core–mantle boundary).

3.1. Models

3.1.1. AK135 and PREM

As a baseline, we consider the SEM fit for QIZ 2009 to the standard global Earth models in Fig. 7, considering PREM, AK135, PREM2, and PREM core +3D mantle model s362wmani. QIZ was chosen as the

Table 1
Measured amplitude and time of the phase-weighted, slant stacked data in Fig. 5.

<table>
<thead>
<tr>
<th>Station</th>
<th>Amplitude a</th>
<th>Cdiff Time b, s</th>
</tr>
</thead>
<tbody>
<tr>
<td>TAM</td>
<td>0.31</td>
<td>6.9</td>
</tr>
<tr>
<td>XAN</td>
<td>0.25</td>
<td>5.7</td>
</tr>
<tr>
<td>QIZ</td>
<td>0.24</td>
<td>3.7</td>
</tr>
<tr>
<td>PTGA</td>
<td>0.20</td>
<td>3.7</td>
</tr>
<tr>
<td>ENH</td>
<td>0.06</td>
<td>4.9</td>
</tr>
</tbody>
</table>

a Relative to PKIKP.

b Delay relative to PREM.

and locations (Ekström et al., 2012) are used in modeling the earthquake sources. Incorporating a 3D mantle and crust, we include within the SEM synthetics the energy scattered from structure above the core (e.g., crust, upper mantle, D′′ and the core–mantle boundary).

Fig. 5. Antipodal data for TAM, PTGA, QIZ, ENH, and XAN are phase-weighted, slant stacked. For each site, the range of distances in the stack are indicated. The Cdiff time is shown relative to PREM time of PKIKP. The white inset view [A] is stacked along the Cdiff ray parameter \( p = 2.07 \text{ s/}^{\circ} \) (average of PREM and AK135). The gray space has ray parameters <2.07, and are not aligned in the stacking but show earlier phases for context. The Inset View shows that the stacked data for Cdiff arrive later than the theoretical travel time (PREM or AK135). Due to the amplitude of the TAM stack, it is not included within the Inset View [B]. Measured peak amplitude and timing of Cdiff are listed in Table 1. The ENH stack has the smallest amplitude stack, as comprised by 8 events.

Fig. 6. The stacked data for TAM and PTGA are compared. The shaded rectangle between PKIKP and PKP-Cdiff indicates the boundary between inner and outer core propagation. Left of the shading there is consistency between TAM and PTGA (except for pPKIKP due to differing earthquake depths). In particular both TAM and PTGA appear to share common inner core structures for PKI100-IKP and PKI250-KP (Butler and Tsuboi, 2021). However, right of the gray shading, neither amplitude or phase relationships suggest commonality in structure between TAM and PTGA at the bottom of the outer core.
largest (both seismic trace amplitude and stacked amplitude) among the small C\text{diff} sites (QIZ, PTGA, ENH, and XAN) for SEM synthesis. For the global models in Fig. 7, the relative C/PKIKP ratios are <0.07—much smaller than the slant-stacked QIZ, PTGA, and XAN ratios (0.2–0.25) in Table 1. For PTGA, XAN, and ENH sites in Fig. 8 we confirm the small effect of mantle scattering observed in Fig. 7 at QIZ for a PREM+3D mantle. None of the global Earth models match either the amplitude of apparent C\text{diff} at QIZ or the slant stacked amplitudes of C/ PKIKP (Fig. 5 and Table 1). ENH is unique in this data set in exhibiting both a small waveform at the C\text{diff} time as well as a small, stacked amplitude (0.06) in Table 1 comparable with the global ratios. Hence, ENH alone is consistent with standard global models. Significantly however, we find that except for ENH, the SEMs for standard global Earth models—whether following a positive velocity gradient at the base of the outer core (PREM) or a decreasing velocity gradient (PREM2 and AK135)—do not fit the antipodal C\text{diff} data.

3.1.2. TAM, PTO, and ECAL

The study of Butler and Tsuboi (2010) first attempted to model the antipodal data from TAM Tamanrasset, Algeria, which exhibited energetic arrivals near the C\text{diff} travel time. This is evident in Fig. 2 for the 1992 event, which is one of three nearly identical earthquakes (1992, 2001, 2011) analyzed by Butler and Tsuboi (2010) and Tsuboi and Butler (2020). The TAM 1992 and 2001 events were modeled (see Fig. S4 in Supplement) by Butler and Tsuboi (2010) with a narrow low-velocity zone at the base of the outer core—a 15% velocity decrease at the inner-outer core boundary relative to PREM, returning to the PREM structure over a linear gradient of 50 km thickness. This hypothetical structure yields large amplitude, later arrivals following C\text{diff} travel times comparable to those observed, as well as increased amplitudes for PKIKP+.

The relatively short period of the oscillations (~5 s) following the TAM C\text{diff} time in Fig. S4, when combined with the mean velocity of the LVZ yields a wavelength λ ~ 50 km, suggests a way forward. As a first order improvement to the model without changing velocities, we simply widened the top of the LVZ to 100 km wide, decreasing the velocity in a linear gradient to a base 50–75 km wide at the bottom model.

Fig. 7. QIZ 2009 data are compared with SEMs computed for standard global 1-D Earth models. The arrival time of C\text{diff} is plotted at the mean of PREM and AK135 times. The QIZ C\text{diff} arrival exceeds the apparent C\text{diff} amplitude of each global model. The relative amplitude of C\text{diff}/PKIKP is <0.07 for each of the global models. PREM (Dziewonski and Anderson) is the whole Earth model without an ocean. PREM+3D mantle replaces the 1-D PREM mantle with the 3D mantle of Kustowski et al. (2008). PREM + PREM2 IOCB replaces the bottom of the outer core with the PREM2 model of Song and Helmberger (1995). AK135 is the global model of Kennett et al. (1995). At the base of the outer core PREM has a positive gradient, whereas PREM2 and AK135 velocity gradients are smaller and flatten at the IOCB.

Fig. 8. Data from four antipodal stations exhibiting a relatively small C\text{diff} arrivals (QIZ, ENH, XAN, and PTGA) are plotted with ±2σ pre-event noise levels (dashed green line). The blue traces are modeled with a PREM core and a 3D-mantle s362wmami (Kustowski et al., 2008). The additional red trace for ENH is modeled with a PREM core and a 3D-mantle SGLOBEran (with Chang et al., 2015). Both 3D mantles show very similar waveforms. Note that distances for the SEM synthetics are consistent with GCMT mechanisms (Ekström et al., 2012).
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arrive together at the antipode have varying thicknesses. Furthermore, given the reasonable fit to TAM data in Fig. 9 we did not sample the space of models for TAM where the LVZ slow velocity $> 8.8 \text{ km/s}$. Considering the low velocity structure at the IOCB which serves as a waveguide for $C$, the phase is not equivalent to $C_{\text{diff}}$ and may be more appropriately called $C_{\text{LVZ}}$ to include LVZ refraction. Nonetheless, there continues to be diffraction above the LVZ—similarly, $P_{\text{diff}}$ diffracts around the CMB, within the higher velocity mantle overlying the lower velocity outer core.

The analog data for PTO are broadly consistent with the nearby TAM observations (Fig. 3). $P_{\text{diff}}$ has been observed at TAM (Butler and Tsuboi, 2020). The 1968 Inangahua earthquake (see Table S1) was a complex event whose source-time function was 20 s in duration (Anderson et al., 1993). The large amplitude arrival following PKIKP may include the inner-core, reflected phase $PKI_{\text{IO}}$, $IKP$ observed at TAM by Butler and Tsuboi (2021). The PTO observation was recorded photographically on an analog WWSSN Sprengnether 15–100 seismometer (Peterson and Hutt, 2014) which has its peak response at 15 s, whereas the TAM sensor is the very broadband Streckeisen STS-1.

To corroborate the fit between TAM and PTO observations, we reviewed current seismic stations on the Iberian peninsula, and noted that ES.ECAL is antipodal to the New Zealand alpine fault, which hosted a Mw 6.5 earthquake in 2013 (see Table S1). The similarity of PTO and ECAL observations (both at 179.3$^{\circ}$) is striking, and further underscores that nearby diameters (~2$^{\circ}$ separation) show similar wave propagation characteristics.

Whereas the stacked amplitude ratio $C/PKIKP$ at TAM is $\sim$0.3, the trace amplitude ratio of $C/PKIKP$ is about 0.7 for both TAM 1992 and PTO 1968, the ratio at ECAL 2013 is $\sim$0.55, suggesting structural and velocity commonalities within the IOCB. The overlapping coverage of TAM, PTO, and ECAL antipodal annuli is shown in Fig. 4. Having obtained a reasonably simple SEM $C$ model which fits to TAM $C$ data and finding corroboration of large $C$ amplitudes at PTO and ECAL—which overlaps antipodal propagation coverage with TAM—we now consider

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**Fig. 9.** The TAM data for the 1992 event (Fig. 2) are phase-weighted, slant stacked for a $C_{\text{diff}}$ ray parameter $p = 2.07 \text{ s/}^{\circ}$. To the right of the $C_{\text{diff}}$ theoretical arrival, $C_{\text{diff}}$ are time-aligned; to the left the ray parameter is $<2.07 \text{ s/}^{\circ}$. Although not time-aligned, these earlier phases are shown for context. Two models [C] are plotted in [A] and [B]. The models (V1.a) differ by the thickness of the base of the low velocity zone (50 and 75 km) where the velocity is 8.8 km/s; both LVZ models begin 100 km above the IOCB. The fit of the SEM synthetic (resolution $>3.5 \text{ s}$) to the stacked data is not perfect or unique; rather the point is that a relatively simple model can explain many of the TAM $C_{\text{diff}}$ features. In particular, [A] fits the initial part of $C_{\text{diff}}$ (blue), whereas [B] fits the later part of TAM (violet). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

**Fig. 10.** This model series (V1.a–V1.e) starts with the TAM model (Fig. 9c) maintaining the basic structure, and successively increases the LVZ slow velocity in steps: 8.8, 9.25, 9.5, 9.75, and 10 km/s. The corresponding SEMs are plotted in Fig. 11 with a resolution $>3.5 \text{ s}$. The thickness of the LVZ is 100 km, with a gradient to 50 km thickness where the velocity is minimum to the IOCB.

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the Chinese and Amazon seismic stations which have measurable, but smaller C than TAM, PTO, and ECAL.

For SEM modeling we chose QIZ as representative of the Chinese stations, with its visible waveform arrival near the time of Cdiff as well as its significant stacked amplitude in Table 1. The full scope of the modeling effort for QIZ is shown in Figs. 10–13. Therefore, QIZ also serves as a proxy for PTGA and XAN in SEM synthesis, given the substantial Earth Simulator time necessary in computing each model.

We first modeled successive iterations on the TAM model to see whether an increasing velocity in the LVZ within a fixed structure (100 to 50 km wide) would converge to a model approximating QIZ. These models are shown in Fig. 10, whereas SEM fits are presented in Fig. 11. For each model iteration, the C arrivals are proximally comparable to the stacked QIZ ratio to PKIKP in Table 1. However, the SEM then grows larger than the QIZ data, and subsequently declines. With decreasing velocity, the relative window of energy does not change substantially. Nonetheless, with decreasing velocity the evolution of the TAM SEM waveform may be observed in the successive models. The challenge is twofold: increase the initial amplitude at the waveform may be observed in the successive models. The challenge is

3.1.3. Data coverage of the IOCB

The Cdiff coverages of the base of the outer core are plotted in Fig. S3 (TAM example), Fig. 4 (TAM, PTO, and ECAL), and Fig. 14 (plots of all sites and their annular surfaces). In each instance, the area sampled is projected to Earth’s surface, which provides context with other core and CMB study regions. Figs. 4[AB] and 14[AB] show areas where paths for TAM, PTO, and ECAL are well separated from paths for QIZ + ENH + XAN + PTGA, and hence may support differing structures. For Fig. 14 [CD] the areas of all coverage overlap around the South Africa–Indian Ocean and North America–East Pacific. In areas where the thin and thicker LVZ overlap in the models, we may reasonably conclude as a presumption that the thinner zones take precedent for simplicity—Butler and Tsuboi (2021) considered the effect of overlapping, antipodal structure on waveform amplitudes.

Several key areal metrics derived for Fig. 4 include: TAM coverage alone, 24%; TAM + PTO + ECAL coverage, 24% without overlap by QIZ + ENH + XAN + PTGA; and no Cdiff sampled, 33%. From these we derive the coverage by only QIZ + ENH + XAN + PTGA as 43%. Based upon the models considered, about 24% of the outer core exhibits a substantially slow LVZ (Fig. 4), whereas 43% is consistent with a thinner LVZ (Figs. 12, 14), and 33% is not sampled in this study.
3.1.4. Rigidity at the base of the outer core

The prospect for non-zero rigidity at the bottom of the outer core has been proposed by Tsuboi and Saito (2002) and Cormier (2009) extending 40 and 400 km above the IOCB, respectively, with shear wave gradient from 0 km/s at the top to 0.5 km/s at the IOCB (Cormier, 2009). We have tested this hypothesis in Fig. 15, where we have computed SEM synthetics for three distance ranges above the IOCB—400, 100, 40 km—incorporating the 0–0.5 km/s shear velocity positive gradient. In the time frame presented by the antipodal waveforms, there appears to be no observable effect upon the waveforms. Hence, the antipodal data set for the outer core finds no waveform evidence favoring a small rigidity at the bottom of the outer core.

3.2. Model summary

Figure 16 presents a summary of the best fitting models. We have been able to fit the large TAM arrival near the time for $C_{\text{diff}}$ with a low velocity channel 100 to 50–75 km thickness at the base of the outer core. The similar PTO and ECAL both show large apparent $C_{\text{diff}}$ arrivals. There is a large overlap in the propagation surfaces of TAM, PTO, and ECAL (Fig. 4). These three observations suggest that apparent $C_{\text{diff}}$ may be modeled as $C_{VZ}$.

For other sites considered—QIZ, XAN, and PTGA—the relative amplitude (Table 1) of the phase-weighted, slant stacked, $C_{\text{diff}}$/PKIKP ratios (0.2–0.25) are larger than any of the comparable $C_{\text{diff}}$ arrivals (<0.07) from the global 1-D earth models PREM, PREM+PREM2 and AK135. Given that the range of velocity structures at the IOCB for the global 1-D Earth models encompasses both positive and flat velocity gradients at the IOCB, there is little room for varying the structure to fit the larger amplitudes observed for the apparent $C_{\text{diff}}$ arrival. For AK135 the “flat” gradient at the IOCB extends 50 km above the boundary, which is about one wavelength at 5 s period. A wider velocity plateau might give entry to longer periods, but not necessarily improve the data fit for the periods of $C_{\text{diff}}$ observed.

Unlike the other Chinese stations, ENH shows $C_{\text{diff}}$/PKIKP ratios of 0.06 (Table 1), which conform well to values for the global Earth models (Fig. 7). As the location of ENH is midway between XAN and QIZ, and overlaps their propagation paths (Fig. 14); we include ENH in their group.

For TAM the recourse was a substantial low velocity zone extending to slow velocities ~8.8 km/s for model V1.a in Fig. 10. A low velocity zone can increase amplitudes. However, testing this for QIZ (Fig. 11) was unsuccessful, in part due to long ringing apparently generated by the LVZ structure. Through trial and error we have found two models that quantitatively fit the observations for QIZ: V3.B and V2.C in Fig. 12. By extension, these models may fit XAN and PTGA, which show comparable times and amplitudes (Table 1). For the ENH, the global models with a 3D mantle can reasonably match the observations, suggesting...
models of attenuation in an iron-nickel alloy, or as alloyed with a lighter, faster element(s). Further synthesis and analysis using the Earth Simulator-4 may answer these questions.

Vertical topography (2–14 km) at the IOCB has been invoked in many papers to explain the amplitude and time variations observed on the IOCB reflected phase, PKIKP and its coda, e.g., Krasnoschekov et al., 2005; Cao et al., 2007; Song and Dai, 2008; Zou et al., 2008; Dai et al., 2012; DeSilva et al., 2018; Wu et al., 2022. PKIKP variation for quasi-vertical ray propagation is not simply matched to \( C_{\text{diff}} \) horizontal ray-sheet propagation over an IOCB arc of ~27° encompassing 24% of the IOCB, as observed herein (Figs. 4, 14, and 17). In principle, topographically-generated scattering at the IOCB should decrease the initial amplitude of \( C_{\text{diff}} \), while increasing its coda duration (e.g., Zou et al., 2008). However, our simple models without topographic structure show the opposite—the SEMs have “ringing” not expressed in the antipodal data.

4. Discussion

For nearly 40 years seismologists have analyzed data to establish the \( V_p \) gradient at the base of the outer core. Early models included a positive gradient in PREM (Dziewonski and Anderson, 1981), a flat gradient for AK135 (Kennett et al., 1995), and a PREM modification to a flat gradient in PREM2 (Song and Helmberger, 1995). Several more recent studies have examined both global and “spot” coverage and attenuation within the basal layer above the IOCB, e.g., Zou et al. (2008), Butler and Tsuboi (2010), Tanaka (2012), Ohtaki et al. (2018), and Adam et al. (2018). Fig. S1 plots a range of these models.

Zou et al. (2008) studied PKPdf — PKPdf travel times and amplitudes from openly available array data in Japan, Germany, Andes, and Tibet in the distance range 154°–160°, finding a flat velocity gradient consonant with AK135 and \( Q_k \sim 300 \). Butler and Tsuboi (2010) modeled antipodal PKPdf data evincing a low velocity layer (~15%) in the basal outer core (Fig. S4), which served as the starting point in this investigation. Tanaka (2012) modeled PKPdf data (~150°–160°) with quasi-hemispheric structure within the inner core (Tanaka and Hamaguchi, 1997) and found the basal outer core is slower above the eastern hemisphere, exhibiting a ~0.1% low velocity zone.

Ohtaki et al. (2018) measured PKIKP and PKPbc travel times and dispersion with respect to PKIKP at the base of the outer core beneath the Northeast Pacific and Australia, above the western and eastern quasi-hemispheres, respectively, of the inner core. The maximum velocity difference between the two regions is 0.04 km/s, with a basal, flat velocity gradient consistent with AK135. Adam et al. (2018) measured differential travel-times and amplitude ratios of PKPbc, PKPdf, PKPdf phases over non-polar paths in the epicentral distance range (149°–171°) distributed globally, to constrain the average \( P' \)-velocity and bulk attenuation profile with depth at the base of the outer-core and at the top of the inner-core. Their study includes an “M-phase”—a large energy in the coda of the PKPbc and PKPdf not predicted by current 1-D reference seismic models, “but most likely originates at the base of the outer-core.” (Adam et al., 2018). Their model of the basal 100 km of the outer core is on average about 0.5% slower than in the reference model AK135—10.238 km/s at the IOCB. We show in Fig. S6 the unconvincing fit of the M-phase model to our QIZ antipodal data.

Global Earth models (1-D and 1-D + 3D mantle) do not fit the antipodal data of Butler and Tsuboi (2021) sampling the basal outer core boundary, whether the model approaches the IOCB with a positive velocity gradient (PREM) or a decreasing, flattening gradient (AK135 and PREM2). We have followed the path of Butler and Tsuboi (2010) and considered Earth models with a low velocity zone (LVZ) immediately above the IOCB. By application of the spectral element method for modeling the antipodal waveforms using the EarthSimulator4, we have converged to two model groups.

Antipodal paths for Chile–China (QIZ, XAN, ENH) and Sulawesi–Amazon (PTGA) form one group characterized by a thin LVZ (20–50

Fig. 13. The model series plots the SEM fits to QIZ 2009 for models in Fig. 12. The model in red was synthesized at a resolution of 3.5 s, whereas the models in blue are computed at a 1.6 s resolution. The waveforms are aligned on the ```

either the absence of an LVZ or a thinner (< 20 km) LVZ layer.

We have focused primarily on the elastic inner and outer core, intending to clarify possible structures which conform to the antipodal data. We also have limited ourselves to simple, “flat” interfaces between radial core layers, without attempting to model the possible effects of scattering at these boundaries, i.e., IOCB topography. We have briefly explored changing the attenuation in the basal outer core—from a PREM bulk \( Q \) of ~57,300 to \( Q = 120 \) —to estimate its effects in the LVZ layer on the “ringing” observed in the SEMs in Fig. 13. For a simple diffraction path length of ~582 km, and frequency ~ 0.3125 Hz, the change in \( Q \) from PREM should decrease the \( A \) amplitude by ~40%. In Supplemental Fig. S5, we see that the SEMs are essentially unchanged from changing the PREM \( Q \). This suggests that the simple approximation of the effect of increased attenuation at the base of the outer core is more complex than can be ascribed to simple layers. Possible considerations include lower values of \( Q \), a transition zone between the high \( Q \) outer core and low basal \( Q \) lower \( Q \) at the top of the inner core, or a frequency dependent \( Q \). Each possibility must also be reconciled with possible
Fig. 14. $C_{\text{off}}$ coverage of the base of the outer core is shown projected to Earth’s surface. The format of the plots is based on Fig. 4 and S3. Globe views are centered upon: [A] the Mediterranean ($40^\circ$, $9^\circ$); [B] Kermadec Islands north of New Zealand ($-37^\circ$, $-179^\circ$); [C] South Indian Ocean ($-45^\circ$, $54^\circ$); and [D] Off Oregon coast. Stripes are oriented along the propagation direction at $5^\circ$ azimuthal intervals. Each annulus covers $\sim 24\%$ of the base of the outer core. Paths are shown for TAM (white transparency), PTO & ECAL (white), QIZ (magenta), ENH (blue), XAN (green), and PTGA (yellow). Note that some overlaps obscure the full path length. Colored pins show the endpoints of their respective antipodal diameters: [AB] TAM (orange) and PTO & ECAL (white); [D] PTGA (yellow, lower right edge). We note that for the Tonga-Algeria and New Zealand–Iberia paths, PKIKP traverses the thin LVZ. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Fig. 15. Possible waveform evidence is tested for a low shear velocity gradient near the base of the outer core between 0 and 0.5 km/s extending down to the IOCB. Three ranges are modeled by SEM: for boundary intervals from 400, 100, and 40 km (above the IOCB), extending down to the IOCB. The base model is PREM + PREM2 at the IOCB. QIZ is shown for data context only. None of the waveform models differ in a substantial way from $V_s = 0$ km/s (PREM2 IOCB).
km thick, layer–gradient, respectively) with a velocity of ~10.0 km/s. ENH is consistent with either a PREM outer core, or thinner LVZ (Fig. 16). A second group of paths for Tonga–Algeria (TAM) and New Zealand–Iberia (PTO, ECAL) shows large amplitude, complicated waveforms for TAM, corroborated by data from PTO and ECAL. Historically, PTO is the first antipodal station of Rial and Cormier (1980). ECAL is a newly discovered antipodal site. TAM waveforms are modeled with a negative gradient from 100 km to 50–75 km above the IOCB, and a constant velocity of 8.8 km/s at the base of the IOCB. The antipodal annular coverage by the combined paths sample ~ 67% of the IOCB. The thin LVZ modeled at ~10.0 km/s is consistent with ab initio molecular dynamics simulations of the properties of liquid iron (10.19 km/s) at the pressure and temperature at the IOCB (Ichikawa et al., 2014). The thin LVZ is consistent with “a stably stratified dense layer at the bottom of the outer core, depleted in light elements relative to the bulk of the outer core” (Cormier et al., 2011). The essence is that the thin LVZ results from depletion of lighter elements, leaving liquid iron as the framework.

For the TAM anomaly, the slower velocity transition to 10.0 km/s occurs within 100 km of the IOCB, along the gradient to 8.8 km/s (Fig. 9). However, when the constitution of the base of the outer core is broadly liquid iron, further removal of light element components has diminishing return on slowing velocities. Based upon the Helmholtz free energy, equation of state estimates at the pressure and temperature at the base of the outer core the P-wave velocity of liquid iron is 5.5–5.7% slower than PREM, or ~ 9.8 km/s (Dorogokupets et al., 2017). To achieve slower velocities approaching 8.8 km/s, we considered within Appendix A in the Supplement the effect of heavy siderophile (“Fe loving”) elements in the LVZ. Whereas siderophiles will contribute to the low velocities, they are not sufficient in numbers to account solely for the observed slow LVZ for TAM.

Are there alloyed compositions of Fe and siderophiles (in low concentrations) that may reduce velocities below that of liquid Fe at the temperature and pressure for the base of the outer core? This challenge falls to mineral physics in much the same way as when the introduction of the PREM model in 1981 invited an explanation for the apparently slow shear velocity of Fe in the solid inner core.

The model in Fig. 9 fits the TAM data. When asked about his F-layer

Fig. 16. The best fitting SEM synthetics are shown. [A] displays TAM fitting model V1.a from Fig. 9. [B] shows QIZ fitting models V2C and V3-B from Fig. 13 at high resolution, 1.6 s. [C] displays ENH from Fig. 8, modeled with a PREM core and two 3D mantle models (s362wmani and SGLobeRani). Please refer to the captions for the original Figs. 8, 9, and 13.
(LVZ) solution for the bottom of the outer core, Sir Harold Jeffreys stated, “Presumably there are an infinity of others, but there is no apparent way of deciding which is the right one, and the first question to decide was whether the times could be fitted at all.” (Birch, 1952). Further efforts may determine a model for the TAM (+ PTO and ECAL) data without a slow LVZ, but that path is not yet clear.

The inner core study of Butler and Tsuboi (2021) considered five antipodal diameters applied herein. Of these QIZ, ENH, and XAN were modeled with PREM-like structure at the top of the inner core, whereas TAM and PTGA presented evidence of shear wave reflections from structure 100 to 250 km below the ICB.

It is significant that TAM presents substantial structure both above and below the IOCB, where the actual connection across the ICB may only be speculated. Furthermore, the change for PTGA from evidence of inner core structure homologous with TAM to a simple, thin LVZ comparable with QIZ, XAN and ENH at the base of the outer core presents another interesting puzzle for speculation. The TAM anomaly is also immediately below a higher velocity region (Fig. S7) detected in PKPbc data by Souriau (2015) in a review of large scale, outer core heterogeneity.

Nearly 2258 km above the IOCB is the CMB, which shows broadly heterogeneous structure both for shear waves (e.g., Kustowski et al., 2008; Chang et al., 2015) and compressional waves (Li et al., 2008; Simmons et al., 2012). Fig. 14 has shown the diverse propagation paths diffracted/refracted in the basal outer core, as projected to the Earth’s surface. In Fig. 17, we show the resemblance of the TAM propagation path with LLVSP structure at the CMB, as presented by models s362wmani by Kustowski et al. (2008) and SGLOBErani by Chang et al., (2015). The geocentric angle of the projected PKPc<sub>diff</sub> propagation path is ~27°, which corresponds to ~582 km at the IOCB and ~ 1640 km at the CMB. Note that the TAM IOCB path apparently aligns with about 90% of the boundary region between the LLVSPs of Africa and the Pacific. Only in the southern Indian Ocean is there apparent LLVSP overlap extended from southern Africa toward Kerguelen. In contrast, ENH—whose propagation is orthogonal to TAM—and nearby XAN and QIZ antipodal paths project across both LLVSPs (Fig. 14).

Based solely upon visual inspection, it appears that more complicated PKPc<sub>diff</sub> propagation (TAM) compares with “normal” regions of the CMB, while simpler PKPc<sub>diff</sub> propagation (ENH, XAN, QIZ, and PTGA) shows in contrast significant coverage projected onto the LLVSPs. This is an unexpected, apparent linkage from seismic structures and processes at the base of the mantle with those at the base of the outer core and within the inner core. Aubert et al. (2008) state, “As the large-scale lower mantle structure has changed little during that time [the past 100–300 Myr], a connection between its present-day pattern and the upper inner core heterogeneous properties is plausible.” This observational hypothesis will benefit from further analysis to corroborate this apparent alignment.

4.1. Speculation

The apparent alignment of features at the CMB with the IOCB—at a distance comparable to a plane ride between Denver and Washington DC—suggests a physical connection of convective mass and/or heat transport. Chen (2021) proffers that “primordial thermochemical models of mantle convection (McNamara, 2019) show that temperatures inside the LLVSPs are higher than that of the surrounding mantle, and that the lower temperature at the LLVSVP boundaries is unfavorable for
melting.

If we consider that “normal” mantle lies between hotter LLSPV, the cooler, normal mantle may overlay convective downwelling in the outer core, where the hotter LLSPV may then correspond with convective upwelling. The downwelling may bring heat and Fe alloyed with lighter elements, precipitating Fe from the Fe alloy mixture, which is then remelted into a Fe-rich low velocity zone in the basal outer core and melt entrained within the top of the inner core. Deguen et al., 2013 elaborated, “Because light elements are partitioned preferentially into the liquid during solidification, iron-rich melt can be produced through a two-stage purification process involving solidification followed by melting (Gubbins et al., 2008). Based on this idea, Gubbins et al. (2008) have proposed a model for the formation of the F-layer in which iron-rich crystals nucleate at the top of the layer and melt back as they sink toward the ICB, thus implying a net inward transport of iron which results in a stable stratification.”

In contrast, the upwelling convection overlies a simple PREM-like structure at the top of the inner core which passively participates in the convective transfer of heat to the LLSPV. In this speculative regard, the CMB may affect change at the IOCB.

If the inner core is indeed rotating with respect to the mantle, then the alignment of TAM inner core structure with the boundary region between the LLSPVs is either fortuitous, or else in sync with the mantle. Gubbins et al. (2011) state, “...any correlation with mantle anomalies and persistence of locality requires the inner core and, to some extent, the core flow to be locked to the mantle.” Aubert and Dumberry (2011) suggest “the inner core rotation then consists in a very slow rotation of a few degrees per million years, superimposed over large fluctuations (at about a tenths of a degree per year). This suggests that the present-day seismically inferred inner core rotation is a fragment of a time-varying signal, rather than a steady super rotation.” Tsuboi and Butler (2020) measured an apparent rotation varying within 0.05°/yr, based on earthquake triplets antipodal to TAM. This suggests that the present-day seismically inferred inner core rotation is a fragment of a time-varying signal, rather than a steady super rotation.

5. Summary

We have reviewed and modeled antipodal data for PKPqref observed for six antipodal Event-Station pairs (>179° geocentric angle), including paths from Tonga to TAM; Chile-China, New Zealand to Spain, and Sulawesi to Amazon. With the exception of ENH China from Chile, the data are not well-fit by standard 1-D global Earth models e.g., PREM, AK135, & PREM2; or the same with a 3-D mantle included. The TAM PKPqref arrival is large and modeled by a relatively thick (50–75 km) low velocity zone (Vp ~ 8.8 km/s) at the basal outer core. For QIZ Hainan Island, China, a smaller PKPqref is observed, which is still larger amplitude than global Earth models, and is consistent with a basal LVZ ~ 10.0 km/s, and either 20 km thick or 50 km gradient to ICB. TAM is representative of ECAL and PTO; whereas QIZ is a proxy for PTGA and XAN.

The TAM data observations show heterogeneous, complex structures at the top of the inner core and at the base of the outer core. Antipodal propagation modeled at sites in China or Amazon show thinner, simpler structures and higher velocity LVZ at the base of the outer core. The antipodal pathway for TAM is orthogonal to that of ENH, which shows simple “PREM-like” structures above and below the ICB.

Finally, reviewing Large Low Shear Velocity Provinces (LLSPV) at the CMB, we find a broad resemblance for the TAM pathway at the ICB projected to the CMB in regions between the Pacific and Africa LLSPVs. The simpler PKPqref paths (e.g., QIZ) project on to the LLSPVs. The ostensible connection between the TAM paths and the intermediary regions between the LLSPVs may not be consistent with steady, super rotation of the inner core with respect to the mantle.

CRediT authorship contribution statement

Rhett Butler: Conceptualization, Methodology, Investigation, Data curation, Visualization, Writing – review & editing. Seiji Tsuboi: Methodology, Investigation, Software, Writing – review & editing.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

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Appendix A. Supplementary data

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References
