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SEISMOLOGICAL INVESTIGATION OF THE ONTONG JAVA PLATEAU AND SURROUNDING PACIFIC UPPER MANTLE

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SEISMOLOGICAL INVESTIGATION OF THE ONTONG JAVA PLATEAU AND
SURROUNDING PACIFIC UPPER MANTLE

BY

AARON C. HIRSCH

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
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OF

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ABSTRACT

The southwest Pacific is a tectonically complicated region consisting of two large igneous provinces (LIPs) on the ocean floor formed approximately 120 Ma. The formation of these two plateaus, the Ontong Java Plateau (OJP) and Manihiki Plateau (MP), along with one further south, is difficult to ascertain due to their remote location. Through the use of seismic modeling, we investigate these two features and the surrounding southwest Pacific to better understand their formation and deformation history.

3-dimensional vertical and tangential seismic wave speed models were computed for the two plateaus and surrounding upper mantle in the southwest Pacific. Pseudo 2D modelling of P/Pn phase arrivals across the OJP crust generated estimates of crustal thickness and mantle wave speed. We then attempted to model the effects of attenuation and wave speed beneath the OJP on Q_{ScS} using 3-dimensional simulations of ScS phases. The seismic modelling provides new datasets for evaluating the LIPs and their formation and deformation history in the southwest Pacific.

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I must thank my family for their support, though their understanding of what I do consists of "working with rocks". Without the support of my wife, Elizabeth Hirsch PharmD, FCCP, none of this would be realized, so thank you. I also want to thank my daughter, Isla. You made it so easy to write this during distance learning with all the completely random noises you must make so you can "concentrate". Thank you, Callan, because if I left you out, you would get mad that I mentioned your sister but not you.

I wanted to say, "I did this for me" and that may be true, but I stood on the shoulders of all who supported me. Thank you for lifting me up.

PREFACE

This dissertation, consisting of three chapters, examines the seismic structure and depths of the Ontong Java plateau and the surrounding Southwest Pacific. In all three chapters, manuscript format is in use.

Manuscript one, titled “Vertical and Tangential wave speed structure of the Ontong Java Plateau and surrounding upper mantle” investigates the upper mantle seismic wave speeds and Vsh (tangential) anisotropy of the southwest Pacific. Abstracts of this work were presented at the AGU annual conference in 2016, 2017 and 2018.

Manuscript two, “Pn analysis of the crust-mantle interface of the Ontong Java Plateau” investigates the crustal thickness and wave speeds of the crust and uppermost mantle beneath the Ontong Java Plateau. Abstracts of this work were presented at the AGU annual conference in 2018.

Manuscript three, “Sensitivity kernels from ScS and ScS²: Interpretation constraints on core reverberations around Ontong Java Plateau” models attenuation and wave speed structure along the ScS phase path and its effects on synthetic seismograms.

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CHAPTER 1

Vertical and Tangential wave speed structure of the Ontong Java Plateau and surrounding upper mantle

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ABSTRACT

Two large oceanic igneous provinces (LIP), the Ontong Java plateau (OJP) and Manihiki plateau (MP) in the Southwest Pacific, took shape from a complicated, but poorly understood geological history. Unraveling the formation and deformation of these Pacific LIPs is not straightforward due to limited available data, a remote location, and atypical geology. Origin hypotheses include melting of a plume head or a fast-spreading triple junction, but distinguishing between these requires a further understanding of 120 Ma of deformation

We performed a transverse-isotropic, scattering-integral, full-waveform tomography utilizing both ambient noise empirical Green's functions and regional earthquake data. Our tomographic model improves upon previous work using permanent and temporary seismic stations, increased model space, and utilizing three components of seismic data (vertical, radial, and tangential).

We image a fast $V_{sv} > V_{sh}$ anomaly in the central OJP while the remaining OJP is V_{sv} (vertical) $< V_{sh}$ (tangential) anisotropic below the crust to ~ 100 km depth. The MP exhibits similar fast V_{sv} wave speeds suggesting a shared formation. This fast V_{sv} wave speed under the OJP with surrounding ambient mantle wave speeds, along with previous geochemical and geophysical datasets, is indicative of a downwelling of eclogite from an 120 Ma partially entrained heterogeneous plume in a spreading center.

INTRODUCTION

The Ontong Java Plateau (OJP) (Figure 1) initially formed approximately 120 Ma with an estimated erupted magma volume of 44 to 57 Mkm³ over 6 – 14 My making it the largest known Large Igneous Province (LIP) (Coffin and Edholm, 1994; Gladchenko et al., 1997; Tejada et al., 2002). In addition to the OJP, the Manihiki Plateau (MP) and Hikurangi Plateau (HP) are believed to be a part of this initial erupted volume covering over 4 Mkm² of ocean floor before moving apart (Taylor, 2006; Ito and van Keken, 2007).

Two main hypotheses have been suggested for the formation of OJP: 1) a plume head at the initiation of a hotspot and 2) a fast-spreading triple junction.

Plume Source

The plume source hypothesis would generate the surface expression of OJP through decompression melting of a Rayleigh-Taylor instability originating at the mantle transition zone or deeper at the core-mantle boundary (Griffiths et al., 1989; Campbell, 1998). This buoyant anomaly would rise as a large plume head with a smaller tail resulting in a significant amount of melting erupted as a LIP followed by a hotspot track of ocean islands. In this instance, the hotspot track has been associated with the Louisville ridge (Neal et al., 1997; Chandler et al., 2012). Geochemical analysis from rock samples in the Solomon Islands indicate a homogeneous ocean-island basalt source with minor element enrichment including siderophile elements, which is consistent with a core-mantle boundary derived plume source (Tejada et al., 1996; Michael, 1999; Neal et al., 1997; Ely and Neal, 2003). Additionally, eclogite

xenoliths (garnet clinopyroxenite) were similarly observed in Solomon Island rock samples of thrust OJP crust suggesting recycled crust brought to the surface during a large mantle upwelling (Ishikawa et al., 2007).

The plume source hypothesis, in isolation, is not universally accepted. OJP is believed to have erupted subaqueously during its entire duration as evidenced by the presence of pillow lavas and oceanic microfossils in samples dredged from the plateau, but studies by Farnetani and Richards (1994) and Korenaga (2005) suggest a large plume head should generate significant uplift to elevate the crust above sea level during eruption.

Fast-Spreading Triple Junction

Passive rift upwelling driven by triple junction spreading of oceanic plates has been suggested as an alternate to the plume source hypothesis (Korenaga, 2005). The Tongareva triple junction represented by the Pacific, Phoenix, and Farrallon plates was active in the area of OJP emplacement (Larson, 1997; Billen and Stock, 2000; Viso et al., 2005; Taylor, 2006). Additionally, sea floor magnetic lineations in the region of OJP indicate a very fast half-spreading rate of 7.7 cm/yr (Larson, 1997). Upwelling due to passive rifting does not generate substantial uplift, which supports a submarine eruption. This hypothesis suffers, however, from a volume perspective because passive rift upwelling would require very a high melt fraction with 100% melting of a 25% eclogite-bearing mantle at the ridge to produce the required igneous crust found at OJP (Korenaga, 2005). In addition, the eclogite fragments would need to be small and in thermal equilibrium with the surrounding upwelling mantle so that the material could be carried towards the surface (Korenaga, 2005); due to its high density, if the

eclogite fragments were too large this would negatively impact a spreading center's ability to bring this material to the surface excluding any inherent thermal buoyancy.

Plume entrained in a Fast-Spreading Triple Junction

An alternative hypothesis for the origin of OJP is to combine the two main hypotheses of a plume source entrained in the flow of a fast-spreading triple junction (Dordevic et al., 2016). This hypothesis merges the advantages of each hypothesis while eliminating most contradictory evidence and studies. The fast-spreading Tongareva triple junction and hot spot track associated with the Louisville Ridge were at the correct location at the time of initial OJP formation. In addition, the combined hypothesis has the necessary high (>25%) melt fraction (Mahoney et al., 1993; Tejada et al., 1996) and a plume head, accommodated by a nearby fast-spreading ridge, would diffuse some of the uplift by ridge capture of at least part of the rising hot mantle plume. Gladchenko et al. (1997) suggested a near-ridge plume created an asymmetric plateau, thickened over OJP and thinner over MP and HP with magma migrating through irregular fractures.

After formation, the LIP, consisting of the OJP, MP, and HP, broke apart intermittently between 120-80 Ma and drifted towards their current locations (Billen and Stock, 2000; Worthington et al., 2006; Hochmuth et al 2015). The OJP experienced several episodes of volcanism between the onset of the LIP and when it struck the Solomon arc subduction zone between 25-22 Ma (Musgrave, 1990; Mahoney et al., 2001). The interaction of the OJP with the subduction zone coincided with the cessation of volcanism at OJP, plate re-arrangements, a subducted Pacific plate breaking off beneath the Solomon Arc, and a subduction polarity reversal with

the Australian plate (Kroenke et al., 1986, Yan and Kroenke, 1993, Musgrave, 1990; Petterson, 1995; Petterson et al., 1999; Mann and Taira, 2004; Hochmuch et al., 2015).

Several previous seismic studies have been conducted on and around OJP. The crust has been studied using active source seismic studies (Furumoto et al., 1970; Furumoto et al., 1976; and Hussong et al., 1979), the crust and upper mantle have been studied utilizing gravity and magnetic techniques (Rose et al., 1968; Sandwell and Renkin, 1988; Schubert and Sandwell, 1989; Nakanishi et al., 1992; Gladchenko et al., 1997; Ito and Taira, 2000) and passive source seismology has looked at the upper mantle and deep structure utilizing Rayleigh wave tomography, SKS splitting, core reverberations, SS precursors and receiver functions (RF) (Richards et al., 2000; Klosko et al., 2001; Gomer and Okal, 2003; Tharimena et al., 2016; Suetsugu et al., 2018; Tonegawa et al., 2019). Previous passive seismic studies by Richardson et al. (2000) and Gomer and Okal (2003) indicated a large low wave speed “root” or “keel” between 30 and 300 km depths. Suetsugu et al. (2019) found overall S wave speeds 2.4% less than IASP91 in the mantle beneath OJP and suggested the low wave speeds Gomer and Okal (2003) observed may have been the same. Covellone et al., (2015) utilized full-waveform techniques and imaged highly abnormal upper mantle shear wave speeds, $>4.75\text{km/s}$ in the central and eastern OJP at depths between 30 and 100 km, and attributed these to an unusual composition, garnet and clinopyroxene residual from melting pyroxenite. i.e., eclogite entrained within a rising plume. The propagation paths used by Richardson et al. (2000) did not cover the same area where the faster shear wave speeds were observed by Covellone et al. (2015). Most recently, Tharimena et al. (2016) attributed similarly fast but slightly slower shear wave speeds

to above 80 km depth over the northern OJP while Tonegawa et al. (2019) found a strong seismic discontinuity beneath the central and eastern OJP at ~60 km depth.

To investigate the OJP and the surrounding Pacific mantle, we performed a comprehensive vertical, scattering-integral, full-waveform tomography between periods of 25 and 200 seconds utilizing both ambient noise empirical Green's functions and seismic data from regional earthquakes. Our tomographic models improve upon previous work using permanent and temporary seismic stations from an increased model area and utilizing three spatial components of seismic data (vertical and tangential).

In addition to a Vsv vertical-vertical (Z-Z) ambient noise and earthquake tomography, we also performed a Vsh transverse-transverse (T-T or tangential) ambient noise and earthquake scattering-integral, full-waveform tomography. Thus far, only the vertical-vertical ambient noise scattering integral tomography has been utilized to extract Rayleigh-wave empirical Green's functions (EGFs) (e.g., Shen, et al., 2012; Gao and Shen, 2012; Gao and Shen, 2014; Covellone et al., 2015, Gao and Shen, 2015; Flinders and Shen, 2017). Additionally, the authors are aware of only one T-T ambient noise tomography. Wang et al. (2019) calculated the 3D EGF sensitivity kernels for T-T and radial-radial (R-R) using adjoint tomography; a spectral-element full-waveform method (e.g., Tromp et al., 2008). Ambient noise T-T tomography has not been attempted previously, presumably for several reasons. T-T and R-R EGFs utilize Love-waves which have a much lower SNR than the Z-Z component making them less desirable to work with considering the computational expense of retrieving ambient noise EGFs (Bensen et al., 2007; Shen et al., 2012). Computing the

sensitivity kernel for T-T or RR EGFs using the scattering integral method requires twice the computational expense and more than double the storage space as two simulations are required for each receiver. Here the resulting Strain Green Tensors (SGTs) are rotated for each measurement; this is described more thoroughly in the methods section. With the addition of our tangential tomographic model, we will be able to better discriminate upper mantle structures that will aid in determining the origin of OJP. Together with the vertical tomography, this will allow us to obtain improved tomographic resolution around OJP and the Pacific upper mantle between 35 and 300 km depth. This improved model will enhance our understanding of the tectonic history of the OJP and MP regions, and the Pacific Indo-Australian plate boundary.

METHODOLOGY

We determined the wave speed structure of the upper mantle beneath the OJP utilizing an iterative, transverse-isotropic and vertical, scattering-integral, full-waveform tomography (Zhang et al., 2012). We utilize empirical Green's functions generated from the raw ambient noise seismic signal (Shen et al., 2012; Gao and Shen, 2014) in addition to data from regional earthquakes. Seismic stations and earthquakes are used from a large area, consisting of NW Australia in the southwest to Hawaii in the northeast, to accurately model the OJP in the sparsely instrumented region of the West Pacific (figure 1.1). A two-step process similar to that of Covellone et al. (2015) is used to determine the wave speed model. First, the ambient noise data is processed (Bensen et al., 2007; Shen et al. 2012) to extract EGFs between seismic stations.

These EGFs greatly improve the coverage in the model because they are not reliant on earthquake paths. We utilize the scattering-integral (SI) method to calculate synthetic EGF within a 3D wave speed model using a finite difference methodology (Zhao et al., 2005; Zhang et al., 2007; Chen et al., 2007a,b; Zhang and Shen, 2008). The simulation also computes a strain green tensor (SGT). The simulated EGFs are compared to the data to derive a frequency dependent travel time anomaly which is then used in combination with the SGT to compute sensitivity kernels (Fréchet derivatives). The differential travel-time measurements determined from the EGFs, forward simulations, and the sensitivity kernels generated from SGTs are used to invert for Earth structure and update the 3D Earth reference model using a linear least squares matrix inversion.

When the long-period wave speed structure from ambient noise has been sufficiently determined, earthquake data are included in the tomographic process. This process is iterated until the improvements in our model are negligible; where the travel-time differences between the synthetics and data are minimized with little or no improvement with additional iterations and the number of usable measurements does not substantially increase. Inversions of the vertical and tangential datasets are conducted separately.

The tomography methods for the tangential component are similar to that of the vertical component with some important deviations that require data and volume rotations (Wang et al., 2019). For a complete description of the ambient noise and earthquake scattering-integral tomography method utilizing the vertical component see Covellone et al. (2015), Zhao et al. (2005) and Gao and Shen (2014). What follows is

a description of the full-waveform scattering-integral tomography method for the tangential component.

Data Preparation

All continuous three component, North (N), East (E), and Vertical (Z), broadband permanent and temporary seismic sensors, actively recording for a minimum of 1 year between 1990 and 2015 within the model space (Figure 1.1), were selected and downloaded from the IRIS DMC. Following the ambient noise processing by Shen et al. (2012), Gao and Shen (2014), and Covellone et al. (2015), the instrument response was removed, the continuous data was cut into daily records, resampled to 1 Hz, and global earthquakes greater than M_w 5.5 were removed. Prior to cross-correlation, a frequency time-normalization (FTN) was applied to each component. Horizontal components, North and East, for a pair of stations, for the recording station and its “virtual” source i.e., the other station, were first rotated to the tangential and radial coordinate system, defined by the interstation azimuth. Then, cross correlations of the tangential components were computed for daily records between the pairs. The cross correlations are then summed where the completed sum represents the T-T EGFs while monthly stacks are used to quantify the EGF error.

Earthquake data within the model area were downloaded from the IRIS DMC with M_w 5.0-7.2, 0 and 400 km depths, and recorded from 1990 to 2017. The earthquakes were initially chosen with a centroid moment tensor (CMT) from the global CMT (GCMT) catalog (Dziewonski et al., 1981). Additional earthquakes between M_w 4.5-5.0 with a CMT solution, away from the active tectonic boundaries within the model space, were also selected to add spatial coverage.

Seismic sensors in the model space were filtered and selected to optimize the full-waveform tomography. First, the vertical signal-to-noise ratio (SNR) for each station-station pair was calculated based on a window around the predicted Rayleigh wave arrival at five overlapping frequency bands (200-100s, 150-75s, 100-50s, 75-30s, 50-25s); the shortest frequency band is used in the SNR analysis. The set of 260 stations were sorted based on vertical-vertical (Z-Z) station-station pairs with a Z-Z SNR greater than 5. To avoid station clustering, stations were sorted and prioritized with a larger number of station pairs with a high Z-Z SNR. Additionally, if two stations were close physically but active at different times with different station-station pairs, these were retained. We limited the total number of stations to 51 to reduce computational costs and temporary data storage.

Earthquake data was additionally filtered to optimally work with the chosen seismic sensors. The earthquake data SNR for the vertical component was calculated based on a window ± 10 s around the predicted window of the P arrival versus the noise windowed 60s before the predicted P arrival. The earthquake data was initially sorted by year (most recent first), then the number of seismic stations with SNR greater than 5 for each earthquake, and lastly by the total number of seismic stations recorded by this earthquake. The earthquakes were then regularized by distance choosing earthquakes recorded by the most stations with the best SNR resulting in 114 earthquakes.

To minimize differences between the vertical and tangential tomography, the stations and earthquakes used in the Z-Z tomography were also used in the T-T tomography. Like above, the SNR for the T-T EGFs was calculated and found to be

similar but lower compared to the vertical station-station pair. For the tangential datasets, the full-waveform simulations do not require Love wave fundamental mode discrimination as the simulations include both the fundamental mode and overtones (Foster et al., 2014). The R-R EGFs were not usable as the SNR was extremely poor.

Synthetic Waveform Generation

An initial starting model is generated using Crust2.0 (Bassin et al., 2000) and AK135 (Kennett et al., 1995) below. We did not use CUB (Ritzwoller et al., 2002) in the upper mantle like Covellone et al. (2015) to avoid any inversion bias towards a specific shear wave speed structure. For ambient noise synthetic generation, using the non-staggered-grid finite-difference method by Zhang et al. (2012), we propagated seismic waves twice between each receiver acting as the virtual source to every other receiver with one simulation using a horizontal gaussian source to the North and one to the East. Earthquake data was also simulated using the same finite difference wave propagation methodology. We simulated an earthquake event using a CMT solution using a 4s bell-shaped source-time function (STF) similar to that of Covellone et al. (2015). The earthquake STF was generated from the event magnitude using

$$L = 10^{\frac{M_w - a}{b}} \quad 1.1$$

$$T_R = \frac{L}{V_R} \quad 1.2$$

where L is the surface rupture length, a and b , from Wells and Coppersmith (1994), are 5.08 and 1.16 respectively, T_R is rupture time, and V_R is a rupture velocity of 2.86 km/s. This inversion problem to update the Earth model can be written as

$$\delta d = A\delta m = \int K_d(\hat{m}, x) \cdot \delta m(x) dV(x) \approx d(m) - d(\hat{m}) \quad 1.3$$

where m is an earth model, d is the data functional of m but also calculated for a starting earth model, $\hat{m}$. The data sensitivity kernels, K_d , are derivatives of the data, d , with reference to the starting earth model volume, V , at every x point within the model (Chen et al., 2007a, 2007b; Covellone et al., 2015). To allow for a direct comparison of the data and synthetic waveforms, a source equalization process was used where the synthetic and data were convolved with the CMT derived STF. The source equalization aligns the data and synthetics in time by the equations

$$S_d(t) = G(t) * \Lambda_d(t) \quad 1.4$$

$$S_s(t) = G(t) * \Lambda_s(t) \quad 1.5$$

$$\hat{S}_d(t) = G(t) * \Lambda_d(t) * \Lambda_s(t) \quad 1.6$$

$$\hat{S}_s(t) = G(t) * \Lambda_s(t) * \Lambda_d(t) \quad 1.7$$

where S_d is the data waveforms, S_s is the synthetic waveforms, G is the Green's Function (GF), and the STFs for the data and synthetic are Λ_d and Λ_s . Unlike, the tangential ambient noise simulations, only one simulation was required for each earthquake. All simulations were completed using two nodes of a multi-node computing cluster with 20 cores per node with solid-state memory with each simulation able to run in parallel. Each synthetic horizontal component (N and E) waveform was extracted at the receiver and rotated to the tangential.

Phase Delay Measurement

The phase delay, dT , is measured between either the ambient noise EGFs or the earthquake data against their respective synthetic by cross-correlation at five frequency bands, 200-100s, 150-75s, 100-50s, 75-0s, and 50-25s. Prior to cross-correlation, the data and synthetics are first filtered using a two-pass butterworth filter.

A generalized seismological data functionals (GSDFs) is used to assign the synthetic waveform, $u_i(\omega)$, generated from equation 1.3 into the corresponding observed ambient noise or earthquake waveform, $\bar{u}_i(\omega)$. This mapping is done in the frequency domain by

$$\bar{u}_i(\omega) = u_i(\omega)^{i\omega[\delta T_p(\omega) + i\delta T_q(\omega)]} \quad 1.8$$

using the phase delay, $\delta T_p(\omega)$, and amplitude reduction, $\delta T_q(\omega)$, though only the phase delay is used (Gee and Jordan, 1992, Chen et al., 2007a,b; Covellone et al., 2015). Measurements that do not meet the minimum SNR and cross-correlation coefficient criteria (Table 2.1 and 2.2) are removed and not used in the inversion.

Scattering-Integral inversion

The scattering-integral (SI) method calculates and stores the sensitivity kernels utilizing a finite frequency full-waveform propagator in 3D with 3D strain and density perturbations (Chen et al., 2007a). The finite frequency SI method has been used by several authors to study the varying tectonic regions around the world (e.g., Chen et al., 2007b; Gao and Shen, 2014; Covellone et al., 2015; Savage et al., 2016; Flinders and Shen, 2017) while finite frequency methods utilizing other inversion techniques has also been broadly used (e.g., Tape et al., 2009; Zhu and Tromp, 2013). The 3D sensitivity kernels computed using the finite frequency SI method do a significantly better job compared to ray-theory methods by more accurately predicting wave speed geometry and perturbations while limiting wave speed anomalies thru reduced smearing (Hung et al., 2004; Becker, 2012). Similar to the SI method, adjoint tomography computes 3D perturbations and sensitivity kernels but uses back propagation from the receiver to the source to generate sensitivity kernels (Tromp et

al., 2008). Both the SI and adjoint methods approach the inverse problem by similar means though diverging computations, with similar end results. Chen et al. (2007a) found that while the adjoint method is better suited and more computationally efficient for global and whole mantle studies, the SI method may be more efficient for regional computations when earthquakes are used, but with large data storage requirements.

Prior to generating the sensitivity kernel, the SGT must be calculated. In the SI method, similar to the synthetic waveforms are calculations, the SGTs are constructed using the same finite difference simulation and 3D reference model, and then are stored for each earthquake or virtual source; this step is computed simultaneously during the 3D forward waveform propagation. To obtain the tangential ambient noise SGTs from a pair of virtual sources to the north and to the east, with a corresponding “good” tangential phase measurement between the rotated synthetic seismogram and the tangential EGF data, the SGTs must be rotated volumetrically. This is done for each virtual source and receiver and stored by transforming the natural coordinate system represented by north, east, and vertical into a general coordinate system of transverse, radial, and vertical (Wang et al., 2019). Using the right-hand rule, an elemental rotation is completed about the vertical axis in three dimensions by

$$A_j^I(\theta) = \begin{bmatrix} \cos \theta & -\sin \theta & 0 \\ \sin \theta & \cos \theta & 0 \\ 0 & 0 & 1 \end{bmatrix} \quad 1.9$$

where A is the rotation matrix and θ is tangential rotation angle between the virtual source and receiver (Wang et al., 2019); in this case the interstation backazimuth. The SGT volumes for the “good” phase anomaly measurements from the earthquake data are not rotated before the kernel calculation. The above SGTs are then used to construct perturbation kernels (J_{in}^{sr})

$$\delta d_{in}^{sr} = \int J_{in}^{sr}(t) \delta u_i^s(x_r, t) dt \quad 1.10$$

for the n th misfit on the i th component of the seismic waveform for the corresponding source, s , and receiver, r where the perturbations, $\delta u_i^s(x_r, t)$, are related to density, p , and strain, c_{jklm} , by equation 1.11 below,

$$\delta u_i^s(x_r, t) = - \int \int \left\{ \sum_j G_{ij}(x_r, t - \tau; x) \partial_\tau^2 u_j^s(x, \tau) \delta p(x) + \sum_{jklm} \partial_k G_{ij}(x_r, t - \tau; x) \partial_l u_m^s(x, \tau) \delta c_{jklm}(x) d\tau \right\} dV(x) \quad 1.11$$

from Zhao et al., 2005; Chen et al., 2007a; Covellone et al., 2015). The sensitivity kernels of the GSDFs in reference to density, p , and strain, c_{jklm} , are calculated, using reciprocity (Aki and Richards, 1980),

$$K_{d_{in}^{sr}}^\rho = - \int dt \int d\tau J_{in}^{sr}(t) \sum_j G_{ji}(x, t - \tau; x_r) \partial_\tau^2 u_j^s(x, \tau) \quad 1.12$$

$$K_{d_{in}^{sr}}^{c_{jklm}} = - \int dt \int d\tau J_{in}^{sr}(t) \partial_k G_{ji}(x, t - \tau; x_r) \partial_l u_m^s(x, \tau) \quad 1.13$$

where $G_{ji}(x, t - \tau; x_r)$ is the transposed Green tensor or receiver Green tensor (RGT) (Zhao et al., 2005; Chen et al., 2007a). The scattering integral is computed by convolving the forward wavefield from the source, s , with the RGT (Chen et al., 2007a, b).

The misfits between the synthetic waveform from the 3D earth model and the observed data are minimized by least-squares. This process is iterated, where the new updated earth model is used to generate synthetic waveform and SGTs, which are then used to calculate new phase anomalies and sensitivity kernels. Ambient noise data was exclusively used first to recover long wavelength structure (Table 1.2). Once the absolute phase delay, dT , decrease per iteration, flattened and the increase in “good”

phase anomaly measurements per iteration slowed or stopped growing, the earthquake data was added. Prior to the least-squares inversion, smoothing and damping parameters were tested to determine the most appropriate combination.

RESULTS

Resolution

The location of the Ontong Java and Manihiki plateaus in the Southwest Pacific has limited its geophysical study. Earthquakes are found primarily along the plate boundary to the south of the OJP and a limited number of usable earthquakes on the Pacific plate result in a directionality bias. Seismic stations are largely located, many temporarily, on the sparse islands around the region, while few ocean bottom seismometer deployments have been completed (Suetsugu et al., 2018, Laske et al., 2009). The most recent work by Covellone et al., 2015 dramatically improved upon previous seismic studies (Furumoto et al., 1976; Hussong et al., 1979; Richardson et al., 2000; Gomer and Okal, 2003; Klosko et al., 2001) utilizing ambient noise and earthquake waveforms measurement using a more robust seismic network than previously available. In this study, we further improved upon the work by Covellone et al. (2015) by increasing the model domain size to incorporate additional seismic stations and utilizing a selective group of temporary stations to increase the number of data measurements for ambient noise analysis. Together, these approaches yield significantly more measurements for the analysis of earthquakes; 51 stations and 114 earthquakes used in this study versus 54 stations and 105 earthquakes in (Covellone et al., 2015). Additionally, while the previous seismic tomographic study of Covellone

et al. (2015) utilized only the vertical measurement from the seismic data, we have included the tangential data for anisotropic analysis. Our improved vertical model and new tangential model resolution can be seen on Figures 1.2 and 1.3 respectively, which shows the computational model domain with a 3.5° and 5° harmonic positive and negative wave speed anomaly pattern with no depth variation for the vertical and 5° and 10° for the tangential. The recovered anomalies are plotted as horizontal slices at 79, 123, and 234 km depth. The recovered vertical resolution is generally very good around the OJP, where the red outline (Figure 1.2) defined in Covellone et al. (2015) is well sampled. Poor sampling due to limited path coverage results in lower resolution in the southern corners, and between Hawaii (Latitude 20° , Longitude 200°) and the Marshall Islands (Latitude 5° , Longitude 180°). Away from the OJP, outside of the red outline, the vertical model is not well sampled nor resolved. Within the region of interest (red outline) the vertical wave speeds are well resolved between 18 and 300 km depth at 5° anomaly sizes and more centrally between 35 and 250 km at 3.5° anomaly sizes as demonstrated by the resolution tests (Figure 1.2). The vertical resolution is similar to that of Covellone et al. (2015). Vertical interpretations will be confined to the area around the OJP, inside of the red outline (Figure 1.2), at 3.5° and 5° for all depth ranges and MP at 5° for all depth ranges and 3.5° at a more limited depth range.

The recovered tangential resolution is good around the OJP, inside the red outline (Figure 1.3), with similar exceptions as the vertical model but with more significant smearing and amplitude loss due to lower data quality and quantity: SNR of horizontal measurements, poor sampling, and limited quality event-station paths. As with the

vertical model, the tangential model's resolution is very poor outside the region of interest, red outline (Figure 1.3). Tangential interpretations will be confined to the areas inside of the region of interest, red outline (Figure 1.2) for the OJP at 5° for all depth ranges. The region of interest is the same as that of the vertical model though the area where resolution is considered good is smaller. As such, care must be taken due to smearing along the edges, especially along the eastern and southern edges, when interpreting features and wave speeds.

Data Fit

The number of measurements per model iteration and travel time residual, dt , are used to determine the robustness and model fit for the vertical and tangential results. Figure 1.4 shows the travel time residuals and number of measurements for the Vsv (vertical) (Figure 1.4a, b) and Vsh (tangential) (Figure 1.4c, d) data sets. Eight iterations were completed for the vertical model with noticeable improvements in ambient noise and earthquake data with each iteration (Figure 1.4a). Additionally, the number of measurements generally increases with additional iterations (Figure 1.4b) as the model improves, allowing more measurements to meet the selection criteria. The decrease in travel time residuals and increase in measurements is not smooth due to long period ambient noise data being used for the first 3 iterations, all periods of ambient noise for iterations 4-6, and earthquake event data included for iteration 7 and 8 (Table 1.1). Five iterations were completed using tangential measurements; four with ambient noise only and the fifth utilizing ambient noise and earthquake data (Figure 1.4c).

The number of ambient noise tangential measurements initially increased, slightly decreased after the 2nd iteration, and then slowly increased upon more iterations (Figure 1.4d). Iterations beyond number 5 for the tangential data were discarded as they showed an increase in travel time residuals for ambient noise measurements, stagnant dt for earthquake measurements, and no change in the number of measurements for earthquakes and ambient noise. Issues related to the iterations beyond number 5 may have been due to the inversion overfitting the solution or the need for damping values stronger than tested.

Other metrics for assessing model fit included visual inspection of waveform simulations of the synthetics and data and analysis of travel time residual histograms to determine absolute amplitude of positive and negative dt .

Seismic Wave Speed

The vertical wave speed depth slices are shown in Figure 1.5 as absolute vertical shear wave speed (V_{sv}) in km/s. At shallow depths ($< 38\text{km}$, Appendix Figure 1.9 and 1.10) the vertical shear wave speed is similar to Covellone et al. (2015) and Richardson et al. (2000) exhibiting a general low V_{sv} wave speed ($< 3.9\text{ km/s}$) southwest of the OJP representative of a thickened Australian plate crust, 4.1 km/s along the boundary between the Pacific and Australian plate, and a fast Pacific plate ($\sim 4.5\text{ km/s}$). This strong vertical shear wave speed difference between the Pacific and Australian lithosphere is a consistent feature in most shear wave tomographic models (Appendix Figure 1.9, 1.10) (Richardson et al., 2000; Covellone et al., 2015). Pacific plate wave speeds in our model are generally in agreement with wave speeds from a smoothed and averaged Pacific plate (Maggi et al., 2006). Additionally, as compared

to the most recent full-waveform Vsv model (Covellone et al., 2015), in the current model (Figure 1.5) the Vsv wave speeds are slightly slower (4.2-4.3 km/s vs 4.4-4.5 km/s) in the oceanic region of the Australian plate south-southwest of OJP at depths between 40 and 100 km depth but similar wave speeds at depths between 114 – 210km depth. Another similar feature to Covellone et al. (2015) are the fast (>4.5 km/s) anomalies North and Northeast of OJP at depths below 114 km that track generally towards MP (Figure 1.5b, c). Overall, the similarities in our wave speeds with Covellone et al. (2015) demonstrates the model reliability and robustness of our observations.

In the upper mantle, the OJP and MP exhibit strong Vsv differences as compared to the surrounding mantle. The very fast (4.75 km/s) anomalies over the central part of OJP (Lat. 162°E, Lon. 5°S; Figure 1.5b,c) are consistent with Covellone et al. (2015) which has the same high Vsv anomalies at depths less than 100 km, slightly more to the NE (Figure 1.5b,c), but still within the OJP; this feature is not observed by Richardson et al. (2000). This fast Vsv speed feature can be seen in Figure 1.6 in all cross-sections, and in Figure 1.5b, c which corresponds approximately to a large negative residual gravity anomaly centered at latitude -2°, longitude 161° (Ito and Taira, 2000). At 100 to 250 km depth, below the fast anomaly, is a spatially extensive slower wave speed region (4.2 km/s – 4.4 km/s) of typical of normal oceanic mantle.

The Manihiki Plateau (MP) shows a similar fast Vsv structure to the OJP but with a few differences. The MP has a strong fast (>4.75 km/s) wave speed anomaly at depths greater than 30 km while the OJP is less coherent at shallow depths; both anomalies extend to ~165 km depth. Additionally, this fast wave speed anomaly is

widespread across the southern half of the MP from 7.5°S to 15°S as compared to the more irregular nature below the OJP (Figure 1.5b,c). This North-South discontinuity cuts across known sub-provinces of the MP including an unfaulted region in the middle (Hochmuth et al., 2015). The wave speed model of Covellone et al. (2015) is inconsistent with our vertical wave speed model at the MP. While the previous model (Covellone et al. 2015) did not target this plateau, the MP in their model is similar to the surrounding mantle at all depths unlike in our model. The MP is on the edge of the ‘high resolution’ model domain in Covellone et al. (2015) and the dissimilarity may be due to smearing along the model perimeter. The current model has significantly better vertical shear wave model resolution over MP compared to Covellone et al. (2015), allowing better imaging of the Vsv anomalies in this region (Figure 1.3). A fast Vsv 4.75 km/s anomaly beneath both MP and OJP between 40 and 100 km depth is consistent with plate reconstructions showing OJP, MP, and HP formed together (e.g., Chandler et al., 2012, Bird, 2003).

The tangential wave speed depth slices are shown in Figure 1.7 as absolute shear wave speed (Vsh) in km/s. The Vsh (tangential) model is smoother with larger spatial anomalies compared to the Vsv (vertical) model as a result of data quality and quantity. Several Vsh wave speed sensitivity tests were completed, varying the starting model and number and type of measurements used per iteration, with each final model having similar features and wave speeds. Slower Vsh (tangential) wave speeds associated with a crustal signature are mostly gone by 38 km depth similar to those of the Vsv (vertical) model due to poor shear wave speed sensitivity to crustal depths. Wave speeds below 38 km depth in the Pacific Ocean are up to 0.4 km/s slower than

average mantle Vsh wave speeds (4.55-4.6 km/s). Slower Vsh wave speeds (4.2-4.3 km/s) are observed to the North and East of the OJP, diminishing with depth to 210 km. A slow Vsh (4.2 km/s) anomaly surrounded by fast Vsh wave speeds is observed at similar depths and in the same central OJP region as the fast Vsv anomaly. The MP has a similarly slow Vsh wave speed (4.2 – 4.3 km/s) over the entire structure to a depth of 210 km before diminishing into regional background Vsh wave speeds. A very fast (> 4.8 km/s) Vsh wave speed anomaly is observed in our model between OJP and MP at 10°S between 180° and 170°W (Figure 1.7a, b, c) coincides with a strong polarization anomaly in Debayle and Kennett (2000) and the Robbie Ridge (Hochmuth et al., 2015). Additionally, in areas where polarization anisotropy from Debayle and Kennett (2000) is smaller, we see slower Vsh wave speeds like those to the North and East of OJP. This generally corresponds to major and minor fracture zones within the Nova Canton Trough including the Phoenix and central Pacific fracture zones (Hochmuth et al., 2015).

Several differences exist between the Vsv (vertical) and Vsh (tangential) wave speed models. The fast Vsv wave speeds (4.7 km/s) under OJP are coincident with a slow Vsh wave speed (4.2 km/s) anomaly that cuts across the Plateau from East to West. This anti-correlation is also observed beneath the MP with fast Vsv wave speeds and slow Vsh wave speeds. These differences are apparent when comparing Figure 1.6 and 1.8 above 100 km depth and West of 155°E . The Vsh (tangential) wave speed model is also of much longer period compared to the Vsv (vertical) model (Figures 1.2, 1.3). This is evident in the southern part of the Vsh model in the deeper

depths where there is significant smearing and high wave speed compared to what is expected.

DISCUSSION

We have identified several intriguing Vsv (vertical) and Vsh (tangential) wave speed anomalies in the vertical and tangential tomographic models deserving of additional discussion. As noted in the Results section, we will focus on the following three wave speed anomalies, 1) a fast 4.7 km/s Vsv but slower 4.2 km/s Vsh wave speed under the OJP, 2) fast (4.7 km/s) and slower (4.4km/s) Vsv wave speeds bifurcating MP with a $\sim$ 4.3 km/s Vsh anomaly, and 3) an area North and northeast of the OJP with a fast 4.6 km/s Vsv and slower 4.2 km/s Vsh.

The fast 4.7 km/s Vsv anomaly under the OJP is similar to that observed and modeled by Covellone et al. (2015). In our model, the fast wave speed anomaly is more localized in the central and east-central OJP compared to the previous study. The location coincides generally with that of the Vsv model from Covellone et al. (2015), where the fast (>4.75 km/s) wave speed anomalies are at similar positions but not as fast in this study (4.7 km/s). This may be attributed to the use of different starting models where Covellone used CUB (Shapiro and Ritzwoller, 2002), a global Vs wave speed model, but we used AK135 (Kennett et al., 1995), where the influence of existing anomalies may affect the inversion. The difference may also be attributed to the difference in the number of measurements used in the final model. Substantial work was done to use the highest quality datasets resulting in 8000 ambient noise and earthquake measurements used in our model compared to 3300 in Covellone et al.

(2015). To determine if the fast V_{sv} wave speed anomaly was required, Covellone et al. (2015) capped wave speeds to different amounts and found that the >4.75 km/s V_{sv} anomaly beneath the OJP was necessary; wave speeds capped at 4.5 km/s degraded the data fit. Richardson et al. (2000) observed a lower wave speed beneath the OJP to 300km depth but used only 4 receivers with all sources to the south which resulted in poor coverage over the high wave speed area we observed. Additionally, the authors confined the wave speed perturbations to 300 km, proactively restricting any deeper wave speed sensitivity and their model was further complicated by trade-offs between crustal thickness and upper mantle wave speeds. Recently, Tharimena et al. (2016) found a similarly fast (>4.75 km/s) wave speed beneath the OJP using SS precursors while Tonegawa et al. (2019) observed a high frequency 4.5 km/s wave speed discontinuity modeled slightly shallower using RF. These recent studies in combination with Covellone et al. (2015), where the same scattering-integral full-waveform tomography method was used but with slightly different datasets, gives us confidence in the >4.7 km/s V_{sv} (vertical) wave speed anomaly beneath the OJP (Figure 1.5).

Covellone et al. (2015) additionally investigated if the fast V_{sv} wave speed was due to anisotropy. In their synthetic testing, they found a better data fit with $V_{sv} < V_{sh}$, compared to $V_{sv} = V_{sh}$ or $V_{sv} > V_{sh}$. The tests simulated earthquakes (3) along the southern edge of the OJP recorded at a seismic station to the North. Our V_{sv} and V_{sh} models show contradictory results to those of anisotropy testing of Covellone et al. (2015). Along the same path, we observe $V_{sv} < V_{sh}$ in the southern and northern OJP and $V_{sv} > V_{sh}$ in the central OJP between 40 and 100 km depth where the fast

OJP anomaly is detected (Figure 1.5b, c and 1.7b, c). Additionally, the area between the OJP and one of the seismic stations used in the Covellone et al. (2015) tests also shows $V_{sv} > V_{sh}$. This demonstrates that a simple anisotropy test from Covellone et al. (2015) across the OJP may cross several anisotropic regimes making it difficult to isolate a preferred wave speed direction.

Lastly, Covellone et al. (2015) calculated upper mantle compositions compatible with fast wave speeds and concluded that the most likely lithology is an ultra-high pressure (UHP) eclogite comprising mostly of clinopyroxene and garnet. An eclogite source for the fast wave speed is incompatible with a V_{sv} (vertical) $< V_{sh}$ (tangential) anisotropy ($\sim 11\%$), assuming the composition used as the source is accurate and V_{sh} wave speeds are acceptable. Eclogite is quasi-isotropic ($< 2\%$ anisotropic), lacking any meaningful anisotropy (Bascou et al., 2001). The UHP eclogite source proposed by Covellone et al. (2015) contained a high proportion of garnet, which minimizes the anisotropy of eclogite because natural samples were randomly oriented by omphacite which has a strong lattice preferred orientation (LPO), but omphacite aggregate, which is also randomly orientated, is also weakly anisotropic (Bascou et al., 2001). If the omphacite in eclogite is aligned during deformation, anisotropy would be greater ($> 2\%$ anisotropic) (Bascou et al., 2001). The lack of anisotropy in eclogite helps to distinguish it from the more general anisotropic upper mantle peridotite composition (Bascou et al., 2001). Calculated V_{sv} wave speeds for peridotite (4.5 km/s) are not as fast as the anomaly below the OJP (Covellone et al., 2015), though peridotite is anisotropic ($\sim 7\%$). An eclogite source cannot be ruled out as a source of the V_{sv} wave speed anomaly as it is seismically fast, and the eclogite composition fits with

models for the origin of the OJP. This suggests that the fast V_{sv} anomaly with the observed V_{sv} (vertical) $>$ V_{sh} (tangential) anisotropy may be composed of a seismically fast, weakly anisotropic eclogite but the source of anisotropy is from tectonic forces during formation or more recent. Additionally, we cannot rule out that the absolutely V_{sh} (tangential) wave speeds are over-fitted and a more reduced V_{sv} (vertical) $>$ V_{sh} (tangential) anisotropy representative of compositional eclogite is equally possible.

The MP also exhibits interesting V_{sv} and V_{sh} wave speed anomalies. The V_{sv} wave speed is faster in the South (>4.7 km/s) than the North (~ 4.3 km/s) with a well-defined transition between the two (Figure 1.5), particularly at depths <100 km. The V_{sh} wave speed of the MP, on the other hand, is more uniform at ~ 4.3 km/s but our resolving power based on our checkerboard tests (Figure 1.3) these values may be unreliable and subject to over-interpretation. The checkerboard resolution over the MP does recover the polarity of anomalies but the magnitudes are severely diminished. If we accept these observations, they show that the MP is isotropic on the northern half between 30 and 100 km depth and V_{sv} (vertical) $>$ V_{sh} (tangential) anisotropic to the South comparable to that of the central portion of the OJP indicative of a similar formation. We can more definitely say that based on the poor V_{sh} (tangential) resolution on the East side of the model, that the V_{sh} (tangential) wave speeds are relatively faster in the South compared to the North but we cannot postulate anisotropy due to the uncertainties in this part of the V_{sh} (tangential) model.

The possibility that the MP and the OJP were once a single large plateau with the HP is a limiting factor to any of the formation hypotheses. Any OJP or greater Ontong

Java Nui (Chandler et al., 2012; Hochmuth et al., 2015) formation must fit several criteria from this study, previous seismic studies (Richardson et al., 2000; Covellone et al., 2015; Tharimena et al., 2016; Tonegawa et al., 2019; Stern et al., 2020), geochemical analyses (e.g., Ishikawa et al., 2004, 2007; Korenaga 2005), geodynamical consideration and plate reconstructions (Hochmuth et al., 2015). The formation of the LIP, or its current tectonic structure, requires a fast V_{sv} (vertical) wave speed in the upper mantle under the central OJP (Covellone et al., 2015) surrounded by tangential V_{sv} (vertical) < V_{sh} (tangential) anisotropic peridotite. A slower V_{sv} (Richardson et al., 2000; Tharimena et al., 2016) and radial anisotropic anomaly below this, to depths of ~250-300 km, is also required. Whether these seismic anomalies were generated during the formation of the OJP or more recently is discussed below.

Any formation hypothesis must take into consideration the location of the OJP and MP at ca. 120 Ma when it was near the proposed Louisville hotspot track and Tongareva triple junction (Larson, 1997; Billen and Stock, 2000; Viso et al., 2005; Taylor, 2006, Hochmuth et al., 2015). Additionally, a formation hypothesis must take into account the geodynamics of an isostatically compensated buoyant (Korenaga, 2005) overly thickened oceanic crust. It must take into account a large volume erupting almost entirely subaqueous except for evidence from an ODP 192 drill core along the OJP eastern salient which showed evidence of subaerial volcanic activity (Thordarson 2004; Svensen et al., 2019). There may certainly be more evidence of subaerial eruptivity but the number of well logs and dredging locations are sparse. A lack of subaerial eruptivity, or evidence of, does not necessarily mean the plateau was

not above water during formation. The OJP did not systematically enlarge equally over time and it is quite possible that various portions were elevated while subaerial volcanism and plateau building was occurring elsewhere. Additionally, the plateau did not grow only from extrusive volcanism. Crustal growth from below or within the OJP could push parts of the plateau above the ocean surface which had previously erupted subaqueously. Any proposed formation of the OJP must account for geochemical and petrological evidence like a CMB signature from siderophile enrichment and garnet clinopyroxenite (eclogite) xenoliths observed in thrust OJP crust in the Solomon Islands as well high percentage of melt potentially associated with eclogite from recycled crust proposed in the upper mantle (Ely and Neal, 2003; Ishikawa et al., 2004, 2007; Korenaga 2005). Lastly, the observed variable overthickened crust across the OJP (Gladchenko et al., 1997; Tonegawa et al., 2019) must be considered as it is the end result of the formation. In the following section, we will discuss the three formation hypotheses already introduced, 1) plume, 2) fast spreading center, and 3) a plume partially entrained in a spreading center. Recent mechanisms like convective removal and viscous drainage that could replicate the above features will additionally be discussed (Lee et al., 2011). Some proposed formations or mechanisms like a bolide impact (Korenaga et al., 2005), imbricated oceanic material, or basal traction (Lee et al., 2011) will not be discussed as they are lacking specific evidence in for of these specific features.

Formation: Plume

A high-temperature heterogeneous plume head generates the required melt at the right time and location to produce the OJP and fits many of the criteria for its origin.

An entrainment of eclogite from subducted recycled crust (Ishikawa et al., 2004, 2007) would produce fast V_{sv} wave speeds beneath the OJP and V_{sv} (vertical) $>$ V_{sh} (tangential) anisotropy indicates a vertical alignment of the material. Several authors have shown the importance of eclogite to the formation of the OJP (Korenaga, 2005; Ishikawa et al., 2007; Covellone et al., 2015). Seismically, it is necessary for the fast wave speeds but not absolutely necessary to explain the $V_{sv} > V_{sh}$. Eclogite is only weakly anisotropic (Bascou et al., 2001) and while we lack confidence in the absolute wave speed of the V_{sh} (tangential) model due to smearing and lower data quality (Figure 1.3) it would still be weakly anisotropic. A garnet enriched eclogite like that found as xenoliths in the OJP outcrops in the Solomon Islands (Ishikawa et al., 2007) and proposed by Covellone et al. (2015) is also only weakly anisotropic due to the isotropic garnet are randomly orientated with the more anisotropic clinopyroxene (Bascou et al., 2001). Additionally, eclogite has been invoked as a method to limit uplift due to an impinging plume (Sobolev et al., 2011); though ODP Leg 192 expedition to the OJP found evidence of volcanoclastic sediments with carbonized wood remains associated with isolated subaerial eruptions (Thordarson 2004; Svensen et al., 2019). While Druken et al., 2013 and Stern et al. (2020) propose a frozen-in radial pooling of the plume head at the time of formation, the $V_{sv} > V_{sh}$ would not be indicative of an active nor frozen-in plume tail. 120 Ma of ambient mantle flow and subsequent collision with the Solomon Island subduction zone should have overprinted any mantle anisotropic signature. Due to the significant volume of partial melting required to generate the OJP (e.g., Korenaga, 2005), much of that upper mantle anisotropy would have been dissipated as the melt volume erupted eventually

replaced by ambient depleted mantle. Additionally, we only observe a vertical V_{sv} (vertical) $>$ V_{sh} (tangential) upwelling over active plumes and mid-ocean ridges (Becker et al., 2008). The more likely explanation for the anisotropy is downward dripping, foundering, or delaminating of the entrained eclogite (Lee et al., 2011; Covellone et al., 2015). This would additionally aid in explaining the resulting continuing buoyancy of the OJP. The 4.2 km/s lower wave speed anomaly (Figure 1.6) beneath is the resulting melt depleted and dehydrated lower density keel from the now passed mantle plume (Richardson et al., 2000; Klosko et al., 2001; Covellone et al., 2015; Tharimena et al., 2016). The OJP resists subduction at the Solomon Island trench due to its crustal thickness and buoyancy from the underlying depleted mantle keel and delamination of dense eclogite (Mann and Taira, 2004; Miura et al., 2004). A plume source can explain the geochemistry and the observed seismic tomography but cannot fully explain the crust. A plume head source erupting in the ocean would hypothetically generate crust of uniform thickness if the plume head were chemically homogeneous but this is not the case at the OJP where Gladczenko et al. (1997) and Tonegawa et al. (2019) observe an asymmetric crust across the OJP. Additionally, the lack of subaerial eruption is not a limiting factor to this hypothesis. It is more than plausible that the limited samples taken are not representative of the surficial eruptive history or that most of the surface of the OJP was subaqueously erupted but most of the crustal thickness was intruded.

Formation: Fast spreading center

A fast-spreading triple junction like the Tongareva triple junction is in the correct position and could theoretically generate the required melt to generate the OJP.

Though this hypothesis accounts for general lack of surficial uplift at the onset of eruption (Korenaga, 2005), it cannot account for many of the seismological features. At an active spreading center, one expects $V_{sv} > V_{sh}$ like that of the current East Pacific rise while radial V_{sv} (vertical) $< V_{sh}$ (tangential) anisotropy would be stronger in old oceanic lithosphere (Becker et al., 2008). As Covellone et al. (2015) and this study have observed, a fast V_{sv} in the shallow lithosphere is incompatible with a 120 Ma spreading center. One would expect azimuthal anisotropy frozen-in from the spreading or radial anisotropy from mantle shear (Becker et al., 2008) but azimuthal anisotropy is weak around the OJP between 50-400 km (Tharimena et al., 2016). Additionally, a slow V_{sv} wave speed root is not analogous with an ancient spreading center. Lastly, a typical spreading center, ignoring any influence from a nearby hotspot, would generate crust of uniform thickness as the OJP and other LIPs were generated between 6 and 14 Ma which is too short of a time scale to generate significant crustal differences due to mantle potential temperature (Dalton et al., 2014). To generate the observed crustal thicknesses at OJP a change in mantle temperature or mantle composition is required due to the large and variable crustal thickness; the mantle temperature and composition variations would be on the high end which suggests influence from a plume source (Dalton et al., 2014). Korenaga (2005) required a heterogeneous mantle with 100% melting of 25% eclogite in a heterogeneous mantle to generate a calculated uniform crustal thickness of 24 km across the OJP. Recent works have calculated variable crustal thicknesses of the OJP between 20km near the edges and up to 40 km in the thickest region (Gladchenko et al., 1997; Tonegawa et al., 2019). The percent of partial melting at a spreading center

to generate these crustal thicknesses is unlikely would far exceed those calculated by Korenaga (2005).

Formation: Plume partially entrained in a spreading center

The formation hypothesis of a heterogeneous plume head being partially entrained in a spreading center for a portion of its duration fits the supporting observations for each of the individual hypotheses while negating the drawbacks (Gladchenko et al., 1997; Tonegawa et al., 2019). An entrained plume fits the seismic observations where we found V_{sv} (vertical) $>$ V_{sh} (tangential) surrounded by V_{sv} (vertical) $<$ V_{sh} (tangential) anisotropy in the upper mantle representative of a delaminating or foundering dense eclogite above a melt-depleted keel. Additionally, a plume, entrained, at least partially, eliminates the subaerial eruption criticism (Korenaga (2005) by allowing that excess flexure to be absorbed by the spreading center (Gladchenko et al., 1997). Druken et al. (2013) has shown that the strain at a spreading center can shear a plume head as it approaches. Lastly, while a spreading center and plume would hypothetically generate fairly even crustal thicknesses, a partially entrained plume would be expected to generate asymmetric crustal thickness similar to what has been observed.

Convective removal and Viscous drainage

A mantle with V_{sv} faster than V_{sh} does not require the anisotropy to be upwelling like at plume or spreading center but can also represent downwelling due to convective removal from delamination, foundering, basal traction, or viscous drainage (Lee et al., 2011). The inferred eclogite in the upper mantle of the central OJP is denser than ambient peridotite and the proposed melt-depleted peridotite keel.

Delamination via density driven forces, where the previously entrained dense and seismically fast eclogite delaminates or peels from the thickened lower crust of the OJP could explain the observations in the upper mantle; additional analysis and modeling would be required to constrain this as a proposed phenomenon. Another density driven method to generate lithospheric sinking, similar to delamination, would be foundering, where the dense lower crust breaks away and sinks (Lee et al., 2011). Basal traction requires a thinned asthenosphere associated with a thickened lithosphere with a low viscosity beneath resulting in basal shear stresses driving removal but that has not been noted for this region or in any of our tomographic observations and is primarily observed beneath continents. Lastly, viscous drainage, involving the garnet-rich eclogite could “drain” into the asthenosphere due to their high densities and low viscosities (Lee et al., 2011). This process has only been proposed for continents but the OJP has been proposed as a proto-continent and mechanisms by which this could occur in oceanic plates would be similar. Korenaga (2005) and Covellone et al. (2015) have proposed downwelling to explain the OJP. Korenaga (2005) suggested delamination as early as 90 Ma to account for late-stage volcanism and Covellone et al. (2015) explained the atypical buoyancy of the OJP by downwelling eclogite associated with an observed seismically fast anomaly similar to this study.

The area North and northeast of the OJP centered at $\sim 5^{\circ}\text{N}$, 170°E exhibits V_{sv} (vertical) $>$ V_{sh} (tangential) persistently between at depths below the crust (Figure 1.5 and 1.7). This could additionally be representative of sinking lithosphere or, as previously stated concerning the resolution of the V_{sh} (tangential), it may be a product of poor resolution along the edges of the model. Due to the size of the anomaly, it

appears to be real though the magnitude of the V_{sv} (vertical) $> V_{sh}$ (tangential) should be treated with caution. Delamination and foundering would be atypical for this region. Though the oceanic crust in this region is older than 100 Ma, the area around the anomaly is not known for over-thickened crust or being buoyantly heterogeneous indicative of density-driven delamination or foundering. If the lithosphere was being removed, one would expect a rebounding observed in the crust which is also not observed in the seafloor topography. With the OJP having stopped and partially reversed subduction at the Solomon Island trench (Taira et al., 2004; Hochmuth et al., 2015), a transfer of stress could hypothetically generate the onset of a subduction zone jump from the South to the North. Studying this hypothesis would require additional analysis of the region like additional imaging and the locating of earthquakes, specifically smaller earthquakes (Hicks et al., 2017) as the region lacks the seismometer density required to observe these events. Interpreting and postulating the causes of or reasons behind these V_{sh} (tangential) anisotropic observations is extraordinarily difficult. It is more so in the Pacific where the V_{sh} (tangential) anisotropy variations are large and do not correlate well with sea floor age (Ekström and Dziewonski, 1998, Becker et al., 2008).

CONCLUSIONS

We used vertical and tangential waveforms from a combination of ambient noise and earthquakes to resolve the V_{sv} (vertical) and V_{sh} (tangential) seismic wave speed of the Ontong Java Plateau, the Manihiki Plateau, and the surrounding southwest Pacific. The vertical V_{sv} model's resolution is a further improvement over previous

research with substantial progress in our model relative to the initial 1D starting model and number of measurements used through iterations (Figure 1.4). The tangential Vsh model is the first using the full waveform scattering integral technique and first for this region at this scale.

We observed regions of Vsv wave speeds >4.7 km/s beneath the central part of the OJP surrounded by slower wave speeds similar to that of Covellone et al. (2015) and beneath the Manihiki plateau at similar depths. We observed slower tangential wave speeds (4.1-4.2 km/s) at the same plateau locations as these fast Vsv anomalies but $V_{sv} < V_{sh}$ wave speeds in the surrounding plateau regions. We additionally observed a fast (>4.7 km/s) $V_{sv} > V_{sh}$ anomaly north-northwest of the Ontong Java Plateau.

The $V_{sv} > V_{sh}$ wave speed anomalies beneath the Ontong Java Plateau and Manihiki Plateau are consistent with delamination or foundering of eclogite from a heterogeneous plume source (Tejada et al., 1996, Neal et al., 1997; Michael, 1999; Ely and Neal, 2003) which was partially entrained at a spreading center (Gladczenko et al., 1997; Dordevic and Georgen, 2016). This interpretation is consistent with an entrained eclogite proposed by Covellone et al. (2015), crustal thickness estimates, and lack of substantial subaerial eruptions. The observed $V_{sv} > V_{sh}$ to the North is indicative of lithospheric downwelling associated with downwelling, a subduction zone jump, or could be anomalous Vsv (vertical) $>$ Vsh (tangential) anisotropy in old (>80 Ma) oceanic lithosphere but this is speculative.

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Iteration	Data Type	dT Threshold (seconds)	Damping	Smoothing
1	AN long period	35	11	11
2	AN long period	35	14	20
3	AN long period	35	20	20
4	AN	35	18	18
5	AN	35	18	18
6	AN	35	16	16
7	AN and EQ	35	90	90
8	AN and EQ	35	60	60

Table 1.1: Vsv (vertical) summary of inversion parameters

Iteration	Data Type	dT Threshold (seconds)	Damping	Smoothing
1	AN long period	35	10	10
2	AN long period	35	12	12
3	AN	35	16	16
4	AN	35	14	14
5	AN and EQ	35	60	60

Table 1.2: Vsh (tangential) summary of inversion parameters

Figure 1.1: 51 Seismic stations (green inverted triangles) and 114 earthquake events (red circles) used in this study in the Pacific Ocean. The computational model is outlined in red. The black arrows show modern Pacific plate motion. Seismic stations, earthquake events, and computational model are the same for the vertical and tangential models. The locations of Japan (JP), Hawaii (HI), Australia (AUS), the Ontong Java Plateau (OJP), and Manahiki Plateau (MP) have been labeled in black for reference.

Figure 1.2: Computational model domain perturbed with 3.5 and 5 sized harmonic positive and negative 5% wave speed anomalies. Our vertical recovered solution is plotted at three depths into the model. The region with high quality recovered resolution is outlined in red.

Figure 1.3: Computational model domain perturbed with 5 and 10 sized harmonic positive and negative 5% wave speed anomalies. Our tangential recovered solution is plotted at three depths into the model.

Figure 1.4. (a,c) The change in the dT through iterations. (b,d) Number of measurements through iteration. Black circles are from ambient noise data, red circles are from earthquake.

Figure 1.5: (a-f) Depth slices through the best resolved region of the Vsv (vertical) model (figure 1.2). Color scale is absolute Vs in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

Figure 1.5: continued

Figure 1.6: V_{sv} (vertical) cross-section depth profiles through areas of interest around the Ontong Java Plateau. (a) North-South cross-section at longitude 160deg between latitude 15S and 15N. (b) North-South cross-section at longitude 157deg between latitude 15S and 15N. (c) East-West cross-section at latitude -3.0deg between longitude 150W and 180W. (d) East-West cross-section at latitude -5.0deg between longitude 150W and 180W. The inset map from figure 1.5c shows the location of the cross-sections in relation to the Ontong Java Plateau. The depth range for each cross-section is 25-350km depth. Color scale is absolute V_s in km/s

Figure 1.7: (a-f) Depth slices through the best resolved region of the Vsh (tangential) model (figure 1.3). Color scale is absolute Vs in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

Figure 1.7: continued

Figure 1.8: V_{sh} (Tangential) cross-section depth profiles through areas of interest around the Ontong Java Plateau. (a) North-South cross-section at longitude 157deg between latitude 15S and 15N. (b) East-West cross-section at latitude -3.0deg between longitude 150W and 180W. The inset map from figure 1.7c shows the location of the cross-sections in relation to the Ontong Java Plateau. The depth range for each cross-section is 25-350km depth. Color scale is absolute V_s in km/s

APPENDIX

Figure 1.9. 19 and 28 km depth slices through the best resolved region of the vertical model (figure 1.2). Color scale is absolute V_s in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

Figure 1.10. 19 and 28 km depth slices through the best resolved region of the tangential model (figure 1.3). Color scale is absolute V_s in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

Figure 1.9: Depth slices through the best resolved region of the Vsv (vertical) model (figure 1.2). Color scale is absolute Vs in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

Figure 1.10: Depth slices through the best resolved region of the Vsh (tangential) model (figure 1.3). Color scale is absolute Vs in km/s. Thick black lines denote the Ontong Java Plateau and Manihiki Plateau, thin black lines are coastlines, and thin gray lines are plate boundaries.

CHAPTER 2

Pn analysis of the crust-mantle interface of the Ontong Java Plateau

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ABSTRACT

The Ontong Java plateau (OJP) took shape from a complicated, but poorly understood geological history in the southwest Pacific. Unraveling the formation and deformation of this LIP is not straightforward due to limited available data, remote location, and atypical geology. Analyzing the crustal structure of the OJP will help us understand its origin: melting of a plume head or a fast-spreading triple junction.

We perform an analysis of first arrivals of P and Pn waves that travel along the OJP crust-mantle interface utilizing synthetic seismograms to constrain the crustal thickness and velocity structure along cross-sections from East to West. Using the FK integration method, a multitude of 1D crust and uppermost mantle models were used to compare P/Pn arrivals from synthetic seismograms and earthquake wave forms.

Sediment and crustal thicknesses and wave speeds from this method are consistent with many previous results. Crustal thicknesses were ~20km thick along the East and West edges and thickened to 40 km in the West central OJP with a steady ~6 km/s mid-crustal and 8.4 km/s uppermost upper mantle wave speed. These fast uppermost upper mantle wave speeds, crustal thicknesses are consistent with an asymmetric plume source interacting with a ridge.

INTRODUCTION

The Ontong Java Plateau (OJP) (Figure 1) initially formed approximately 120-125 Ma with an estimated erupted magma volume of 44 to 57 Mkm³ over 6 – 14 My making it the largest known Large Igneous Province (LIP) (Coffin and Edholm, 1994; Gladchenko et al., 1997; Tejada et al., 2002). The primary hypotheses for the

formation of OJP are a 1) plume head from the initiation of a hotspot (Griffiths et al., 1989; Campbell, 1998) or 2) a fast-spreading triple junction (Larson, 1997; Billen and Stock, 2000; Viso et al., 2005; Taylor, 2006), while a hybrid hypothesis involves a combination of a plume at a triple junction (Gladczenko et al., 1997; Dordevic and Georgen, 2016; Chapter 1 of this study). The plume source hypothesis would generate a buoyant Rayleigh-Taylor anomaly (Griffiths et al., 1989; Campbell, 1998) as a large plume head with a smaller tail extending to the core-mantle boundary, as determined from geochemical analysis (Tejada et al., 1996; Michael, 1999; Neal et al., 1997; Ely and Neal, 2003) resulting in a significant amount of initial melting represented as a LIP followed by a hotspot track of ocean islands (Neal et al., 1997; Chandler et al., 2012). In this instance, the hotspot track has been associated with the Louisville ridge (Neal et al., 1997; Chandler et al., 2012). A passive, fast spreading rift hypothesis (Larson, 1997) would generate upwelling driven by very fast spreading of a half rate of 7.7 cm/yr at the Tongareva triple junction represented by the Pacific, Phoenix, and Farrallon plates in the area of OJP at the time of emplacement (Larson, 1997; Billen and Stock, 2000; Viso et al., 2005; Taylor, 2006). Both primary hypotheses have unresolved discrepancies that have not been adequately addressed. Given a plume source, the emplacement volume at OJP, MP, and HP would generate significant lithospheric extension and would uplift the crust above sea level (Farnetani and Richards, 1994; Korenaga 2005), though only subaqueous rock samples have been observed in dredges and drill cores (Mahoney et al., 2001; Kroenke et al., 1991; Winterer et al., 1971; Packham et al., 1975) except for one instance on the eastern salient (Thordarson, 2004). The passive, fast-spreading rift hypothesis suffers from a

volume problem requiring 100% melting of eclogite from a 25% thermally equilibrated eclogite source to generate the necessary quantities (Korenaga, 2005). An entrained plume in a fast-spreading triple junction utilizes the best of each hypothesis above while eliminating or compensating for their deficiencies (Gladchenko et al., 1997, Dordevic and Georgen, 2016). A plume, partially or temporarily, entrained in a triple junction would still have the same geochemical origin and generate the necessary melt fraction while negating any requirement of significant uplift to accommodate the upwelling.

After emplacement of the Ontong Java Nui, made up of OJP, MP, and HP, breakup soon began. At 120-118 Ma, MP began breaking off from HP along its southern margin (Billen and Stock, 2000; Worthington et al., 2006) and the OJP separated to the West (Chandler et al., 2012). The OJP separation with the MP terminated at approximately 80 Ma. As the OJP continued passively north-northwest drifting on the Pacific plate (Hochmuch et al 2015), episodic volcanism occurred at approximately 90, 60, 40, and 22 Ma (Mahoney et al., 2001) while slowly accruing over 1km of pelagic sediment in places (Musgrave, 1990). At 25-22 Ma the OJP struck the subduction zone at the Solomon arc resulting in arc volcanism cessation from 20-15 Ma and plate re-arrangement (Kroenke et al., 1986, Yan and Kroenke, 1993, Musgrave, 1990) due to most of the OJP being unsubductable (Hochmuch et al., 2015) except for the removal of small segment of the lower OJP (Mann et al., 1996, Petterson et al., 1997; Petterson et al., 1999). A subduction polarity reversal with the Australian plate at 12-6 Ma resulted in significant deformation and the formation of the New Britain and San Cristobal trench and uplift at 4-2 Ma of the Malaita

anticlinorium and Solomon Islands overthrust (Petterson, 1995; Petterson et al., 1999; Musgrave 1990). Additionally, during this time, the Solomon Island arc is transferred from the Australian plate to the Pacific plate (Musgrave, 1990). The OJP sits in its current location with its southern terminus at the North Solomon Trench (Cooper and Taylor, 1985), the Caroline Seamounts, Lyra and East Mariana basins to the northwest, and Nauru basin to the northeast (Gladczenko et al., 1997; Keating et al., 1984).

Several drilling, dredging, seismic, and geochemical studies have been conducted to look at the OJP and its underlying structure. Gomer and Okal, (2003) and Suetsugu et al. (2019) utilized ScS waves to study the attenuation structure and found a higher than average Q in the upper mantle, Tharimena et al. (2016) used receiver functions and observed mid-lithospheric discontinuities at 80 and 282 km depth, Klosko et al. (2001) and Richardson et al. (2000) modeled a large low wave speed root using SkS splitting and earthquake tomography, respectively, while Covellone et al. (2015) observed a fast V_{sv} wave speed anomaly in the upper mantle using full waveform tomography. These authors focused on the upper mantle under and around the OJP. Other authors used gravity and magnetics to study the OJP crust and upper mantle and observed an anomalous buoyant and thickened crust (Rose et al., 1968; Sandwell and Renkin, 1988; Schubert and Sandwell, 1989; Nakanishi et al., 1992; Gladczenko et al., 1997; Ito and Taira, 2000). Additionally, Furumoto et al. (1970); Furumoto et al. (1976), and Hussong et al. (1979) investigated the crust and crust-mantle transition using active source refraction techniques but focused mainly on the West and central OJP. They observed thickened crust but only Furumoto et al. (1976) observed a fast

lower crust. Besides the biased refraction coverage on the western edge, the surveys are spread too wide apart or short in length to account for potential crustal variations. The East portion of the OJP is especially devoid of crustal investigations. More recently, Tonegawa et al. (2019) used an OBS deployment around the OJP to study Receiver Functions (RF) of the crust and mantle modeling a thickened crust in the middle and a thinner but still thick crust along the East and West edges and Stern et al. (2020) investigated the P-wave speed of the MP and HP with active and passive seismic methods and observed extremely fast (8.7-8.8 km/s) uppermost upper mantle.

Compared to the seismic crustal studies, drilling and dredging have sampled the OJP more broadly. Five Offshore Drilling Program (ODP) expeditions have cored the sediments and crust of OJP. DSDP legs 7, 30, and 89 each penetrated the OJP once, only sites 289 and 586, which are located adjacent to each other, penetrated past the sediments and into basaltic basement (Winterer et al., 1971; Packham et al., 1975; Moberly et al., 1986). Additionally, ODP 192 drilled 6 total sites, 5 of which pierced OJP basement, and ODP 130 drilled 5 sites, two of which penetrated basement (Kroenke et al., 1991; Mahoney et al., 2001). Outcrops of the OJP have also been sampled along the Solomon Islands on Malaita, Santa Isabel, San Cristobal and sections on Ramos and Ulawa (Tejada et al., 2002). Two distinct compositions with no evidence of mixing were observed, neither of typical midocean ridge basalts. The authors hypothesized that the samples represented a geochemically heterogeneous peridotite and eclogite plume head (Tejada et al., 2002).

There is still ongoing debate regarding the crust and crust-mantle transition of the OJP. The West and East OJP boundaries are poorly sampled, and there are

discrepancies between existing geophysical datasets and crustal thickness estimates where supplementary information could provide insight. We analyzed and modeled Pn body waves that traverse the OJP and refract along the crust mantle interface to investigate crustal structure, thickness and mantle wave speeds. This analysis could help elucidate the true crustal boundaries and aid in the volumetric calculations of the OJP utilizing the variations of the crust-mantle depth. Lastly, these results will allow future studies to discriminate how OJP thickness relates to melt volume, its origins, how OJP may have deformed along its journey to its current location and how it relates to the HP and MP.

Pn waves are seismic body waves that refract along the crust-mantle interface (Moho). The Pn waveforms at large epicentral distances, like those in this study, sample increasingly below the Moho, which is seismically faster than the crust. The crustal thickness can be interpreted based on when the Pn waveforms arrive, and thicker crust produces a larger travel time delay. The following is a description of our crust and uppermost upper mantle forward modeling methodology focusing on Pn arrivals utilizing earthquakes and stations around the OJP. We use the Pn arrivals to determine the crust-mantle structure of OJP utilizing Pn waves to investigate its deformation history from eruption to the current location at the Solomon Island OJP convergent zone.

METHODOLOGY

The remote location of the OJP, the sparseness of seismic sensors and the bias of earthquake events to the South does not allow for many geophysical techniques to

investigate the crust and uppermost upper mantle. To model and predict the thickness and wave speeds of the sediment, crust, and uppermost mantle, we forward model using a frequency-wavenumber (FK) scheme in pseudo two-dimensions (Zhu and Rivera, 2002). A one-dimensional model is derived from existing local and regional wave speed data, which is then used to derive Greens Functions (GF) for the model space. Synthetic waveforms are generated from the GFs in combination with event and station information. These synthetic waveforms are generated for a multitude of sediment and crustal thicknesses and wave speeds which are then compared to the real event-station data. These time differentials are used to determine the best-fitting crustal thickness and wave speed model for locations along an East-West transect across the OJP.

Data Preparation

Earthquake-station pairs, to be used in the analysis, are initially selected if both the event and station are in between the latitude of -15° and 6° and longitude 140° and 180° which corresponds to area over the OJP (Figure 2.1). Additionally, the event criteria include, having a known CMT solution (Dziewonski et al., 1981; Ekström et al., 2012) between Mb 5 and 7.5, from 1990-2017 with a focal depth less than 400 km. The seismic station criteria must be a broadband (BH) sensor and less than 15 degrees epicentral distance to avoid obscuring the Pn arrival from other arrivals due to the upper mantle triplication. Additionally, we required the event-station path to cross or be adjacent to the OJP.

From this initial grouping of waveforms with the instrument response removed, consisting of 10 minutes prior to the theoretical P arrival and 60 minutes after, we

selected only waveforms on the vertical component with a signal to noise ratio (SNR) greater than five based on amplitude of noise prior to the P arrival vs the amplitude around the P arrival. We used only the vertical component as Pn waveforms are nearly identical on the vertical and radial component while the vertical component has higher SNR (Savage et al., 2003). This resulted in 96 events with 117 event-station pairs with 7 unique stations, only one of which is to the South-Southeast of the OJP. We divided this group of event-station pairs by location, ray-path, and Pn waveform similarity. To be similar, the Pn waveforms were required to have the same amplitude polarity, shape, and relative strength within a group and depth range defined as 0-20, 20-40, 40-60, 60-100, and greater than 100 km depth (Savage et al., 2003). The grouping resulted in a total of 7 locations (Figure 2.1). Within these locations, 39 representative waveforms (Table 2.1) were selected with the best SNR for each depth range to further reduce the computational expense of forward modeling.

As the theoretically calculated P/Pn arrival is based on a simplified wave speed model, an updated P/Pn arrival was hand-picked. A new P/Pn arrival search started at ± 5 seconds from the theoretical arrival, resulting in 34 P/Pn arrivals were within this range. Of the 5 observed outside this time range, 4 were within an additional 1.5 seconds and 1 was observed well outside the normal range most likely due to misapplication of a theoretical P/Pn arrival (Table 2.1).

1D Model Generation

Before a forward model can be calculated, one-dimensional models were generated. One-dimensional models were used since the source-receiver ray paths in this study only travelled in one direction (~South to North) thus the seismic refraction

reciprocal, which requires ray paths in the opposite direction to determine variable thickness and dip was impractical. Existing regional and local wave speed models were used as starting points. The forward modeling required a minimum of 4 inputs consisting of depth in km, density, P-wave speed, and S-wave speed. Where density was not defined, but calculated using

$$\rho = 0.77 + 0.32V_p \quad (2.1)$$

where ρ is density and V_p is P-wave speed. Where P-wave speed or S-wave speed is not defined a V_p/V_s ratio from the layer above and below was used. Additionally, slowness (Q_s and Q_p) was treated as 500 for Q_s and $2 \times Q_s$ for Q_p as a default; this did not factor into the analysis or calculation since we were concerned only with the P/Pn arrival time not its relative amplitude. The one-dimensional models were subsequently changed to allow for the testing of the sediment and crustal thickness thru the modification of the depth component. Additionally, where necessary, the uppermost upper mantle wave speed was substituted with refraction wave speeds from previous studies.

Synthetic Waveform Generation

Synthetic seismograms were calculated thru a two-step process utilizing the frequency-wavenumber (FK) integration method (Zhu and Rivera, 2002; Wang and Hermann, 1980). This method uses the Thomson-Haskell propagation matrix (Thomson, 1950; Haskell, 1953; Kennett, 1973), also referred to as the Haskell matrix to relate displacement-stress vectors at the top and bottom in a multi-layered half-space (Haskell, 1964; Wang and Hermann, 1980; Zhu and Rivera, 2002)

$$\mathbf{a}_n = \mathbf{E}_n \mathbf{\Lambda}_n (z_n - z_{n-1}) \mathbf{E}_n^{-1} = \mathbf{E}_n \mathbf{\Lambda}_n (d_n) \mathbf{E}_n^{-1} \quad (2.2)$$

where $\mathbf{E}_n$ is a similarity matrix, $\mathbf{\Lambda}_n$ is a displacement-stress term, and d_n is the layer thickness. The Haskell matrix assumes boundary conditions with continuity of displacement and stress across the interfaces with stress disappearing at the free surface and waves vanishing in the half-space (Zhu and Rivera, 2002). In the first step, the displacements are computed for the one-dimensional layered model for a point source using wave number integration summation. This generates a series of greens functions for the vertical, radial, and transverse (clockwise) components which provide the general horizontal radiation patterns. The vertical radiation pattern for a double-couple displacement is

$$\begin{aligned}
u_z = & \frac{1}{2} \sin 2\delta \sin \lambda u_z^0 \\
& -(\sin \phi \cos 2\delta \sin \lambda - \cos \phi \cos \delta \cos \lambda) u_z^1 \\
& -(\sin 2\phi \sin \delta \cos \lambda + \frac{1}{2} \cos 2\phi \sin 2\delta \sin \lambda) u_z^2
\end{aligned} \tag{2.3}$$

where δ is the fault plane dip angle, λ is the slip direction from the strike of the fault, and ϕ is the azimuth of the seismic station from the point source (modified from Wang and Hermann, 1980). The radiation pattern for the vertical component is a 45° down-dip slip at azimuth 45° for u^0 , vertical dip slip for u^1 at azimuth 45°, and a vertical strike slip at 22.5° at u^2 (Wang and Hermann, 1980). To generate the greens functions in the forward modeling, besides the one-dimensional model, it requires an event depth, event-station distance in degrees, sample rate, and number of total samples.

In the second and final step, synthetic waveforms are generated from the greens functions thru the addition of a source radiation pattern from the source time function

(STF) and other inputs via the FK integration described above, convolved with a triangle STF. The STF was calculated from the event's magnitude from the GCMT catalog (Dziewonski et al., 1981; Ekström et al., 2012) using

$$L = 10^{\frac{M_W - a}{b}} \quad (2.4)$$

where L is the surface rupture length, a is 5.08, b is 1.16 from Table 2A in Wells and Coppersmith (1994) and Covellone et al. (2015). The rupture time is defined as

$$T_R = \frac{L}{V_R} \quad (2.5)$$

where T_R is the rupture time and V_R is the wave speed defined as 2.86 km/s. Besides the STF which includes the event magnitude, the other inputs required to generate the synthetic waveforms are the six moment-tensor elements and the station azimuth from the event. The waveforms are output as velocity and the time series is integrated to get displacement which can be directly compared to the real waveforms.

P/Pn measurement

As part of the synthetic waveform generation, the Pn or first arriving P is known and identified in the synthetic seismograms. These were visually inspected to verify accuracy which is easier as the noiseless waveforms do not start at event initiation but several seconds prior to the first arriving body wave (P/Pn). The difference between the synthetic and real P/Pn arrivals was calculated using

$$\delta P = P_s - P_r \quad (2.6)$$

where P_r is the hand-picked P/Pn arrival from the real waveforms and P_s is the synthetic arrival. One dimensional Earth models that accurately match the Pn waveforms will have a small δP and those matches are used to construct a pseudo 2D

profile of the crustal mantle interface across the OJP with estimates of upper mantle velocity and crustal thickness.

RESULTS

A total of 5604 synthetic seismograms with unique event-station paths were generated using the FK method (Zhu and Rivera, 2002) for comparison against each regional location specific Pn arrival. The synthetic waveforms were generated from one of six starting 1D models (Figure 1.2). These six 1D starting models were chosen as they represented global datasets or were derived from local and/or Pacific datasets. The base starting models are GT1r5 (Nishimura and Forsyth, 1989) which represents 110+ Myr Pacific ocean, PA5 from Gaherty et al. (1996) developed for the Tonga region, PA2 from Lerner-Lam and Jordan (1987) generated from data mostly from 100 Ma Pacific oceanic lithosphere, PAC6 from Tan and Helmberger (2007) generated from pure paths across the Pacific ocean, PHB3 (Kato and Jordan, 1999) created from multiple S phases in the western Philippine Sea, and IASPI from Kennett and Engdahl (1991) which is a global 1D model.

Each 1D starting model was modified by thinning or thickening the sediment, crustal thickness and/or replacing the uppermost mantle wave speeds with values from regional refraction surveys (Furumoto et al., 1970; Furumoto et al., 1976, Gladczenko et al., 1997) and ambient noise dispersion curves (Ku et al., 2018). The sediment thickness was varied between 0, 1, 1.25, 1.5 and 2 km based on results from refraction measurements (Furumoto et al., 1970; Furumoto et al., 1976, Gladczenko et al., 1997) and previous drill core results (e.g., Fitton et. al., 2004) when used in conjunction with

wave speed changes to the uppermost mantle from refraction surveys. All other base models used the original sediment thicknesses (GT1r5 = 1.0 and 1.3 km, PA5 = 0.2 and 1 km, PA2 = 1 km, PAC06 = 0.2 km, PHB3 = 0.1 km, IASP91 = 1 km). Sediment thickness of PA5 was additionally tested using 0, 1.3 and 1.4 km which were derived from drilling and refraction analysis (Furumoto et al., 1970; Furumoto et al., 1976, Fitton et al., 2004). This resulted in the generation of 1209 Gt1r5 models, 1693 PA5 models, 730 PA2 models, 869 PAC6 models, 869 PHB3 models, and 234 IASP91 models. Synthetic Pn arrivals from these wave speed models were compared to the corresponding Pn arrival data. The thirty-nine events with Pn data which were selected as representative waveforms from the initial subset of over 100 events had an average event-station path of 13.4°.

Several sources of error exist for the measurement of Pn arrivals and the generation of synthetic Pn waveforms. Source depth errors may occur from the use of event origin information from the GCMT catalog (Dziewonski et al., 1981; Ekström et al., 2012) where event depths which were not independently calculated. The average difference between the theoretical Pn arrival calculated using IASP91 (Kennett and Engdahl, 1991) and hand-picked is 0.23 seconds with a standard deviation of 4.21. Sources of error associated with the synthetic waveforms are due to inaccurate moment tensors, source depth from the GCMT and the use of a 1D velocity model over the source-receiver path. Within most of the propagation paths, the region between where the earthquake and station is tectonically complicated including low-velocity basins, subduction zones, and the thickened crust of the OJP large igneous province. Using a simplified 1D velocity structure does not account for the real-world

complexity and introduces potential sources of measurement error. That being said, we are attempting to determine how the Earth models change across the study region. The resulting Pn arrival times from varying the crustal thickness (Figure 2.3) is significant enough to overcome any errors described above. On average, the Pn arrival time varies by 3 seconds per 10 km of crustal thickness (Figure 2.4).

Location 1: West of OJP

A total of 5 Pn event arrivals were used, and 548 synthetic waveforms generated (Figure 2.1, Table 2.1). All source-receiver paths for this region are West of the OJP and oriented primarily North-South. Four of the Pn events were each compared against 125 synthetic waveforms while the fifth event, 081795F (Table 2.1), was compared against 48 models due to its remote location away from known seismic refraction surveys and its unique source-receiver path. Compared to other Pn events for this location, the best fitting models for event 081795F has the thinnest crust at 5-10 km. When event 081795F is removed from the analysis, the preferred model has a slightly thickened crust. Each Pn arrival (event to station pair) preferred different models but the models shared similarities. A deep event, 042595B, and an event located off structure, 081795F, preferred simple, oceanic like, wave speed models, PAC06 and PA2 respectively. The remaining Pn arrivals require models modified by included refraction wave speeds (Figure 2.2). The overall best fitting model for all Pn arrivals is PAC06 with a 20km thick crust and 8.4km/s mantle wave speed. Best fitting models for each depth range prefer Pacific specific models without any refraction wave speed modifications (Table 2.2).

Location 2: Far West OJP

Loc. 2 is located on the far western edge of the OJP. Four Pn event-station pairs were used with each compared to 144 synthetic waveforms for a total of 576 generated seismograms (Table 2.1). Each Pn ray path is oriented South-North with approximately 40% of the ray path outside the OJP. All event depths are 15km or less. The best fitting model to each Pn arrival are Pacific and regional models (PAC06, PA5, and PHB3) modified by refraction wave speeds. The overall best fitting model for all far western OJP Pn arrivals is a modified model PAC06 with 2 km sediment, 30.5 km crust, and 8.4 km/s wave speed mantle.

Location 3: West OJP

Loc. 3 is located just to the East of Loc. 2 (Figure 2.1) with over 80% of the source-receiver path within the OJP. Eleven Pn event-station pairs (Table 2.1) were compared to 111-142 synthetic waveforms for a total of 1500 waveforms. The source depths of each western OJP events varied across a broad range of depths (Table 2.1). Each depth range preferred a different wave speed model, but all required between 1.25-2 km of sediment and crust greater than 30km. The shallowest depth range, which would be the most sensitive to crustal and Moho wave speeds, prefers a mantle wave speed of 8.4 km/s. The next shallowest source depth ranges (20-60 km) require an 8.4 km/s mantle wave speed, while source depths least sensitive to crustal wave speeds (>60 km depth) require a mantle wave speeds of 8.6 km/s. The overall best fitting model for all West OJP Pn arrivals is PA2 with 2 km of sediment, 30.5 km of crust, and 8.4 km/s wave speed uppermost mantle wave speed like that of shallower source depths.

Location 4: West-central OJP

Loc. 4 is located just slightly East of Loc. 3 (Figure 2.1) and like Loc. 3, most (greater than 90%) of the source-receiver ray path is within OJP. Eight Pn event-station pairs (Table 2.1) were each compared to 138 wave speed models for a total of 1101 waveforms. Source depths of events in this region vary across a broad depth range from shallow to > 100 km (Table 2.1). Wave speed models grouped by their source depths were divided into two populations. Events with deep source depths preferred a fast refraction wave speed of 8.6 km/s in the mantle and a thick total crust of 39.5 km. Shallower source depths preferred a 16km crust and simple GT1r5 wave speed model. Similarly, events with sources at 20-40 km depth required a simple PA5 wave speed model but the crust is unrealistically thin at 3 km and is not indicative of the overall location nor the source depth range. The best fitting wave speed models for the Pn arrivals from events with source depths between 20-40 km have crustal thicknesses 30 and 50 km; the mean of the preferred best fitting models for each are 32.32 and 23.14 km respectively. For the shallowest events, the best fitting wave speed models, for the individual Pn arrivals, have crustal thicknesses of 39.5, 16.3, and 20.3 km (appendix 2.1). The overall best fitting model for all West central OJP Pn arrivals is PAC06 with 0.2 km of sediment, 39.8 km of crust, and 8.43 km/s mantle wave speed.

Location 6: East-Central OJP

Loc. 6 is located between Loc. 4 to the West and Loc. 7 to the East with source-receiver paths crossing the east-central OJP; location 5 was absorbed into location 4 and 6. Four Pn event-station pairs (Table 2.1) were compared against a combined total of 634 synthetic waveforms. Source depths of east-central events varies across a

wide range (Table 2.1), but Pn arrivals require a simple Pacific derived wave speed model (PA2 and PAC06) with a mantle wave speed of 8.17 and 8.43 km/s respectively. The shallowest (0-20 km) and deepest (>100 km) depth range require a thicker crust, 23.25 and 39.5 km while the mid-source depth range prefer an anomalously and unreasonable thinner crust. The mean of the crustal thickness of the most preferred best-fitting models is 16.31, 5.92, 19, and 20km from shallowest to deepest depth ranges. The range of crustal thickness per event-station pair for the best fitting models were unusually large compared to the other locations (3-26.3km, 4-10km, 5-39.5km, 5-32.5km). Though individual best fitting models were varied, the overall best fitting model for all East-central Pn arrivals preferred a PA2 wave speed model with a 23.25 km thick crust and a slightly slower uppermost upper mantle wave speed of 8.17 km/s compared to locations to the West.

Location 7: East OJP

Loc. 7 is located on the eastern edge of OJP with 50% or more of the source-receiver ray path located off the OJP (Figure 2.1). Seven Pn event-station pairs (Table 2.1) were compared to a combined total of 1107 synthetic waveforms. Source depths of eastern OJP events are from shallow to > 100 km depth (Table 2.1) and they all prefer a Pacific derived wave speed model (PA2, PA5, and PAC06). The shallowest source depth range require 22.5 km thick crust with a refraction derived mantle wave speed of 8.4 km/s. The two deepest depth ranges, which are least sensitive to crustal wave speeds, require simpler structure and 5 km thin crust. The overall best fitting model for all eastern OJP Pn arrivals is GT1r5 with 1 km of sediment, 4 km of crust, and 8.17 km/s mantle wave speed. The best fitting model correlates with the depth

ranges that are least sensitive to crustal thickness but has the more arrivals. Here the deep Pn arrival results may have biased the best fitting model towards a thin crust as compared to the other locations where most depth range groups and/or the mean of the crustal thickness for each Pn event are similar to the best fitting model. Additionally, this region may be the most complicated region as the Pn raypaths are sampling different wave speed structures with the deep arrivals preferring an oceanic crust and shallow arrivals preferring a thickened OJP type crust. The most likely best fitting model is probably somewhere between one derived from this analysis of deep events (4 km thick crust and 8.1 km/s) and from the 0-20 km depth range results (22.5 km thick crust and 8.4 km/s); a thick crust and 8.4 km/s mantle wave speed is used as the preferred wave speed model.

DISCUSSION

Modeling of the synthetic Pn arrivals provided insight into the complicated crustal structure of the OJP. To the west, outside the bounds of the OJP, our analysis shows that the crust is estimated to be 20 km thick. On the western side of the OJP the crustal thickness is 32.5 km increasing to 40km, 23.25 km in the central portion and 22.5 km in the East (Figure 2.5). The overlying sediment also varies across the OJP. The best fitting model for each location exhibit thin 0.2 km thick sediment to the West of the OJP, 2 km thick sediment on the far western side of OJP thinning to 0.2 and then thickening again to 1.25km in the central part of OJP and 1km on the East (Figure 2.5).

Crustal thickness

A large range of overall crustal thicknesses were tested including varying the wave speed of the crustal layers (Figures 2.2 and 2.4). Several studies have been conducted looking at the crustal thicknesses of the surrounding OJP area including gravity and magnetics (Rose et al., 1968; Sandwell and Renkin, 1988; Schubert and Sandwell, 1989; Nakanishi et al., 1992; Gladchenko et al., 1997; Ito and Taira, 2000), active source reflection/refraction (Furumoto et al., 1970; Furumoto et al., 1976; Hussong et al., 1979), regional and local seismic tomography from earthquakes and ambient noise (Richardson et al., 2000; Covellone et al., 2015), OBS receiver functions (RF) (Tonegawa et al., 2019), and SS precursors (Tharimena et al., 2016). Many of these results correlate well with crustal thicknesses determined by this study. While normal oceanic crustal thickness, depending on age, is in the range of 5-10 km (e.g., White et al., 1992) with a thicker crust associated with greater age. Tonegawa et al. (2019) observed 10 and 13km thick crust to the West and East of the OJP utilizing RF which is slightly greater than normal 120Ma oceanic crust. To the South of OJP, near the Solomon Islands, crustal thicknesses range from 13-15km (Furumoto et al., 1970) from refraction measurements, 13-15km (Gladchenko et al., 1997) from joint gravity and refraction analysis, and 20.4 (Ku et al., 2018) from ambient noise. Though our study has no sensitivity to the South of OJP, the crustal thicknesses West of OJP are 20 km while the Eastern side of OJP is 22.5 km thick. These values are much greater than several of these previous studies but were calculated within 3-5° of the 4000m bathymetric outline commonly used to delineate the OJP boundary. The crustal thickness from our study to the West of OJP and Tonegawa et al. (2019) suggests that the OJP extends beyond this arbitrary boundary making the volume and

size of OJP even larger. The results for Loc 7. (East OJP) have approximately 50% of the ray path outside of the OJP bathymetric boundary with a preferred crustal thickness of 22.5 km making the measurement speculative.

Over the main portion of OJP, most studies found a range of thickened crust. Refraction studies found crustal thickness ranging from 19-25 km over south-central OJP to 39-42 km in the North (Furumoto et al., 1976) which when averaged correlates well with Location 2 and 3 of this study while the thicker crustal estimate of the North OJP results is similar to Location 4. Hussong et al. (1979) calculated an average crustal thickness of 37 km for OJP which is also similar the central OJP results for this study. A gravity study by Furumoto et al. (1970) found the crustal thickness to be greater than 20 km over OJP while Gladczenko et al. (1997) calculated the central OJP had 30 km thickened crust which thinned towards the margins to 15 km utilizing gravity and existing refractions results. Richardson et al. (2000) calculated a crustal thickness of 38 km with similar thinning. Unlike Richardson et al. (2000), updated shear wave tomography studies by Covellone et al. (2015) and Chapter 1 of this dissertation found the thickened crust was absent below 30 km though all three studies lack both the near surface and vertical sensitivities to be conclusive. Tharimena et al. (2016), utilizing SS precursors, suggests a crustal thickness of 28 ± 4 km which is on the lower end of the previous studies. Receiver functions by Tonegawa et al. (2018) had the most similar results to this study with approximately 20 km along the Western and Eastern edge with the OJP thickening between 30-40 km over the central OJP before thinning again to around 20 km on the Eastern edge of OJP before transitioning to normal >100 Ma oceanic crustal thicknesses, which compares well to this study

with 20 and 22.5 km along the West and East boundaries and 32.5 and 40 in the central OJP. Similarly, the thicknesses observed of 18-25km on the eastern OJP (22.5 km) are comparable with crustal thicknesses of the MP (24-27 km) and the HP (24 km) (Stern et al., 2020). Overall, the results of this study agree with most of the previous work and are within the range for all the previous studies while utilizing a new methodology for this area.

Sediment thickness

Unlike estimates for crustal thickness from various geophysical techniques, sediment thickness at the OJP has been well documented through various drilling expeditions and programs at select locations across much of central and parts of Eastern OJP. Most crustal thickness studies, except active source refraction profiles and drilling programs, ignore the thin sediments and only relay bulk thicknesses. Though known in certain areas around the OJP, we allowed the sediment thickness to vary between 0 and 2 km unlike other studies (Stern et al., 2020). Sediment thicknesses across OJP are variable along the edges while consistent in the central OJP. In the central OJP, three sites, IODP 192 holes 1183 and 1186 and DSDP hole 289 have sediment thicknesses of 1.130 km, .968 km, and 1.250 km (Mahoney et al., 2001; Winterer et al., 1971). In IODP 192, hole 1184, along the Eastern salient of the OJP, were found to be 0.201 km before reaching volcanoclastics. While further South but closer and similar to the central OJP, hole 288 bottomed at just under 1 km failing to find basement (Mahoney et al., 2001). Along the northern edge of the OJP, sediment thickness thins from 1.458 km at IODP 130, hole 807, in the north-central, (Kroenke et al., 1991) to ODP hole 803 at 1.187 km, and IODP 192 hole 1187 at 1.185

km. It thins dramatically further to the southeast at IODP 192 hole 1185 at 0.308 km (Mahoney et al., 2001). The variable sediment thickness correlates well with the estimated eruption depth below the sea surface where estimated (Roberge et al., 2005). At site 807 eruption depth is ~3km with 1.458 km of sediment with shallowing eruption depth with thinning sediment thickness. The only outlier is 1183 which erupted at 1.072 km depth but has a sediment thickness of .968 km. Similar sediment thickness sites were estimated to have erupted below 2 km (Roberge et al., 2005).

The sediment thicknesses in the central OJP region align well with the preferred models of Location 2 and 3 while Location 4, which most directly samples the high plateau, prefers a thinner 0.2 km thick sediment. Wave speed sensitivity to thin sediment layers is small even for the slower wave speeds (Figure 2.5) and only a limited number of sediment thickness and wave speeds were used. As such, the preferred model for Location 4 may have accommodated the wrong sediment thickness by thickening the total crust, decreased wave speed of the crustal layers, or both. The East-central OJP location 6 prefers sediment thicknesses (1.25 km) more like that of the high plateau compared to most of the thin sediment areas the event-station pairs traverse. Similarly, to the East-central OJP, location 7 additionally prefers thicker sediments than the region it sampled, 1 km modeled vs. 0.2 km. Insensitivity to thin slow sediments at the surface would account for this discrepancy where changes in total crustal thickness and wave speed are preferred. Sediment thickness results from Tonegawa et al. (2019) were also insensitive to thin sediments with RF observations of > 1km along the far West OJP, 0.75-1 km in the central OJP, and ~ 1km along the eastern salient. Modeled sediment thickness results match

relatively well with the drilling information for areas with thick sediments in the West-central OJP but poorly in areas with thinner sediments like those to the East. This is a surprising result as the model has little, if any, sensitivity to sediment thickness and may be purely coincidental. The variable sediment thickness of the OJP deserves additional scrutiny but is outside the scope of this work.

Mantle Wave Speeds

P-Wave speeds in the mantle are remarkably consistent at 8.17-8.4 km/s across the OJP in our study (Figure 2.5). While Locations 1, 2, 3, 4, and 7 have narrow event-station paths thru the OJP, the 8.17 km/s mantle wave in the East-central OJP area represents the largest area location tested with the most diverse ray path trajectories (Figure 2.1). Additionally, one event represented each preferred model depth range (Appendix 2.1) and only one event originated in the crust (Table 2.1) with the two shallowest events (15 and 28 km) preferring 8.17 km/s while the two deeper events (46 and 155 km) preferring 8.4 km/s. Hussong et al. (1979) observed mantle P-wave speeds of 8.3 km/s over OJP while multiple seismic refraction profiles by Furumoto et al. (1970, 1976) detected mantle P-wave speeds between 8.0 and 8.6 km/s; uppermost upper mantle PREM wave speeds are 8.0 km/s while 8.5 km/s is much deeper at 220 km depth (Dziewonski and Anderson, 1981). Similarly, Gladchenko et al. (1997) calculated mantle P-wave speeds of 8.0 km/s but this analysis used refraction wave speeds already discussed. These 8.0-8.6 km/s P-wave speeds are not emblematic of the region as P-wave speeds South of OJP near the Solomon Islands were found to be in the range of 7.3-8.0 km/s (Furumoto et al., 1970) and Ku et al. (2018) calculated very slow mantle P-wave speeds of 7.24 km/s utilizing surface

waves though this could be adjusted utilizing a different V_p/V_s ratio. Very fast P-wave speeds are observed at HP and MP, 8.7-8.8 respectively (Stern et al., 2020), which are more typical of PREM at 300-400 km depth (Dziewonski and Anderson, 1981), which, compared to OJP, conflicts with the single origin hypothesis for OJP, HP, and MP unless events following formation 120Ma generated these differences.

Crustal Wave Speeds

The resulting upper and lower crustal P-wave speeds from the modeled Pn arrivals vary across OJP with some lateral similarities. The Far West and West OJP locations share identical crustal P-wave speed structure with a two-layer upper crust of 5.4 and 6.1 km/s and a lower crust 6.9 km/s (Figure 2.5); wave speeds of 5.0, 6.6, and 7.1 km/s are typical for upper, mid, and lower oceanic crust (Mooney et al., 1998) while 5.5, 6.3, and 6.7 km/s are typical for continental crust (Hadley and Kanamori, 1977). Just to the East, the West-central P-wave structure is a single intermediate crustal P-wave speed of 5.8 km/s though other preferred models for this location have distinct upper and lower crustal P-wave speeds similar to the surrounding locations. Similar to the West OJP profiles, East-central OJP has a layered upper crust of 4.1 and 5.3 km/s but a slower than expected lower crust of 6.2 km/s. These wave speeds are slower than typical oceanic crust (Mahoney et al., 1998).

The P-wave speeds of the upper most upper crust are similar to those determined from drill cores. The 5.4 km/s of the Western OJP locations corresponds to basalts from IODP 130 hole 807 while the slower 4.1 km/s is similar to that of the pillow basalts observed in site 642 (Planke and Edholm, 1994). Furumoto et al. (1976) postulated that the P-wave speeds of 6.1-6.2 km/s are representative of a middle

crustal layer where in our profiles, wave speeds of 5.3 and 5.4 km/s are upper crustal layers and P-wave speeds of 6.9-7.0 km/s are lower crust which is similar to that of gabbro. The fastest lower crust, possibly transitional crust in this study, is similar to that of Gladchenko et al. (1997) (7.1 km/s) though much slower than that observed by Hussong et al. (1979) of 7.8 km/s and Furumoto et al. (1976) of 7.8 km/s which can also be observed in Precambrian shields (Furumoto et al., 1976). These fast, observed, upper most mantle P-wave speeds are confined to a thin layer (Hussong et al., 1979; Furumoto et al., 1976) and our methodology did not account for a two-layer lower crust with a basal layer just above the mantle. Additionally, this fast basal crust was only observed in the central and northern OJP and missing from the South. As our ray paths assume a 1D structure, this cannot be differentiated.

The crustal P-wave speeds and depths of the layers in the Eastern OJP from this study are similar to MP observed by Hussong et al. (1979) with a two-layer upper crust of 5.1 and 6.1 km/s between 6 and 14 km depth and a lower crust of 6.9 km/s.

OJP Formation

The crustal thickness of an anomalous feature in the Pacific Ocean can be used to help elucidate its formation. OJP thought to be a Large Oceanic Province formed 120 Ma due to the initiation of a plume, fast spreading triple junction, or a combination of both (Coffin and Edholm, 1994; Gladchenko et al., 1997; Tejada et al., 2002; Gladchenko et al., 1997); a bolide impact has also been proposed (Korenaga 2005) but not entertained in this study. Continental flood basalts and hotspots have thick crust usually exceeding 40 km while oceanic flood basalts and hotspots are thinner. Multiple studies, including this one, determined that the OJP, at least at its center

(Location 4: West Central OJP), has a thicker crust compared to an average oceanic plateau or hotspot but still less than its continental counterpart, though there are similarities. The lower crust P-wave speed of 6.9-7.0 km/s, higher in other studies, is faster than continental wave speeds (6.7 km/s) but slower than oceanic mantle (7.2-7.7 km/s) (Gladczenko et al., 1997). This makes any direct comparison to standard ocean or continental crust difficult. The thicker than average crust is indicative of a significant volume of magma produced at the OJP especially if the MP and HP were once together (e.g., Hochmuch et al., 2015). The thickened crust (>20 km) along the edges of the OJP determined from this work, as well as RFs from Tonegawa et al. (2019), indicate that the OJP may be larger at its edges than previously delineated by the 4000m bathymetric contour (Coffin and Edholm, 1994). Gladczenko et al. (1997) suggests that the thickened crust West of OJP in the Lyra basin was due to post emplacement extrusives but still part of the larger OJP LIP. Unlike crustal thicknesses to the West, North, and East, the Southern OJP may be thickened due to oceanic under-thrusting at the subduction zone (Furumoto et al., 1970) which may also explain its lack of a fast basal P-wave speed observed in the North in many refraction studies (Furumoto et al., 1970, Furumoto et al., 1976, Hussong et al., 1979).

The mechanism for formation is still debated though most recent works prefer a plume head model (e.g., Covellone et al., 2015) or a plume caught in fast spreading triple junction (e.g., Gladczenko et al., 1997). A simple plume head model would suggest a thickened central crust with thinner crust around the edges of the anomaly. If the plume head became detached or separated from the plume conduit, as it approached the surface, it could generate a temporary hiatus in volcanism while, if

continuous, it could create thicker crust compared to the edges in the opposite direction of plate motion. Additionally, the requirement of subaerial eruption is only necessary if the OJP crust was only formed by extrusive pillow basalts. Subaqueous eruptive evidence can still be preserved if the OJP is uplifted while the crust thickens internally thru intrusions, then sink back below with no indication of subaerial activity. Based on geochemical analysis and plate reconstruction, OJP, MP, and HP were once contiguous (Taylor, 2006; Ito and van Keken, 2007). MP and HP are both thought to have crustal thicknesses of 20-25 km which are similar to those estimated for the edges of OJP. This would place the thickest crust over central OJP with a large area to the East with 20-25 km thick crust. This would be more indicative of a plume interacting with a ridge than a simple plume model. Gladchenko et al. (1997) also suggested a near-ridge plume creating a non-uniform plateau with thick laterally changing crust with magma migrating thru asymmetric fractures. Other authors have suggested underplating from more recent volcanic activity at 60, 40, and 20 Ma around OJP (e.g., Tejada et al., 2002; Simonetti and Neal, 2010) but this would result in a warmer and lower P-wave speeds in the lower crust compared to the middle crust, which is not observed in other seismic studies and not tested here.

Recent seismic studies and plate reconstructions have made comparisons of the MP and HP to the OJP accessible. Plate reconstruction (e.g., Taylor, 2006; Hochmuth et al., 2015) and geochemistry (Ito and van Keken, 2007), by far, relate these isolated structures back to a single larger plateau. Hochmuth et al. (2015) reassembled the three plateaus thru rifting, stretching, and shearing which takes into consideration the thickened crust outside of the OJP boundary. We found 6.9 and 7.0 km/s P-wave

speeds in the lower crust from our modeling and other seismic refraction studies found 7.8 km/s (Furumoto et al., 1976; Hussong et al., 1979); very fast lower crust wave speeds were not tested in our models. The higher wave speed (7.8 km/s) observed by Furumoto et al. (1976) is thought to be olivine and pyroxene crystal fractionation from a LIP trapped above the mantle (Ridley and Richards, 2010; Karlstrom and Richards, 2011). A similar high P-wave speed of 7.3 km/s transitional crust found under the MP provides good correlation between the OJP and MP (Hochmuth et al., 2015). Dissimilarly, extremely fast P-wave mantle wave speeds of 8.7-8.8 km/s were observed at the MP and HP by Stern et al. (2020) while we modeled mantle P-wave speeds of 8.4 km/s under the OJP in this study is much slower; PREM wave speeds at 400 km depth go from 8.9 to 9.13 km/s (Dziewonski and Anderson, 1981). This discrepancy between MP mantle wave speeds could be attributed to the difference in age of each analysis with the more recent MP and HP P-wave speeds being ~45 years newer but such a fast wave speed beneath the OJP would have been very distinct even in the older refraction surveys. A reexamination or new OJP survey may result in more similar mantle P-wave speeds.

CONCLUSIONS

A dataset of synthetic P/Pn waveforms, generated from the forward modeling of one-dimensional sediment, crust, and uppermost upper mantle of various thicknesses and wave speeds (Figure 2.2) using the FK integration method, were compared to P/Pn arrivals that traversed the OJP and its surroundings. A pseudo two-dimensional model that transects the OJP from East to West was created from the one-dimensional models

from each location (Figure 2.1) where the time difference between the synthetic arrival vs. the real data are minimized (Figure 2.5). The sediment and crustal thicknesses are relatively consistent with existing drilling and refraction datasets. Additionally, the wave speeds for the crust and uppermost upper mantle are consistent along the two-dimensional profile and with most other studies.

Along the edges of the OJP, the crustal thickness calculated to a thickness of 20-23 km, similar to that of the MP and HP which during and immediately after formation resided to the West and Southwest respectively (Hochmuth et al., 2015). Within the bathymetric boundaries of the OJP, the crustal thickness varied from 32.5 km on the far western side and 40 km in the West center then thinning to the East at 23.5 km. A consistent mid-crust P-wave speed of 5.8-6.1 km/s was observed while a high P-wave speed of 6.9-7.0 km/s was detected sporadically beneath the OJP. A strong uppermost upper mantle P-wave speed of 8.4 km/s was coherent across most of the OJP which, though consistent with several previous wave speed studies, is slower than recent studies of the MP and HP.

Our analysis, along with most other seismic analysis of the OJP, agree that a plume origin is the most likely. The high P-wave speed in the lower crust at the OJP is indicative of a LIP formation while the uppermost upper mantle P-wave speed are dissimilar to the other plateaus but geochemically similar. Overall, the pseudo two-dimensional crustal thickness and wave speed model of the OJP is consistent with a formation more closely associated with a significant volume of magma produced by a plume head associated with a LIP interacting with a ridge resulting in a variable asymmetric crust.

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Loc	EQ Name	EQ Mag	EQ Dep (km)	EQ Lat	EQ Lon	EQ Date	EQ Time (GMT)	Station Name	Dist Deg	SAC Pn Arrival (s)	User Pn Arrival (s)
1	201408030022A	6.9	19.4	0.89	146.23	08/03/2014	00:22:12	AU.RABL	7.79	103.98	103.98
1	101594A	5.6	15	-3.67	152.06	10/15/1994	00:39:33	XT.TKK	11.04	151.69	149.0
1	062595A	5.5	15	-3.16	150.76	06/25/1995	02:10:42	XT.TKK	10.6	149.66	144.5
1	042595B	5.6	57	-6.07	147.64	04/25/1995	06:15:12	XT.TKK	14.08	185.75	189.0
1	081795F	5.5	28	-5.44	153.5	08/17/1995	10:01:37	XT.PNI	13.21	175.97	190.0
2	201607251938A	6.4	12	-2.97	148.18	07/25/2016	19:38:50	PS.PATS	14.04	194.23	190.0
2	200511051048A	5.8	13	-3.17	148.61	11/05/2005	10:48:26	PS.PATS	13.88	191.49	187.5
2	103102A	5.3	15	-3.38	148.95	10/31/2002	01:35:20	PS.PATS	13.8	191.53	192.0
2	060602C	5.7	15	0.8	148.58	06/06/2002	23:53:55	PS.PATS	12.32	167.91	167.5
3	111700Q	6.2	17	-5.26	152.34	11/17/2000	21:02:20	PS.PATS	13.41	165.95	184.0
3	071503E	5.9	15	-3.79	152.3	07/15/2003	18:46:43	PS.PATS	12.14	166.87	163.0
3	200512111420A	6.2	12	-6.73	152.36	12/11/2005	14:20:52	PS.PATS	14.28	199.71	195.0
3	201504301045A	6.7	38	-5.58	151.83	04/30/2015	10:45:09	PS.PATS	13.93	188.11	188.11
3	051799B	5.5	36	-5.43	152.82	05/17/1999	10:08:02	PS.PATS	13.36	182.42	182.0
3	201106160003A	6.1	21	-6.12	151.21	06/16/2011	00:03:43	PS.PATS	14.69	198.89	194.0
3	201505010806A	6.8	49	-5.47	151.76	05/01/2015	08:06:10	PS.PATS	13.87	187.1	187.1
3	020600C	6.6	45	-6.06	151.19	02/06/2000	11:34:01	PS.PATS	14.65	195.0	198.5
3	051699A	6.0	42	-4.99	152.76	05/16/1999	00:51:32	PS.PATS	12.99	170.32	171.5
3	011999A	5.8	87	-4.72	153.66	01/19/1999	03:35:43	PS.PATS	12.38	163.46	157.0

Table 2.1. Table of Earthquakes-Station pairs used in the Pn analysis

3	051099C	6.5	144	-5.38	150.97	05/10/1999	20:33:09	PS.PATS	14.81	186.88	190.25
4	041901I	6.0	16	-7.56	155.79	04/19/2001	21:43:50	PS.PATS	14.52	196.47	200.0
4	090696K	5.6	15	-7.5	155.84	09/06/1996	17:03:52	PS.PATS	14.45	198.77	198.77
4	201505070710A	7.1	12	-7.36	154.49	05/07/2015	07:10:33	PS.PATS	14.61	192.42	193.0
4	201304140132A	6.0	29	-6.74	154.56	04/14/2013	01:32:27	PS.PATS	14.00	191.2	193.0
4	200609011018A	6.3	46	-7.01	155.42	09/01/2006	10:18:59	PS.PATS	14.05	188.65	185.0
4	201404110707A	7.1	44	-6.78	155.09	04/11/2014	07:07:33	PS.PATS	13.90	183.59	178.0
4	101703B	6.2	138	-5.5	154.07	10/17/2003	10:19:11	PS.PATS	13.0	175.39	175.39
4	201404190104A	6.6	35.8	-6.92	154.99	04/19/2014	01:04:10	PS.PATS	14.1	190.36	196.0
6	200501222030A	6.2	28	-7.66	159.69	01/22/2005	20:30:22	PS.PATS	14.47	197.51	197.51
6	110295A	5.5	15	-9.95	159.9	11/02/1995	16:08:45	XT.NAU	11.7	164.89	162.0
6	042194B	5.8	46	-5.77	153.95	4/21/1994	03:51:54	XT.KOS	14.26	189.21	189.21
6	091695A	5.8	152	-6.18	155.5	09/16/1995	01:03:42	XT.NAU	12.72	170.89	172.5
7	201507180227A	7.0	12	-10.3	165.1	07/18/2015	02:27:40	IU.TARA	13.99	191.85	187.5
7	201302081112A	6.3	12	-10.91	165.81	02/08/2013	11:12:25	IU.TARA	14.1	185.37	186.5
7	201302071859A	6.4	15	-10.94	165.59	02/07/2013	18:59:24	IU.TARA	14.23	193.16	191.5
7	200708121205A	5.8	48	-11.49	165.98	08/12/2007	12:05:23	IU.TARA	14.51	198.79	197.5
7	201306050447A	5.9	47	-11.47	166.0	06/05/2013	04:47:30	IU.TARA	14.48	198.27	198.27
7	201201090407A	6.6	42	-10.56	165.04	01/09/2012	04:07:21	IU.TARA	14.1	192.15	192.15
7	201403270349A	6.0	116	-12.13	166.46	03/27/2014	03:49:47	IU.TARA	14.85	199.68	205.0

Table 2.1. Continued

Loc.	Depth (km)	Model	Sed.	Crust	Total (km)	Mean (s)
1	0-20	PAC06	0.2	19.8	20.0	0.933
1	20-40	PA2	1.0	4.0	5.0	6.85
1	40-60	PAC06	0.2	29.8	30.0	0.0099
1	All Depths	PAC06	0.2	19.8	20.0	0.9324
2	All Depths	PAC06F76p2	2.0	30.5	32.5	1.5475
3	0-20	PHB3F76p3	2.0	30.5	32.5	2.19
3	20-40	GT1r5F76p2	2.0	30.5	32.5	1.98
3	40-60	PA5F76p2	2.0	30.5	32.5	1.78
3	60-100	GT1r5F76q3	1.25	38.25	39.5	2.92
3	>100	PA5F76q2	1.25	38.25	39.5	0.18
3	All Depths	PA2F76p2	2.0	30.5	32.5	3.73
4	0-20	GT1r5	1.3	15.0	16.3	2.833
4	20-40	PA5	0.0	3.0	3.0	1.055
4	40-60	PAC06F76q2	1.25	38.25	39.5	2.52
4	>100	PA5F76q2	1.25	38.25	39.5	0.48
4	All Depths	PAC06	0.2	39.8	40.0	4.288
6	0-20	PA2F70a1	1.25	22.0	23.25	0.02
6	20-40	PA2	1.0	4.0	5.0	0.1
6	40-60	PAC06	0.2	9.8	10.0	0.17
6	>100	PAC06F76q1	1.25	38.25	39.5	0.01
6	All Depths	PA2F70a1	1.25	22.0	23.25	0.75
7	0-20	PAC06F70c1	1.0	21.5	22.5	1.16
7	40-60	PA2S	1.0	4.0	5.0	0.876
7	>100	PA5	0.2	4.8	5.0	0.1
7	All Depths	GT1r5	1.0	3.0	4.0	0.88

Table 2.2. Best fitting models information for each location.

Figure 2.1: Map of earthquake and station pairs used in the Pn analysis. Each color corresponds to a specific group of event-station waveforms referred to as Location 1: West of OJP (Green), 2: Far West OJP (Blue), 3: West OJP (Red), 4: West-central OJP (Yellow), 6: East-central (Black), and 7: East OJP (Magenta).

Figure 2.2: 1D V_p input models used to generate synthetic waveforms. GT1 (Red) represents 110+ Ma Pacific Ocean (Nishimura and Forsyth, 1989), PA5 (green) from a Tonga dataset (Gaherty et al., 1996), PA2 (yellow) developed from mostly 100+ Ma Pacific lithosphere (Lerner-Lam and Jordan, 1987), PAC6 (blue) represents the Pacific plate (Tan and Helmberger, 2007), PHB3 (magenta) was derived from a Philippine Sea dataset (Kato and Jordan, 1999), and IASPI (turquoise) is a global 1D model. Each colored 1D V_p profile represents a base model used excluding crustal thickness variations. Gray 1D V_p profiles are variations of the base models using seismic refraction measurements from the Ontong Java Plateau region.

Figure 2.3: Example synthetic Pn arrivals vs. real Pn waveforms. (a) Pn arrivals (black) from EQ 081795F to XT.PNI (13deg) vs. fk synthetic data (red). Model used for synthetic is PA5 velocity model with no water or sediment. Data and synthetics filtered to 5 – 100s. Inset map shows the location of the event Centroid Moment Tensor (CMT), XT.PNI station (red circle), and ray path (red dotted line) over the Ontong Java Plateau (black outline).

Figure 2.4: Plot of the time differential between each earthquake from the corresponding synthetic per crustal thickness for each location region and all locations. The zero crossing is highlighted by the long-dashed line. All data is represented as gray circles. One Pn time differential for each 1D Vp base model, as shown in Figure 2.2, at each location is highlighted. Note: Loc. 7 data above 20 seconds is cropped to highlight changes in crustal thickness.

Figure 2.5: Plot of the best fitting 1D synthetic models for each location along the Ontong Java Plateau. Each Color corresponds to the location in Figure 2.1. The dashed line indicates the approximate location of the Moho between each 1D model. Note the depth change in Moho depth along the East-West transect across the Ontong Java Plateau.

APPENDIX

Loc.	Earthquake	Station	Model	Sed.	Crust	Total (km)	Diff (s)
1	042595B	XT.TKK	PAC06	0.2	29.8	30.0	0.00999
1	062595A	XT.TKK	PAC06F76p1	2.0	30.5	32.5	0.05
1	101594A	XT.TKK	PHB3F76p3	2.0	30.5	32.5	0.02999
1	201408030022A	AU.RABL	PAC06F76q3	1.25	38.25	39.5	0.03999
1	081795F	XT.PNI	PA2	1.0	4.0	5.0	6.85
2	060602C	PS.PATS	PHB3F76p2	2.0	30.5	32.5	0.00999
2	103102A	PS.PATS	PA5Glad97min	0.0	15.2	15.2	0.05
2	200511051048A	PS.PATS	PAC06F76w1	1.5	34.5	35.5	0.08
2	201607251938A	PS.PATS	PAC06F76p2	2.0	30.5	32.5	0.14999
3	200512111420A	PS.PATS	PAC06	0.2	4.8	5.0	0.02999
3	071503E	PS.PATS	PAC06	0.2	19.8	20.0	0.00999
3	111700Q	PS.PATS	GT1r5	1.3	11.0	12.3	0.0
3	201106160003A	PS.PATS	PAC06F76q2	1.25	38.25	39.5	0.02999
3	051799B	PS.PATS	PAC06	0.2	39.8	40.0	0.21
3	051699A	PS.PATS	GT1r5F76q2	1.25	38.25	39.5	0.11
3	020600C	PS.PATS	PAC06	0.2	49.8	50.0	0.36
3	201505010806A	PS.PATS	PAC06	0.2	39.8	40.0	0.12999
3	011999A	PS.PATS	GT1r5F76q3	1.25	38.25	39.5	2.91999
3	051099C	PS.PATS	PA5F76q2	1.25	38.25	39.5	0.17999
3	201504301045A	PS.PATS	GT1r5F76p2	2.0	30.5	32.5	0.03999
4	201505070710A	PS.PATS	PA2F76q2	1.25	38.25	39.5	0.53999
4	090696K	PS.PATS	GT1r5	1.3	15.0	16.3	0.07999
4	041901I	PS.PATS	GT1r5	1.3	19.0	20.3	0.0
4	201304140132A	PS.PATS	PA2	1.0	29.0	30.0	0.05
4	201404190104A	PS.PATS	GT1r5	1.3	40.0	50.3	0.0
4	201404110707A	PS.PATS	GT1r5F76q3	1.25	38.25	39.5	1.02999
4	200609011018A	PS.PATS	PHB3F76q2	1.25	38.25	39.5	0.11999
4	101703B	PS.PATS	PA5F76q2	1.25	38.25	39.5	0.47999
6	110295A	XT.NAU	PA2F70a1	1.25	22.0	23.25	0.02
6	200501222030A	PS.PATS	PA2	1.0	4.0	5.0	0.1
6	091695A	XT.NAU	PAC06F76q1	1.25	38.25	39.5	0.00999
6	042194B	XT.KOS	PAC06	0.2	9.8	10.0	0.17
7	201201090407A	IU.TARA	GT1r5F76p2	2.0	30.5	32.5	0.51999
7	201306050447A	IU.TARA	GT1r5	1.3	11.0	12.3	0.03999
7	200708121205A	IU.TARA	PA2	1.0	4.0	5.0	0.42999
7	201403270349A	IU.TARA	PA5	0.2	4.8	5.0	0.0
7	201302081112A	IU.TARA	PAC06	0.2	4.8	5.0	0.02
7	201507180227A	IU.TARA	PAC06F70c1	1.0	21.5	22.5	0.02
7	201302071859A	IU.TARA	PAC06F70a1	1.25	22.0	23.25	0.02

Appendix 2.1. Best fitting model for each event-station pair.

CHAPTER 3

Sensitivity kernels from ScS and ScS2: Interpretation constraints on core reverberations around Ontong Java Plateau

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ABSTRACT

ScS phases have been used to estimate the attenuation structure in the upper mantle in the southwest Pacific including most recently by Gomer and Okal (2003) which utilized a deep earthquake to calculate Q beneath the Ontong Java Plateau. We generated synthetic waveforms of ScS phases in a 3D spectral element wave propagation program in three increasingly complex wave speed and Q model sets where we modified the wave speed and Q at various depth intervals to test how Q and wave speed affect the amplitudes of ScS phases.

Arrival and amplitudes varied depending on which wave speed and Q model was used but reference and modified waveforms suffered from “ringiness” from unknown sources in the simulation. One-dimensional synthetics of varying Q in the upper mantle show that spectral amplitude ratios are mostly insensitive to these changes. Our observations show that distinguishing amplitude changes associated with Q or wave speed is incredibly difficult and constraining those changes to a specific depth is equally challenging.

INTRODUCTION

The OJP is a complex geologic feature in a tectonically complicated part of the world. The formation of the OJP at around 120-125 Ma is mainly hypothesized to be from the initiation of a deep mantle plume head that erupted subaqueously uninfluenced by other tectonic forces (Griffiths et al., 1989; Campbell, 1998; Neal et al., 1997; Chandler et al., 2012; Tejada et al., 1996; Michael, 1999; Ely and Neal, 2003) or from the onset of a plume that interacted with a fast spreading triple junction

(Dordevic and Georgen, 2016; Gladchenko et al., 1997) both resulting in at least part of the largest known LIP (Coffin and Edholm, 1994; Gladchenko et al., 1997; Tejada et al., 2002). After formation, the LIP, consisting of the OJP, MP, and HP, is hypothesized to have broken apart intermittently between 120-80 Ma and the fragments drifted towards their current locations (Billen and Stock, 2000; Worthington et al., 2006; Hochmuth et al 2015). The OJP experienced several episodes of volcanism between the onset of the LIP and when it struck the Solomon arc subduction zone between 25-22 Ma (Musgrave, 1990; Mahoney et al., 2001). The interaction of the OJP with the subduction zone coincided with the cessation of volcanism at the OJP, plate re-arrangements, a subducted Pacific plate breaking off beneath the Solomon Arc, and a subduction polarity reversal with the Australian plate (Kroenke et al., 1986, Yan and Kroenke, 1993, Musgrave, 1990; Petterson, 1995; Petterson et al., 1999; Mann and Taira, 2004; Hochmuth et al., 2015).

Seismic studies and analyses have made improvements in studying the complicated upper mantle. Regional seismic studies of the OJP and its surroundings have focused on the crust and upper mantle while a few have looked at deeper structure. Using tomographic methods, Richardson et al. (2000) found a large slow seismic shear wave root in the upper mantle below the OJP, however Covellone et al. (2015) and Chapter 1 of this work found a high shear wave speed anomaly beneath the OJP. Seismic attenuation studies, using surface waves, have shown that the OJP is lower than the global average (Selby and Woodhouse 2002; Dalton et al., 2008, Warren and Shearer, 2002) while Romanowicz and Mitchell (2015) showed that 3D attenuation models for the OJP region are variable.

ScS phases have also been used to study the seismic wave speed and attenuation structure of the OJP and surrounding Pacific mantle. ScS phases are S-waves generated by an earthquake that reflects off of the core mantle boundary (CMB) and recorded at the surface still as an S-wave. Usually, deeper events are utilized to avoid any crustal contamination near the source. Phases ScS^2 (ScSScS) and ScS^3 (ScSScSScS) are from the same earthquake that have reflected off the CMB and free surface an additional one or two more times. ScS measurements are typically done by using a version of spectral ratio method (Jordan and Sipkin, 1977; 1980). Spectral amplitudes of the ScS phases are retrieved, smoothed, normalized, and stacked. The spectral ratio is then calculated between ScS phases. ScS attenuation studies found a Q_{ScS} of 155 for the older ocean basins of western Pacific (Sipkin and Jordan, 1980) with specific Tonga to Japan raypaths having a Q_{ScS} of 173, 214, and 172 (Sipkin and Jordan 1980; Chan and Der, 1988; Suetsugu 2000) and a Q_{ScS} of 226 from Papua New Guinea to Taiwan (Revenaugh and Jordan, 1991). From Fiji to Hawaii, the younger oceans had a Q_{ScS} of 169 (Revenaugh and Jordan, 1991) and 138 between Tonga and South America (Chan and Der 1988); PREM Q ranges in the upper mantle are 80 at 80 km depth and 143 between 220 km and the transition zone (Dziewonski and Anderson, 1981). In the Coral sea, between the OJP and Australia, a Q_{ScS} of 109 was determined but the ScS travel path crosses the polarity reversed Solomon arc making interpretation difficult (Gomer and Okal, 2003).

For the OJP structure in the upper mantle, Gomer and Okal (2003) used multiple ScS phases from a single event to calculate a Q_{ScS} of 366 with a lower bound of 253 in conjunction with an upper mantle slow shear wave speed anomaly similar to that of

Richardson et al. (2000). Most recently, Suetsugu et al. (2019) found an average Q_{Scs} of 309 in the mantle below OJP. Taking into account one-dimensional shear wave and Q structure for the lowermost mantle beneath the Pacific, Suetsugu et al. (2019) also found Q_{Scs} of 367, similar to that of Gomer and Okal (2003) in the upper mantle. Additionally, Suetsugu et al. (2019), using a spectral ratio method, found a Q_{Scs} of 324 for the OJP for the same event-station pair used by Gomer and Okal (2003). These OJP Q_{Scs} results have a similar Q structure to that of stable continents (Sipkin and Jordan, 1980; Chan and Der, 1988; Revenaugh and Jordan, 1991; Sipkin and Revenaugh, 1994) and higher than the average Q_{Scs} of the Pacific.

The mantle, specifically the lower mantle, was initially thought to be homogeneous and often ignored while more recent studies have shown that the mantle is very compositionally and seismically heterogeneous with various areas of upwelling and downwelling, subducting slabs, and broken slabs (Ritsema and Lekic, 2020). There are significant heterogeneities in the several hundred kilometers above the core-mantle boundary associated with the D'' layer (Kim et al., 2020). Additionally, a Large Low Shear Velocity Province (LLSVP) is present in the South Pacific, centered at $\sim 200^\circ$ longitude ($160^\circ E$) and stretching $\sim 100+$ degrees wide, with the OJP near its present western edge. The LLSVPs are thought to be denser than their surroundings (Ishii and Tromp, 1999; Trampert et al., 2004; Moulik and Ekström, 2016) and have complex low shear wave speeds with poorly defined edges (Ritsema and Lekic, 2020). Near the edges of the LLSVP, the S-wave speeds change dramatically by several percent across their margins (Ritsema et al., 1998; Ni et al., 1999; Wen 2002; Ritsema and Lekic, 2020).

Recent work has also made improvements in our understanding of the D". Studies looking at the LLSVP in the D" beneath the OJP have shown anisotropy (Deng et al., 2017) and strong wave speed gradients along its edges (Ritsema et al. 1998, Ni et al. 1999, Wen 2002). Additionally, Garnero et al. (2016) has shown that in low shear velocity provinces, P-wave speed is also diminished. Konishi et al. (2017) calculated Q in the Pacific LLSVP at 260 between 2000 and 2850 km and 216 between 2850 and the CMB using earthquakes between Tonga and Fiji with Japan; PREM Q in the lower mantle is 312 (Dziewonski and Anderson, 1981).

Studies of Q_{ScS} , like any seismic method, have potential sources of error and assumptions. Global attenuation models do not always align with Q_{ScS} studies, leading to discrepancies at multiple scales (Suetsugu et al., 2019). Moderately noisy waveforms can lead to errors in ScS amplitude measurements and Q results (Gomer and Okal, 2003). Some Q_{ScS} methodologies assume that all wave speed and attenuation discrepancies are from a heterogeneous upper mantle and not the presumed homogeneous lower mantle (Gomer and Okal, 2003). In recent studies, it has been shown that the lower mantle is not a homogenous simple layer but is in fact also heterogeneous with the D", LLSVP, regions of focusing and defocusing like upwellings, and downwellings, as described above, and the entire ScS path should be reconciled (Chaves and Ritsema, 2016). In other Q_{ScS} methodologies, where a Q model is used in the attenuation measurement, different Q models will generate varying wave speed delays (Butler 1977; Zhou, 2009; Savage et al., 2010; Ruan and Zhou, 2010, Chaves and Ritsema, 2016). Additionally, the ScS waveforms are usually treated as azimuthally isotropic rays and assumed to be unaffected by multipathing

and amplification by heterogeneities (Chaves and Ritsema, 2016). Liu and Tromp (2008) generated ScS sensitivity kernels (Fréchet derivatives) which show that the phase is most sensitive at the event, receiver, and at the CMB reflection points with spatially large, limited sensitivity around the raypath; the spatial extent of the sensitivity kernel is a function of the period. Not only are ScS phases least sensitive along the raypath but the sensitivities can be spatially large, on the order of 10s of degrees in size (Liu and Tromp, 2008). Typically, measurements of Q using ScS multiples are made at periods between 20 and 100 seconds, resulting in spatially expansive Fréchet sensitivity kernels. With strong lateral heterogeneities around the OJP at the surface (Papua New Guinea, Solomon Islands, Solomon arc) and at depth (subducted Pacific plate, LLSVP, D”), Q_{ScS} analyses could be severely impacted. Chaves and Ritsema (2016) generated synthetic waveforms from one-dimensional Q in 3D simulations where they found that wave speed heterogeneities in the deep mantle can not only affect ScS waveform amplitudes but may overwhelm amplitudes solely from attenuation. Additionally, Kanamori and Rivera (2015) had similar results showing Q_{ScS} results are influenced by 3D structure. Moreover, the method for determining Q_{ScS} using the slope of spectral ratio depends directly on the relative travel time between subsequent arrivals (Gomer and Okal, 2003). To investigate Q_{ScS} assumptions and build upon the work by Kanamori and Rivera (2015) and Chaves and Ritsema (2016), we proposed to use a three-dimensional wave simulation to generate synthetic seismograms in modified one- and three-dimensional Q and wave speed models. This will allow us to investigate the effects of

heterogeneity along the wave propagation path, with bounce points on the OJP, on the ScS phases of amplitude and arrival time.

METHODOLOGY

To determine the influence of Q and wave speed structure on Q_{ScS} we use a three-dimensional spectral-element wave simulation to generate synthetic seismograms. We use SPECFEM3D_GLOBE, a software package that uses the spectral-element method to simulate the 3D propagation of seismic waves in the Earth (Komatitsch and Tromp 2002a, b). This program is well suited to this analysis as the global spectral-element mesh was designed to model discontinuities including the core-mantle boundary (CMB) (Liu and Tromp, 2008). To relate this study back to the OJP, we used the 510 km deep earthquake in the Solomon Islands and PS.PATS receiver from the Gomer and Okal (2003) Q_{ScS} study. To generate the suite of Q and wave speed model for the experiment, we modify either the Q or wave speeds in a specific depth range in 1D and/or 3D models. We increase Q by 0.5 – 5.0% in the attenuation model resulting in a “stiffer” medium and more effective energy propagation. Additionally, V_p and V_s were modified by ± 0.5 to 5% at the lowermost lower mantle at the core mantle boundary. The forward simulation calculates synthetic seismograms for the modified model and then the synthetics are compared against similar Q and wave speed models and reference waveforms from known models with no variation.

Model Selection and Modification

The one-dimensional PREM (Dziewonski and Anderson, 1981) and the three-dimensional QRFSI12 (Dalton et al., 2008) Q models in conjunction with the one-

dimensional transversely isotropic PREM (Dziewonski and Anderson, 1981) and three-dimensional S3620iso wave speed model (Kustowski et al., 2008) were chosen for the analysis. These models were selected, in part, because they were already available in the SPECFEM3D_GLOBE software and the Q and wave speed were setup to work with each other when specified; a three-dimensional Q model was incompatible with a one-dimensional wave speed model while the other combinations worked well. Additionally, the selected wave speed models are isotropic which will eliminate anisotropic effects and allow the analysis to focus on how the Q_{ScS} strictly related to isotropic wave speeds and Q. Each Q model was increased by 0.5, 1.0, 2.5, or 5% at a specific depth interval prior to making the model mesh while the wave speed models were modified ± 0.5 , 1.0, 2.5, or 5% at just above the CMB at 2741-2891 km. For each 1D model or 1D reference model, the Q values are modified by the percent increase at the depth interval. The 3D model is calculated from the 1D reference and a perturbation model. All tested models included spatial variations and physics associated with Crust2.0 (Bassin et al., 2000), topography, oceans, ellipticity, gravity, rotation, and obviously attenuation. No noise was incorporated in the resulting synthetics.

Gridding the Mesh

Before performing a simulation in SPECFEM3D_GLOBE, one must define the model mesh parameters and grid the input model. The mesh is considered a “cubed-sphere” which is mapped from the cube to the sphere and honors first and second order discontinuities in PREM (Komatitsch et al., 2002). Additionally, it accommodates crustal thickness variation and topography (Komatitsch et al., 2002).

As we were only using a single earthquake with a source ~ 10 degrees from the seismic station, we use one of the six regional chunks that make up the cubed sphere with a model 60° wide, 60° long, centered on 0° latitude, 165° longitude and with the depth extending to the CMB. The chunk is sliced into elements made up of hexahedral volume elements that align with neighboring elements and that double at increasing depths (Komatitsch et al., 2002; Komatitsch and Tromp, 2002a; Tromp et al., 2008). Each hexahedral volume element is made up of control points which are mapped between the element and the cube-sphere where each volume element must contain a minimum of 8 anchors but can be as large as 27 (Komatitsch and Tromp 1999). In addition to volume elements, there are also boundary elements to consider. Boundary elements are the volume elements along the edges of the model and free surfaces where an anchor within an absorbing boundary is related to the referenced cube-square (Komatitsch and Tromp, 1999).

We utilize 320 spectral elements along each side of the chunk. The shortest period which can be accurately simulated is approximated by

$$\text{shortest period (s)} \cong (256/\text{NEX_XI}) \times \left(\frac{\text{Angular_XI_IN_DEGREES}}{90} \right) \times 17$$

where NEX_XI is the number of spectral elements along one side (Komatitsch and Tromp 2002a). In this study we can accurately simulate periods to 9s. Additionally, the number of slices within our chunk is determined by the number of processors used. We utilized one node on the Bluewaves computing cluster consisting of 20 processors where each processor shares the calculations along its edges with the adjoining slices (Tromp et al., 2008).

Synthetic Waveform Generation

Synthetic seismograms were calculated by a three-dimensional wave simulation from a deep earthquake source (Gomer and Okal, 2003) to receivers using a spectral-element method (SEM) (Komatitsch et al., 2002). The SEM utilizes the adaptability of the finite-element method and the accuracy of the pseudospectral method which simplifies operations and reduces computational costs (Komatitsch et al., 2002). The SEM in SPECFEM3D_GLOBE utilizes the weak form of the wave equation (Komatitsch and Tromp, 1999), which facilitates the accurate body and surface wave diffraction simulations and ensures minimal dispersion and anisotropy (Komatitsch et al., 2002). We perform the wavefield simulation for 60 minutes of simulation time and record the synthetic seismograms at a variety of stations around the OJP but focus on the seismic station PS.PATS which Gomer and Okal (2003) utilized for their Q_{ScS} analysis.

Once the synthetic waveforms are generated, we apply a low-pass filter with a corner at 25s and compare the synthetic ScS and ScS^2 arrivals with modified Q and wave speed models. We additionally compare the synthetic ScS and ScS^2 arrivals from modified models to reference waveforms from unmodified Q and wave speed models.

RESULTS

A total of 87 simulations were computed in SPECfem3D (e.g., Komatitsch and Tromp, 2002a, b; Tromp et al., 2008; Komatitsch et al., 2016) representing a reference model with increasing complexity of the wave speeds and attenuation structures. A1V1 refers to the baseline reference model with a one-dimensional transversely

isotropic PREM wave speed model (Dziewonski and Anderson, 1981) with CRUST2.0 (Bassin et al., 2000) and one-dimensional PREM attenuation. A1V3 is a baseline reference model utilizing the three-dimensional isotropic wave speed model s362 (Kustowski et al., 2008) with CRUST2.0 (Bassin et al., 2000) and one-dimensional PREM attenuation. The last baseline reference model, A3V3, employs the three-dimensional isotropic wave speed model s362 with CRUST2.0 and three-dimensional upper mantle attenuation model QFRSI12 (Dalton et al., 2008). For the simulations presented here, the 1D PREM and 3D QFRSI12 attenuation reference models were modified at discrete depth intervals; the crust (1-35 km depth, CR), upper mantle between the crust and 410 km discontinuity (36-410 km depth, UM), between the 410 and 660 discontinuity (410-660 km depth, TZ), lower mantle along the core-mantle boundary (2741 – 2891 km depth, LLM), and a distributed modification from the crust to the core-mantle boundary (1-2891 km depth) (Figure 3.2). Each attenuation depth interval was modified 0.5, 1.0, 2.5 or 5% and subsequently used in the waveform simulation producing 20 synthetic waveforms to compare against the unmodified reference simulations. Additionally, the wave speed structure was modified between ± 0.5 and 5% at the lowermost lower mantle depth range (2741 – 2891 km) while attenuation was unchanged. This generated an additional 8 wave speed modified synthetic seismograms for comparison. Each of the 28 modified synthetic seismograms per reference model, 20 with modifications to the attenuation structure, 8 with wave speed, have only one variable modified (depth interval with percent change) per waveform. The 511 km deep May 2nd, 1996 M6.6 Solomon Island event analyzed in Gomer and Okal (2003) and described by Dziewonski et al.

(1997) was used as the source with the recording at PS.PATS and an epicentral distance of 11.7° . The focal mechanism ($\phi = 250^\circ$, $\delta = 73^\circ$, $\lambda = -111^\circ$) generated an ScS phase with a small initial positive amplitude followed by a much larger negative amplitude (Figure 3.3).

Synthetics from the three reference models, A1V1, A1V3, and A3V3, were compared against the synthetics from modified models at the phases ScS and ScSScS (ScS²) specifically looking at the differences in arrival and amplitude. A1V1 ScS arrived at 836.793s with an amplitude $-1.122e^{-05}$ while A1V3 ScS arrived later at 838.272s ($\Delta t = 1.479$ s) with an amplitude of $-1.1e^{-05}$ ($d\ln A = -2.0\%$) and A3V3 ScS arrived similarly at 838.467s ($\Delta t = 1.674$ s) with the largest amplitude of the three at $-1.51e^{-05}$ ($d\ln A = 34.5\%$) (Figure 3.3). For comparison, the theoretical ScS arrival using 1D PREM is 837.07s ($\Delta t = 0.277$ s). For the reference models, the ScS phase are similar but with A3V3 produces larger amplitudes. For ScS², A1V1 arrives at 1763.93s with an amplitude of $-4.94e^{-06}$, A1V3 arrives at 1767.53s ($\Delta t = 3.560$ s) with an amplitude of $-4.96e^{-06}$ ($d\ln A = 0.4\%$), and A3V3 arrives at 1769.86s ($\Delta t = 5.930$ s) with an amplitude of $-6.15e^{-06}$ ($d\ln A = 24.5\%$) (Figure 3.3). The 1D PREM theoretical ScS² arrival is later than A1V1 by 2.5s but faster than A1V3 and A3V3 at 1766.434s. Though the reference synthetics ScS phase arrivals are consistent with ScS phase arrivals from other methods, the raw and filtered waveforms appear “ringy” (Figure 3.4).

Irregularities in the reference and modified waveforms resulted in noisy and ringy features. This negatively impacted the analysis such that the modified waveforms for

each reference model are compared to each other and against the reference model in the appendix.

DISCUSSION

Each of the three models tested generated different seismograms. Additionally, the reference seismograms (Figure 3.3) arrived at different times with varying amplitudes. A3V3 had the largest ScS and ScS² amplitudes while also arriving last with the most complex models. A1V1 arrived earliest due to its simple velocity model while A1V3 arrived less than 0.2s earlier than A3V3 but having amplitudes remarkably similar to A1V1. Models with the same wave speed model but different Q background affect the measurement of different arrival times (Zhou 2009, Savage et al., 2010; Ruan and Zhou 2010). Additionally, Chaves and Ritsema (2016) demonstrated that different Q models chosen for the deep mantle can vary the arrival of ScS and ScS² by as much as 2s. The ScS and ScS² arrival difference between the references A1V3 and A3V3 is due to the Q heterogeneity in A3V3 between 0-650 km (Dalton et al., 2008); below 650 km, both Q models are identically 1D (Figure 3.2). The A1V1 and A1V3 amplitudes differences are exclusively related to the wave speed model where A1V1 used PREM (Dziewonski and Anderson, 1981) and A1V3 used s362iso (Kustowski et al., 2008). Similarly, the amplitude difference between A1V3 and A3V3 is singularly due to the attenuation model; A1V3 used PREM attenuation (reference) and A3V3 used QRFSI12 (Dalton et al., 2008).

The waveforms resulting from modifying the attenuation structure did not generate varying seismograms for all of the depth ranges that were tested. At each of

the five depth ranges, 4 attenuation percentages were used but resulted in waveforms that would be indistinguishable in real low-noise data. The attenuation modified waveforms of A1V1 had larger ScS and ScS² amplitudes, A1V3 had very similar amplitudes as the reference, and A3V3 had smaller amplitudes. Similarly, A1V1 and A1V3 had similar noise around the ScS phases while A3V3 was significantly noisier at the same ScS phase. Even though the simulations were supposedly noise free, signal to noise decreases with each ScS phase. The waveforms of A1V1 as expected, as Q increases (attenuation decreases), amplitudes should increase. This also resulted in higher amplitude noise around the ScS phases. The attenuation modified waveforms of A1V3 overlay the reference seismograms which suggests that an attenuation increase between 0.5 and 5.0% at different depth ranges is indiscernible from the effects of a 3D wave speed model. A3V3 waveforms with a modified decrease in attenuation structure with amplitudes less than the reference for ScS and ScS² compared to the A1V1, which is greater than the reference. This unexpected amplitude decrease may be due to how the Q increase was implemented in the 3D attenuation model or an issue with the model implementation. Another oddity in A3V3 is that the modified waveforms are less noisy or ringy than the reference (Appendix 3.8) which is antithetical to the other models. While modifying a 1D attenuation model is straightforward, QRFSI12 has a reference with spherical harmonics to degree 12 and eight radial spline functions (Dalton et al., 2008). To modify QRFSI12, only the reference model was modified which, when combined with the radial splines and implemented into the model may have resulted in an overall lower Q. Regardless of how any changes in Q in the modified QRFSI12 in the upper

mantle resulted in lower amplitudes than the reference, this does not explain the amplitudes resulting from distributed modifications to attenuation from the CMB to the crust and only in the lowermost mantle. QFRSI12 is a global upper mantle attenuation model (Dalton et al., 2008) and any modified Q below 650 km depth is not affected by the radial splines or spherical harmonics. This implies that with an increase of 0.5 to 5.0% in Q in the various depth ranges within a 3D attenuation and wave speed model, the amplitudes are decreased consistently. Additionally, we tested applying changes in Q directly to the 3D attenuation model on implementation which resulted in similar results as above.

The amplitude differences between each modification and reference are indistinguishably small. In A1V1, which is the simplest Earth structure tested, ScS amplitudes from Q modified depth ranges are near indistinguishable; the same is true for ScS2 (Appendix Table 3.3). With additional model complexity from the 3D velocity variations (A1V3), 1D Q modifications are muted by the 3D wave speeds with no distinguishable difference in ScS and ScS2 amplitudes. A3V3 also has significant ScS amplitude overlap with all Q modifications in the depth ranges. In each model structure, the various changes in Q in different depth ranges, ScS phase amplitudes overlap where analysis would result in ambiguity. If wave speed is not considered, any attempt to interpret attenuation structure utilizing ScS phases alone, is extremely difficult and fraught with inaccuracies.

Modifications to the wave speed in the lowermost mantle at the core-mantle boundary (CMB) resulted in more interesting amplitude results than those generated from attenuation changes. Changes in only the A1V1 wave speed up to $\pm 5.0\%$ in the

lower most mantle resulted in an amplitude increase (Appendix Figure 3.9). The amplitude increases due to the wave speed changes are slightly less than those associated with attenuation modifications that encompass the overall range. Wave speed perturbations of 2.5 and 5.0% less than the reference also resulted in an amplitude increase but less so than the smaller wave speed decrease and all wave speed increases. In A1V3 wave speed modified waveforms, no ScS or ScS² amplitude increase is identified similar to that of the attenuation modifications, only arrivals are changed and those are predicated on wave speed increases arriving earlier and wave speed decreases arriving later than the reference. Similar to the attenuation modifications, A3V3 wave speed modified waveforms similarly have an amplitude decrease compared to the reference. ScS and ScS² amplitudes from wave speed modifications overlap with every depth interval modification to the Q structure. Amplitudes for the ScS phases, unless utilizing other datasets, could be inferred to be from attenuation structure in any depth or multiple depth ranges or an unaccounted wave speed anomaly.

In the problem that we setup to solve in this chapter, investigating the effects of wave speed heterogeneity along the ScS wave propagation path and Q_{ScS} assumptions, abnormalities in the expected results included noisy and ringy characteristics in the resulting synthetics. The high magnitude of noise in the reference and modified synthetics is unusual as the synthetics were simulated without noise and it negatively impacts measurements of attenuation. When the 1D reference wave speed and Q model is compared to 1D synthetics like AxiSEM (Figure 3.4), the fluctuations in the amplitudes are obvious. Several possible sources of this “ringiness” are possible. The

oceans and topography were included in the SPECFEM simulations. The oceans are not included directly due to computational limitations, but as an additional downward force to properly simulate the surface waves; this is unlikely to have added additional noise in the simulations. Topography and bathymetry were also included in the simulations and it will generate secondary wave scattering, but at the periods used in the simulations, >25 seconds, this scattering should be minimal. The model domain was set at $60^\circ \times 60^\circ$ using a single chunk of the SPECFEM global model (Komatitsch and Tromp, 2002a). While the grid dimensions were calculated to resolve periods of 9s, it is possible that the single chunk was not sufficiently large to avoid reflections from the absorbing boundary conditions (Komatitsch and Tromp, 1999).

SPECFEM3D uses an absorbing boundary condition that works perfectly perpendicular to the boundary but much less effectively at grazing angles. These potential causes of unwanted noise can be thoroughly tested by modifying the size of the grid to a single chunk. If the noise is reduced or eliminated this would be indicative of a solution while if the noise moves out with time this would imply an issue with the absorbing boundaries. Besides the lateral boundaries, the lower boundary at the core may also be a source of noise if the absorbing boundary has similar issues as described above or if the core is not treated properly. The noisy records do not appear to become ringy until after the initial P wave arrival, which is well before any core phase arrivals. The azimuth between the simulated earthquake does not occur on or near a nodal plane which could have been problematic due to abrupt changes in the radiation pattern. The CMT solution used a half-duration of zero as recommended and any noise generated in the simulation should filter out with

the low-pass filter applied to the waveforms (Komatitsch and Tromp 2002a).

Additionally, source mechanisms that are less complicated can and will be tested to determine their contribution to the noise. Overall, the potential sources of error in the simulation overwhelmed our ability to generate high quality synthetic seismograms for analysis and measure Q with reliability and accuracy.

The analysis we proposed in this study began with a critical analysis of Gomer and Okal (2003) which uses spectral amplitude ratios (Sipkin and Jordan, 1980) to determine Q from multiple ScS phases. As the issues with the synthetic seismograms made it overly burdensome to qualitatively investigate Q_{ScS} changes, we generated several 1D seismograms (Al-Attar and Woodhouse, 2008; van Driel and Nissen-Meyer, 2014) with calculated ScS phase time differentials and spectral ratios accounting for geometrical spreading (Figure 3.5). For each 1D synthetic, PREM Q (Dziewonski and Anderson, 1981) was modified between 40 and 300 in the upper mantle (80-220 km and 80-400 km). The spectral amplitude ratios between ScS and ScS² phases show that there is little sensitivity to Q changes in the upper mantle. The differences in spectral amplitude ratios for an upper mantle Q of 200 and 400 are very small with geometrical spreading accounting for most of the amplitude difference between ScS and ScS² (Figure 3.5). Measurements for Q from synthetics from a model with an upper mantle with a Q of 40 (strongly attenuating from 80 – 400 km) resulted in minimal amplitude loss between ScS and ScS² after accounting for geometrical spreading, and insufficient to recover the prescribed Q in the upper mantle. This suggests that Q_{ScS} measurements using either spectral amplitude ratios (Sipkin and Jordan, 1980) or the time domain (Kanamori and Rivera, 2015) are difficult and

complicated to interpret. The original Q_{ScS} relied on stacking multiple earthquakes at arrays of stations to enhance the contrasts (Sipkin and Jordan, 1980).

In the reference study by Gomer and Okal (2003), the authors acknowledge the lack of lower mantle attenuation information but still attribute a high Q differential to the upper mantle at OJP due to the ray path bounce points. Global and regional attenuation models exhibit variable Q compared to Gomer and Okal (2003) and Suetsugu et al. (2019). The results of the 1D spectral amplitude ratios and others where noise free ScS and ScS^2 are modeled, provide a critical dataset when interpreting core reflected waveforms like those in Gomer and Okal (2003). Our results indicate that where modifications in attenuation, regardless of attenuation changes in the upper mantle, the resulting ScS and ScS^2 amplitude variations are small after accounting for geometrical spreading. Chaves and Ritsema (2016) have shown that the effects of focusing may overwhelm any effects from attenuation and attenuation may not be necessary to elucidate the waveforms. Likewise, Shearer (2015) has shown that wave scattering and attenuation may have a proportionate effect on the waveforms. As previously mentioned, and expected theoretically, arrival time of seismic phases, including ScS , depend on the Q structure of the background model (Zhou 2009; Savage et al., 2010; Ruan and Zhou 2010). More recently, heterogeneity in the lower mantle, specifically near the CMB has disproven the previous assumption of the lower mantle being benign and homogeneous. Lastly, sensitivity kernels associated with long period ScS and ScS^2 are extraordinarily large with complicated sensitivity at the CMB (Liu and Trump, 2008). When a bounce point with a large sensitivity size is near a plate boundary due to the large sensitivity size, like that near

the OJP, the horizontal contrasts can affect the waveform amplitudes (e.g., Lay and Kanamori, 1985; Ji et al., 2005). The above demonstrates the difficulty utilizing ScS and ScS² waveforms and attributing Q to a specific depth range, lateral location, and ignoring any wave speed attributed Q.

While we could not calculate Q_{ScS} differences from our synthetic seismograms using SPECFEM, several assumptions were made in our analysis. Our study ignored any anisotropy where the anisotropy is path-integrated along ScS and the subsequent derivatives would consist of significant anisotropy depending on the location. Only a single variable is changed at a time, wave speed or attenuation, in the three reference models due to the difficulty in distinguishing which variable impacted the amplitudes (Appendix Figure 3.12). Similar to the results of Chaves and Ritsema (2016), our results, though questionable, show that variations in Q can also affect the wave speed; this is expected theoretically. Any future work will have to address the underlying issues regarding the SPECFEM simulations including significantly larger $\pm$ percent changes in Q and utilizing a reference model with 3D attenuation and 1D wave speed to further investigate how Q can affect arrival measurements. Lastly, future work should include a comprehensive dissection of QRFSI12 and S362iso so that Q and wave speed are appropriately investigated.

CONCLUSIONS

Utilizing the deep earthquake used by Gomer and Okal (2003), we generated synthetic waveforms of ScS and ScS² phases in SpecFEM_GLOBE, a 3D spectral element wave propagation program in three increasingly complex wave speed and Q

model sets to test how Q affects the amplitudes of ScS phases. For each simulation and subsequent forward model, we modified the Q models at one of five depth ranges to varying degrees or modified the wave speed at the lowermost lower mantle.

Regardless of depth range, modifications in attenuation or wave speed for a specific model resulted in similar ScS and ScS² amplitudes and waveform arrivals but different from the baseline reference. Unfortunately, the reference and modified waveforms suffered from unwanted noise from unknown sources in the simulation making Q_{ScS} analysis and interpretation untenable.

Arrival and amplitudes varied depending on which wave speed and Q model was used. The ScS and ScS² arrival for the same 3D wave speed model arrived later with larger amplitudes when a 3D Q model was used instead of 1D while the amplitudes did not change between 1D and 3D wave speed models. This was previously demonstrated by Zhou (2009), Savage et al. (2010), Ruan and Zhou (2010) and most recently Chaves and Ritsema (2016). While Gomer and Okal (2003) calculated Q in the upper mantle using only a single receiver and earthquake, 1D synthetics of varying Q in the upper mantle show that spectral ratios and time domain methodologies, at least to single measurements, are minimally to faintly sensitive to these changes. Our observations show that depending on the model and assumptions used in the Q_{ScS} methodology, distinguishing amplitude changes associated with Q or wave speed is incredibly difficult and constraining those changes to a specific depth is equally challenging. In light of these results, we suggest that any study utilizing ScS phases as their primary dataset be interpreted with wariness while any future work should proceed with caution.

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Velocity Model	Attenuation Model	Name	Crust Model	Depth Range (km)	Type	Perc %
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	1-35	Attenuation	+5%
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	36-410	Attenuation	+5%
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	411-660	Attenuation	+5%
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	2741-2891	Attenuation	+5%
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	1-2891	Attenuation	+5%
S362iso	1D PREM	A1V3	Crust 2.0	1-35	Attenuation	+5%
S362iso	1D PREM	A1V3	Crust 2.0	36-410	Attenuation	+5%
S362iso	1D PREM	A1V3	Crust 2.0	411-660	Attenuation	+5%
S362iso	1D PREM	A1V3	Crust 2.0	2741-2891	Attenuation	+5%
S362iso	1D PREM	A1V3	Crust 2.0	1-2891	Attenuation	+5%

Table 3.1. ScS Model tests

Velocity Model	Attenuation Model	Name	Crust Model	Depth Range (km)	Type	Perc %
S362iso	QRFSII12	A3V3	Crust 2.0	1-35	Attenuation	+5%
S362iso	QRFSII12	A3V3	Crust 2.0	36-410	Attenuation	+5%
S362iso	QRFSII12	A3V3	Crust 2.0	411-660	Attenuation	+5%
S362iso	QRFSII12	A3V3	Crust 2.0	2741-2891	Attenuation	+5%
S362iso	QRFSII12	A3V3	Crust 2.0	1-2891	Attenuation	+5%
Transverse isotropic PREM	1D PREM	A1V1	Crust 2.0	2741-2891	Wave speed	+/- 5%
S362iso	1D PREM	A1V3	Crust 2.0	2741-2891	Wave speed	+/- 5%
S362iso	QRFSII12	A3V3	Crust 2.0	2741-2891	Wave speed	+/- 5%

Table 3.1. continued

Figure 3.1: Map of the Ontong Java Plateau (OJP) including the earthquake event (yellow star) and seismic station (red inverted triangle). Multiple recorded and synthetic ScS phases with bounce points recorded at PS.PATS are shown as wide diamond (ScS1) and narrow diamonds (ScS2). ScS3 are not displayed).

Figure 3.2: 1D velocity (Vp-red, Vs-blue) and attenuation (green) reference/starting models. Colored dashed lines are reference velocity and attenuation profiles for 3D models that utilize splines. Colored solid lines are the 1D velocity and attenuation profiles. Areas of depth specific model perturbations are outlined in dashed black and highlighted in red; ‘CR’ crust (1-35 km), ‘UM’ upper mantle (36-410 km), ‘TZ’ transition zone (410-660 km), ‘LLM’ lowermost lower mantle (2741-2891 km).

Figure 3.3: Synthetic ScS, ScS², and ScS³ from 1D (V1) and 3D (V3) velocity and 1D (A1) and 3D (A3) attenuation reference models generated in SpecFEM3D_Globe with Crust2.0.

Figure 3.4: Comparison of 1D synthetic waveforms using yspec (black), SPECfem (SEM) (red), and AxiSEM (blue) for the Gomer and Okal (2003) deep earthquake.

Figure 3.5: (a) 1D Qmu models. (b-e) Calculated ScS phase time differentials and spectral ratios of the modified 1D Qmu models; (b) 300 Q between 80-220 km; (c) 40 between 80-220 km; (d) 300 between 80-400 km; (e) 40 between 80-400 km.

C. 40 Q between 80-220 km

D. 300 Q between 80-400 km

Figure 3.5 continued

D. 40 Q between 80-400 km

Figure 3.5 continued

APPENDIX

Included in the appendix is a description of the results of the Q and wave speed modified waveforms generated in SPECFEM and additional figures and tables.

Table 3.2: ScS arrival and amplitude measurement differentials from reference models

Table 3.3: ScS² arrival and amplitude measurement differentials from reference

Figure 3.6. (red) 1D velocity and 1D attenuation starting/reference model synthetic waveforms (green waveforms from Figure 3.3) vs. 1D velocity and 1D attenuation starting model synthetic waveforms but with individual isolated attenuation perturbations in the crust, upper mantle, transition zone, and core-mantle boundary between 0.5 and 5% (multi-colored). All attenuation perturbed waveforms completely overlap at this scale. Minor waveform amplitude differences and delays exist at very fine scales. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.7. (red) 3D velocity and 1D attenuation starting/reference model synthetic waveforms (blue waveforms from Figure 3.3) vs. 3D velocity and 1D attenuation starting model synthetic waveforms but with individual isolated attenuation perturbations in the crust, upper mantle, transition zone, and core-mantle boundary between 0.5 and 5% (multi-colored). All attenuation perturbed waveforms completely overlap each other and the reference waveform at this scale. Minor waveform amplitude differences and delays exist at very fine scales. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.8. (red) 3D velocity and 3D attenuation starting/reference model synthetic waveforms (red waveforms from Figure 3.3) vs. 3D velocity and 3D attenuation starting model synthetic waveforms but with individual isolated attenuation perturbations in the crust, upper mantle, transition zone, and core-mantle boundary between 0.5 and 5% (multi-colored). All attenuation perturbed waveforms completely overlap each other at this scale. Minor waveform amplitude differences and delays exist at very fine scales. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.9. (black) 1D velocity and 1D attenuation starting/reference model synthetic waveforms (green waveforms from Figure 3.3) vs. 1D velocity and 1D attenuation starting model synthetic waveforms but with individual isolated velocity perturbations in the core-mantle boundary between positive (solid colored) and negative (dashed colored) 0.5 and 5%. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.10. (black) 3D velocity and 1D attenuation starting/reference model synthetic waveforms (blue waveforms from Figure 3.3) vs. 3D velocity and 1D attenuation starting model synthetic waveforms but with individual isolated velocity perturbations in the core-mantle boundary between positive (solid colored) and negative (dashed colored) 0.5 and 5%). Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.11. (black) 3D velocity and 3D attenuation starting/reference model synthetic waveforms (red waveforms from Figure 3.3) vs. 3D velocity and 3D attenuation starting model synthetic waveforms but with individual isolated velocity perturbations in the core-mantle boundary between positive (solid colored) and negative (dashed colored) 0.5 and 5%. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Figure 3.12. (black) 1D velocity and 1D attenuation starting/reference model synthetic waveforms (red waveforms from Figure 3.3) vs. all tested perturbed 1D velocity and 1D attenuation synthetic waveforms. Velocity perturbations at the core-mantle boundary waveforms (dashed colored) and attenuation perturbations in the crust, upper mantle, mantle transition zone, and core-mantle boundary (solid colored) are highlighted. Waveforms are windowed over the predicted arrivals for ScS, ScS₂, and ScS₃ and low-passed filtered at 25 seconds.

Results

Attenuation

A1V1

The waveforms, generated from a modified (0.5 – 5.0%) attenuation structure distributed from the surface to the CMB or in the crust (CR), upper mantle (UM), transition zone (TZ), and lowermost lower mantle (LLM), are near identical, excluding the reference waveform (Figure 3.6). Similar to the reference synthetics generated from 1D wave speed and attenuation models, the modified waveforms also appear noisy. The arrival of ScS for all modified waveforms is 837.63s, 0.837s slower

than the reference. The mean amplitude for the modified waveforms is $-1.4288e^{-05}$ ($d\ln A=27.3\%$) with a standard deviation of $7.955e^{-10}$ ($3.07e^{-06}$ greater than reference). Though the ScS and ScS² amplitudes overlap with each depth interval, upon close inspection, Q changes dispersed thru the crust and mantle creates the largest amplitude increase followed closely by those changes in Q in the TZ and to a lesser degree, the crust, lowermost mantle, and lastly the upper mantle (Table 3.2). Most changes in Q in CR, UM, and LLM result in similar amplitudes as well as small Q modifications in the dispersed case and TZ. The ScS² phase arrival for all modified waveforms is 1766.82, 2.89s slower than the reference. Like ScS, the mean ScS² amplitude is larger than the reference at $-6.787e^{-06}$ ($d\ln A=37.4\%$) with a standard deviation of $1.022e^{-09}$. Modifications in dispersed Q and UM have similar amplitudes across most changes and TZ lesser so while Q changes to CR and LLM result in smaller amplitude spread but equally similar (Table 3.3). These amplitude differences for ScS and ScS² are only observed in the fine details of the “noise free” synthetic waveforms and would likely be impossible to distinguish in real seismograms.

A1V3

Attenuation modifications were simulated using the same weights and same depth ranges as in A1V1. The only difference between the models is A1V3 has a 3D wave speed. The various ScS phase arrival waveforms from the modified attenuation structure are near identical to each other and to the reference waveforms including a consistent ringiness. The arrival of ScS for all modified waveforms is 838.272s, identical to the reference. The mean amplitude for the modified waveforms is $-1.103e^{-05}$ ($d\ln A=-1.7\%$) with a standard deviation of $2.36e^{-11}$ ($3.3258e^{-11}$ greater than

reference). The fine scale amplitude differences between the reference and the depth interval Q modified waveforms is exceedingly small with significant amplitude overlap between each (Table 3.2). The ScS² phase arrival for all modified waveforms is 1767.53, the same as the reference. The mean amplitude for the ScS2 modified waveforms is -4.956e-06 (dlnA=0.3%) with a standard deviation of 7.25e⁻¹¹ (7.72e⁻¹¹ greater than reference). Like the ScS amplitudes, there is significant amplitude magnitude overlap between each depth range but even more so (Table 3.3).

A3V3

As above, A3V3 uses the same weights and same depth ranges as in A1V1 and A1V3. Unlike A1V1 and A1V3, A3V3 modifies the 1D Q reference model used in generating the 3D attenuation model. The ScS phase arrivals arrive at near the exact same time as the reference waveforms though the amplitudes are smaller but similarly noisy. The arrival of the ScS phase for all modified waveforms is 838.348, 0.12s faster than the reference. The mean amplitude for the ScS phase of the modified waveforms is -1.102e⁻⁰⁵ (dlnA=-1.8%) with a standard deviation of 6.214e⁻¹⁰ (4.08e-6 less than the reference). ScS amplitudes associated with distributed Q and the TZ depth range while CR, UM, and LLM depth range amplitudes are fairly consistent between each group and each other regardless of modification (Table 3.2). The mean ScS² phase arrival for all modified waveforms is 1767.636s with a standard deviation of 0.047s (2.224s faster than the reference). The 5% distributed attenuation modification is the slowest ScS² at 1767.76s ($\Delta t=3.83s$) while the 2nd slowest ScS2 arrival are from the remaining distributed attenuation models and 5 and 2.5% for the transition zone and upper mantle modifications at 1767.68s ($\Delta t=3.75s$). The

remaining ScS2 attenuation modified waveforms arrive at 1767.6s ($\Delta t=3.67s$). The mean amplitude for the ScS2 phase of the modified waveforms is $-4.964e-6$ ($d\ln A=0.5\%$) with a standard deviation of $4.67e-10$ ($1.189e-6$ less than the reference. ScS2 amplitudes from the TZ modified depth range shows the largest variability while the remaining depth ranges are all around the mean with little variability (Table 3.3).

Wave speed

The wave speed modified waveforms produced more variability in the ScS and ScS² phase arrival and amplitude compared to the attenuation modified waveforms. Like the reference and attenuation modified seismograms, the wave speed modified waveforms also display this “ringy” characteristic. Modifications were made only to the Vp and Vs wave speed, between ± 0.5 to 5.0% in lowermost lower mantle near the CMB (Figure 3.2).

AIVI

The wave speed modified ScS and ScS2 phases arrive at a range of times around the reference model while the amplitudes are greater (Figure 3.9). The ScS waveforms with 5% and 2.5% increase in Vp and Vs in the lowermost mantle arrives earlier than the reference model $-1.141s$ and $-0.152s$ respectively while the remaining wave speed modified waveforms ($+1\%$, $+0.5\%$, -0.5% , -1% , -2.5% , and -5%) arrive after the reference. The amplitudes are less variable with a mean of $-1.427e-5$ ($d\ln A=27\%$) with a standard deviation of $3.27e-8$, $3.05e-6$ greater than the reference. Though the amplitudes between each model are very small, the amplitudes are largest with the most negative change (-5%) and least at the most positive ($+5\%$) (Table 3.2). This creates a range of amplitudes greater than those from a modified Q in any depth

range. The ScS² waveforms with a 5% increase in V_p and V_s in the lowermost mantle arrive just slightly earlier than the reference model at -0.23s while the remaining wave speed modified waveforms increasingly arrive after the reference based on the increasing wave speed percentage. Amplitudes, like the wave speed modified ScS phase arrivals, are fairly uniform with a mean of -6.592e^{-6} (dlnA=33%) with a standard deviation of 2.45e^{-7} which is 1.65e^{-6} larger than the reference. The ScS² waveforms with 5% and 2.5% decrease in wave speeds have noticeably smaller amplitudes than the other wave speed modified waveforms at -6.18e^{-6} (dlnA=25.1%) and -6.19e^{-6} (dlnA=25.3%) as compared to a mean of -6.727e^{-6} (Table 3.3). All wave speed modified ScS² amplitudes are slightly less negative compared to the Q modified amplitudes for all depth ranges but only on scales currently beyond our abilities to distinguish in real data.

AIV3

The ScS and ScS² waveforms from the wave speed modified simulations vary with percent lowermost mantle wave speed variation while the amplitudes remain consistent with small variations (Figure 3.10). All of the ScS waveforms with an increased wave speed arrive before the reference while those with a decrease in wave speed arrive after. The wave speed modified ScS amplitude are near identical with a mean of -1.1e^{-5} (dlnA=-2.0%) and a standard deviation of 6.59e^{-8} with the largest wave speed increase with a slightly smaller amplitude at -1.09e^{-5} (dlnA=-2.9%) and the largest wave speed decrease with a slightly larger amplitude increase at -1.11e^{-5} (dlnA=-1.1%) as compared to -1.1e^{-5} for the reference but in general, amplitudes decrease with increasing wave speed modification from -5% to 5% (Table 3.2). A

decrease in wave speed results in a slightly more negative ScS amplitude while an increase in wave speed results in a slightly less negative ScS amplitude. The ScS² waveforms result in comparable arrivals as above with an increased wave speed arrive before the reference while those with a decrease in wave speed arrive after. Similarly, the ScS² amplitudes with a largest wave speed increase result in a slightly smaller amplitude at -4.87×10^{-6} ($d\ln A = -1.4\%$) and largest wave speed decrease with a slightly larger amplitude at -5.1×10^{-6} ($d\ln A = 3.2\%$) as compared to -4.956×10^{-6} for the ScS² reference (Table 3.3). The mean ScS² amplitude is -4.965×10^{-6} with a standard deviation of 6.725×10^{-8} . ScS and ScS² amplitudes from wave speed modification are both larger and smaller than the Q modified depth range counterparts with wave speed modifications of $\pm 0.5\%$ in the lowermost lower mantle are most similar (Table 3.2; 3.3).

A3V3

The ScS and ScS² waveforms from the wave speed modified simulations arrivals vary with percent lowermost mantle wave speed variation while the amplitudes are smaller than the reference but comparable to each (Figure 3.11). Like in the A1V3 waveforms, the ScS phase arrival with wave speed increase arrive earlier than the reference ScS arrival while the waveforms with wave speed decreases arrive later than the reference ScS phase. The mean amplitudes of the wave speed modified ScS phase is -1.1×10^{-5} with a standard deviation of 6.65×10^{-8} which is smaller than the reference ScS of -1.51×10^{-5} . Like A1V3, the largest wave speed increase (+5%) results in smallest absolute amplitude as compared to reference while largest wave speed decrease produces a largest negative amplitude but still less than the reference (Table 3.2). The

ScS² wave speed arrivals are similar to the ScS with wave speed increases arriving before the reference and wave speed decreases arriving later. The amplitudes of the ScS² are similarly lower than the reference model with the wave speed modifications resulting in similar results as ScS. The mean amplitude of the wave speed modified ScS² phase is -4.97e^{-6} with a standard deviation of 5.95e^{-8} while the reference is 6.15e^{-6} .

⁶. Similar to the how the arrival compared to the wave speed increase, the amplitudes are smallest with the largest wave speed increase and largest with the largest wave speed decrease. Like A1V3, ScS and ScS2 amplitudes from wave speed modifications are more variable than the Q modified amplitudes for any depth range and are most similar at a wave speed change of $\pm 0.5\%$.

		Arrival			Amplitude		
		A1V1	A1V3	A3V3	A1V1	A1V3	A3V3
	Reference	836.793	838.272	838.467	-1.122E-05	-1.103E-05	-1.510E-05
Attenuation					1×10^{-6}	1×10^{-11}	1×10^{-6}
	Dist 0.5	-0.837	0	0.119	3.068075	-1.876840	-4.081424
	Dist 1.0	-0.837	0	0.119	3.068242	1.591350	-4.081065
	Dist 2.5	-0.837	0	0.119	3.069358	4.903060	-4.080193
	Dist 5.0	-0.837	0	0.119	3.070783	0.382980	-4.079000
	CR 0.5	-0.837	0	0.119	3.067855	1.682310	-4.081585
	CR 1.0	-0.837	0	0.119	3.067871	3.909110	-4.081456
	CR 2.5	-0.837	0	0.119	3.067974	3.603390	-4.081325
	CR 5.0	-0.837	0	0.119	3.068225	3.603390	-4.081041
	UM 0.5	-0.837	0	0.119	3.067770	4.597270	-4.081532
	UM 1.0	-0.837	0	0.119	3.068009	0.970460	-4.081410
	UM 2.5	-0.837	0	0.119	3.068271	5.805690	-4.081232
	UM 5.0	-0.837	0	0.119	3.068698	1.962690	-4.081203
	TZ 0.5	-0.837	0	0.119	3.068052	7.772040	-4.081491
	TZ 1.0	-0.837	0	0.119	3.068072	7.096150	-4.081316
	TZ 2.5	-0.837	0	0.119	3.068731	6.106510	-4.080888
	TZ 5.0	-0.837	0	0.119	3.069951	2.761750	-4.080271
	LLM 0.5	-0.837	0	0.119	3.067805	0.304500	-4.081575
	LLM 1.0	-0.837	0	0.119	3.067827	4.068460	-4.081551
	LLM 2.5	-0.837	0	0.119	3.067627	3.747830	-4.081526
	LLM 5.0	-0.837	0	0.119	3.067676	3.523880	-4.081461
		Arrival			Amplitude		
Wave Speed					1×10^{-06}	1×10^{-08}	1×10^{-06}
	LLM -5	-3.043	-0.228	-2.24	3.075576	11.168622	-3.969348
	LLM -2.5	-3.043	-0.456	-1.022	3.080014	5.718446	-4.023836
	LLM -1.0	-1.218	-1.141	-0.337	3.075714	2.316768	-4.058244
	LLM -0.5	-0.989	-2.359	-0.109	3.071974	1.164375	-4.069832
	LLM +0.5	-0.609	0.229	0.348	3.062694	-1.164733	-4.093361
	LLM +1.0	-0.381	0.457	0.576	3.056339	-2.333920	-4.105227
	LLM +2.5	0.152	1.066	1.261	3.032654	-5.890112	-4.141344
	LLM +5.0	1.141	2.131	2.25	3.975990	-11.90177	-4.202258

Table 3.2. ScS arrival and amplitude measurement differentials from reference models

		Arrival			Amplitude		
		A1V1	A1V3	A3V3	A1V1	A1V3	A3V3
	Reference	1763.93	1767.53	1769.86	-4.937E-06	-4.956E-06	-6.153E-06
Attenuation					1×10^{-6}	1×10^{-11}	1×10^{-6}
	Dist 0.5	-2.890	0.000	2.180	1.849485	6.192565	-1.188833
	Dist 1	-2.890	0.000	2.180	1.849477	9.650502	-1.188746
	Dist 2.5	-2.890	0.000	2.180	1.850641	15.195188	-1.188612
	Dist 5	-2.890	0.000	2.100	1.852716	-4.023033	-1.188585
	CR 0.5	-2.890	0.000	2.260	1.849324	-5.201045	-1.188986
	CR 1	-2.890	0.000	2.260	1.848951	12.365960	-1.188759
	CR 2.5	-2.890	0.000	2.260	1.849132	12.867855	-1.188598
	CR 5	-2.890	0.000	2.260	1.849297	12.867855	-1.188601
	UM 0.5	-2.890	0.000	2.260	1.849046	14.337225	-1.188863
	UM 1	-2.890	0.000	2.260	1.849686	3.306827	-1.188750
	UM 2.5	-2.890	0.000	2.180	1.850285	22.296680	-1.188810
	UM 5	-2.890	0.000	2.180	1.851361	2.980490	-1.188877
	TZ 0.5	-2.890	0.000	2.260	1.849572	5.578231	-1.189058
	TZ 1	-2.890	0.000	2.260	1.849844	16.994470	-1.189075
	TZ 2.5	-2.890	0.000	2.180	1.850108	12.248978	-1.189855
	TZ 5	-2.890	0.000	2.180	1.851542	-0.421418	-1.190591
	LLM 0.5	-2.890	0.000	2.260	1.849086	-1.154625	-1.188661
	LLM 1	-2.890	0.000	2.260	1.849455	3.707165	-1.188920
	LLM 2.5	-2.890	0.000	2.260	1.848791	11.621582	-1.188744
	LLM 5	-2.890	0.000	2.260	1.848472	2.904705	-1.188746
		Arrival			Amplitude		
Wave Speed					1×10^{-06}	1×10^{-08}	1×10^{-06}
	LLM -5	-6.080	-4.720	-2.460	1.246130	14.706509	-1.060655
	LLM -2.5	-6.080	-2.360	-0.100	1.254357	6.192184	-1.135276
	LLM -1	-3.500	-0.910	1.340	1.804138	2.539949	-1.166660
	LLM -0.5	-3.190	-0.450	1.800	1.832815	1.330114	-1.177165
	LLM +0.5	-2.580	0.460	2.710	1.852885	-1.331435	-1.201051
	LLM +1	-2.200	0.920	3.090	1.845709	-2.586294	-1.212191
	LLM +2.5	-1.290	2.210	4.460	1.795883	-5.256693	-1.236018
	LLM +5	0.230	4.340	6.520	1.604707	-8.333107	-1.265447

Table 3.3. ScS² arrival and amplitude measurement differentials from reference models

Figure 3.6: Synthetic output from the modified 1D attenuation and 1D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.7: Synthetic output from the modified1D attenuation and 3D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.8: Synthetic output from the modified 3D attenuation and 3D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.9: Synthetic output from the 1D attenuation and modified 1D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.10: Synthetic output from the 1D attenuation and modified 3D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.11: Synthetic output from the 3D attenuation and modified 3D velocity models vs the synthetic reference. Waveforms are low-pass filtered at 25 seconds.

Figure 3.12: Synthetic output from all 1D attenuation and 1D velocity model changes. Black(original), Velocity (dashed) vs Attenuation (solid)