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Seismic Structure near the Mariana Trench and Deep Earthquake Triggering in the Tonga Flat Slab

Chen Cai
Washington University in St. Louis

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WASHINGTON UNIVERSITY IN ST. LOUIS

Department of Earth and Planetary Sciences

Dissertation Examination Committee:
Douglas A. Wiens, Chair
James A. Conder
Michael J. Krawczynski
Philip Skemer
V. Slava Solomatov

Seismic Structure near the Mariana Trench and Deep Earthquake Triggering in the Tonga Flat Slab
by
Chen Cai

A dissertation presented to
The Graduate School
of Washington University in
partial fulfillment of the
requirements for the degree
of Doctor of Philosophy

August 2018
St. Louis, Missouri
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Acknowledgements

I would like to thank my advisor Douglas Wiens, who has been so supportive in my studies and provided me with great opportunities to participate in exciting scientific projects. His serious attitude, honesty, and enthusiasm for science have significantly influenced my research. I am grateful to all faculty members on my dissertation committee, Slava Solomatov, Phil Skemer, James Conder and Mike Krawczynski who volunteered their time in reading and commenting on this dissertation and gave me advice on various perspectives.

I want to thank Patrick Shore, an excellent mentor and also a great friend, who has helped me on almost everything. I have to thank all the members from Washington University seismology group and the department of Earth and Planetary Sciences for supporting me to do this PhD. I appreciate the friendship and help from Songqiao Wei, Weisen Shen, Martin Pratt, Andrew Lloyd, Melody Eimer, Aubreya Adams, Zhengyang Zhou, Tingying Xu, Zhen Tian, and many others.

Finally, I would like to thank my parents, parents in law and wife who have been continuously supportive of my education and career choice. They selflessly encouraged me to explore new directions in life. This journey would not have been possible if not for them, and I dedicate this milestone to them. And lastly, to Angela Cai, having you as a daughter gave my life so much meaning.

Chen Cai

Washington University in St. Louis
August 2018
This dissertation utilizes multiple techniques of earthquake location and seismic tomography to investigate deep seismicity in the Tonga flat slab region and the upper mantle structure of the Mariana subduction zone. I study the rupture property and aftershock distributions of the largest earthquake ever recorded in the Tonga flat slab region, defined by scattered seismicity and velocity anomalies. These, along with background seismicity, regional 3-D seismic velocity model and tectonic reconstructions of relict Vitiaz subduction zone, suggest the earthquakes are occurring in the final portion of the slab subducted at the now inactive Vitiaz trench. The associated Coulomb stress change shows many of the aftershocks were dynamically triggered, suggesting fossil slabs contain material that is too warm for earthquake nucleation, but may be near the critical stress susceptible to dynamic triggering.

In order to investigate water cycling at the Mariana subduction zone, I conduct tomographic study of Rayleigh wave to examine velocity and anisotropy anomalies associated with slab hydration. Tomographic results show a low velocity zone within the incoming plate prior to the trench down to about 30±5 km, along with significant trench-parallel anisotropy. The low
velocity zone preserves the thickness after the slab is subducted, but the velocity reductions become smaller. An extremely low velocity zone is observed beneath the serpentine seamounts in the outer forearc. These anomalies suggest coexistence of water-filled cracks and mantle serpentinization along the normal faults in the slab before subduction. Water is expelled from the cracks early in subduction, causing a modest increase in the velocity of the subducting mantle, and the water moves upward and causes serpentinization of the outer forearc mantle. The total amount of water carried by hydrous minerals in the Pacific plate at the Mariana trench is estimated from the low velocity zone thickness and associated velocity reduction. If other old, cold subducting slabs contain correspondingly thick layers of hydrous mantle, as suggested by the similarity of incoming plate faulting, estimates of the global water flux into the sub-arc mantle must be increased by about a factor of three over previous estimates.

I further expand the study of the Mariana subduction zone to the Mariana volcanic arc and backarc. I conduct similar tomographic study of Rayleigh wave to help constrain the melt production and transport beneath volcanic arc and backarc spreading center. Tomographic results show a thick deep low velocity zone (LVZ) beneath the Mariana trough with varying depth. A small size slow velocity anomaly beneath the volcanic arc near the top of the mantle is also imaged. An inclined LVZ is observed west of the arc and connected to the deep LVZ. Tomographic results also show a shallower thin slow velocity layer at the top of the mantle distributed in a narrow channel along the central part of the spreading center, where larger magma supply is suggested. The shallow slow velocity anomaly beneath the volcanic arc and the backarc spreading center may both represent a shallow melt reservoir. However, neither of them is imaged to be directly connected to the deep LVZ, suggesting the melt may be transported through a conduit that is too narrow to be resolved by surface wave tomography.
Chapter 1: Introduction

Subduction zone is a key component of Earth’s dynamic system and essential for understanding the processes that control the dynamic evolution of the mantle. These processes include slab dynamics before and after subduction, water circulation, arc magmatism, and the formation of new oceanic lithosphere at backarc spreading centers. Despite its obvious importance, studies of subduction zone structure and evolution are still not sufficient. Thanks to the development of ocean bottom seismographs, more and more subduction zones are now covered by temporary ocean bottom seismographs network. This dissertation covers seismic studies from two subduction zones, employing recently developed methods and techniques. The first section of the dissertation focuses the Tonga subduction zone, studying deep earthquakes in a region west of the Tonga slab defined by scattered seismicity and velocity anomalies. The second section centers on the Mariana subduction zone, with an investigation of the seismic properties of the upper mantle.

Chapter 2 focuses on deep seismicity in the Tonga flat slab region. The Tonga flat slab is defined by a small group of deep earthquakes distributed outside of the Tonga Wadati-Benioff Zone (WBZ) beneath the North Fiji Basin [Hamburger and Isacks, 1987; Okal and Kirby, 1998; Chen and Brudzinski, 2001], extending westward as far as 700 km. These unusual earthquakes, often referred to as ‘outboard’ events, are commonly thought to occur within a detached slab segment [Hamburger and Isacks, 1987; Okal and Kirby, 1998; Chen and Brudzinski, 2001], as indicated by seismic tomography [van der Hilst, 1995; Amaru, 2007; Simmons et al., 2012; Fukao and Obayashi, 2013]. However, the origin of this detached slab still remains unsolved. The sparse seismicity and absence of large magnitude earthquakes with aftershocks have
prevented a thorough study of the outboard earthquake characteristics as well as the associated slab geometry. We study source property and aftershock distributions of the largest earthquake ever recorded in this region, using data from a temporary seismic network deployed on the Fiji Islands. This aftershock sequence provides important insight into the mode of seismic nucleation in previously subducted materials, as well as the subduction history and the deep slab geometry in this region.

All following chapters focus on the Mariana subduction zone. The Mariana subduction zone provides an excellent setting to evaluate subduction zone processes because it includes a wide variety of tectonic features, including an extremely old subducting plate with widely spreading normal faulting [Oakley et al., 2008; Emry et al., 2014], a highly faulted forearc with actively venting serpentinite seamounts [Fryer et al., 1985, 1999], a remnant arc, an active modern volcanic arc and back-arc spreading center [Stern et al., 2003]. The intraoceanic environment also prevents contamination of continental crust for geochemical studies of arc and backarc magmatism. At the trench, the old Pacific plate (> 150 Ma) subducts beneath the Philippine Sea plate with a direction of N80°W and a rate of ~4.5 cm/yr in the region of this study [Seno et al., 1993]. The subducted slab becomes nearly vertical below depths of about 250 km [Engdahl et al., 1998] and no slab rollback is observed [Stern et al., 2003; Schellart et al., 2007]. The spreading rates vary along the back-arc spreading center, but all are very slow (~25 mm/yr at 18°N) [Seama and Fujiwara, 1993; Kato et al., 2003]. Geochemical studies have shown slab-derived volatiles and fluid mobile elements at the forearc, arc and the backarc along with variation along the arc chain and the backarc axis [Stolper and Newman, 1994; Newman et al., 2000; Taylor and Martinez, 2003; Pearce et al., 2005; Kelley et al., 2006; Barnes et al., 2008].
Two separate temporary seismic arrays were deployed in the Mariana subduction zone. The 2003-2004 Mariana Subduction Factory Imaging Experiment, which consists of 20 broadband land stations and 58 semi-broadband ocean bottom seismographs, densely covers the active backarc spreading center and sparsely extends to the forearc and the West Mariana Ridge. A more recent deployment between 2012-2013, consists of 20 ocean bottom seismographs and 7 island stations, covers the subducting Pacific plate, the Mariana forearc and volcanic arc.

In Chapter 3, we apply the ambient noise tomography method and the two-plane-wave tomography method to the data from the 2012-2013 deployment. A Bayesian Monte-Carlo algorithm [Shen et al., 2013] is then adopted to get SV-velocity models near the central Mariana trench. These strategies significantly improve the tomographic resolution at shallow depths, so that we are able to construct a comprehensive model of the uppermost mantle. We then analyze the results and estimate the water volume carried by the subducting Pacific plate at the Mariana subduction zone and apply this estimation to global subduction zones. In Chapter 4, we deliver thorough discussions of the results from Chapter 3 and develop a water cycling model for the Mariana subduction zone.

In Chapter 5, we combine data collected by the 2003-2004 deployment and the 2012-2013 deployment. We apply the ambient noise tomography and the two-plane-wave tomography for SV-velocity with the Monte-Carlo inversion, mainly focusing on structure around the Mariana forearc, arc and backarc. This model, along with previous studies, helps us better understand the melt production and transport beneath the Mariana volcanic arc and back-arc spreading center.
References


Chapter 2: Dynamic Triggering of Deep Earthquakes within a Fossil Slab

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2.1 Abstract

The 9 November 2009 Mw 7.3 Fiji deep earthquake is the largest event in a region west of the Tonga slab defined by scattered seismicity and velocity anomalies. The mainshock rupture was compact, but the aftershocks were distributed along a linear feature at distances of up to 126 km. The aftershocks and some background seismicity define a sharp northern boundary to the zone of outboard earthquakes, extending westward towards the Vitiaz deep earthquake cluster. The northern earthquake lineament is geometrically similar to tectonic reconstructions of the relict Vitiaz subduction zone at 8-10 Ma, suggesting the earthquakes are occurring in the final portion of the slab subducted at the now inactive Vitiaz trench. A Coulomb stress change calculation suggests many of the aftershocks were dynamically triggered. We propose that fossil slabs contain material that is too warm for earthquake nucleation, but may be near the critical stress susceptible to dynamic triggering.
2.2 Introduction

The Tonga subduction zone has long been identified as having more deep earthquakes than any other region in the world. Most of these deep earthquakes occur within the continuous slab currently subducting along the Tonga trench, while a much smaller group is sparsely distributed outside of the Tonga Wadati-Benioff Zone (WBZ) beneath the North Fiji Basin [Hamburger and Isacks, 1987; Okal and Kirby, 1998; Chen and Brudzinski, 2001], extending westward as far as 700 km (Figure 2.1a). These unusual earthquakes, often referred to as ‘outboard’ events, are commonly thought to occur within a detached slab segment [Hamburger and Isacks, 1987; Okal and Kirby, 1998; Chen and Brudzinski, 2001], as indicated by seismic tomography [van der Hilst, 1995; Amaru, 2007; Simmons et al., 2012; Fukao and Obayashi, 2013]. However, the origin of this detached slab still remains unsolved. The sparse seismicity and absence of large magnitude earthquakes with aftershocks have prevented a thorough study of the outboard earthquake characteristics as well as the associated slab geometry.

The 9 November 2009 (M\text{\text{w}}=7.3; depth=591 km) earthquake beneath the Fiji Islands was the largest ‘outboard’ earthquake ever recorded in the Tonga region. This earthquake occurred in a heretofore aseismic location, with the nearest previously detected earthquakes located at least 60 km away. According to International Seismological Center (ISC) catalogue, the event was followed by a sequence of 15 aftershocks with duration of 21 hours that extended east-southeastward as far as 126 km (Figure 2.1b). These events were recorded by a temporary seismic network deployed on the Fiji Islands (Figure 2.1a), which enables detailed study of the mainshock rupture properties as well as the locations and focal mechanisms of some larger aftershocks. This aftershock sequence provides important insight into the mode of seismic
nucleation in previously subducted materials, as well as the subduction history and the deep slab
geometry in this region.

2.3 Data and Methods

2.3.1 Body Waveform Inversion for the Mainshock Rupture Properties

We use vertical component broadband data from both local (< 10°) (Figure 2.1a) and Global
Seismic Network (GSN) teleseismic stations (30°-85°) (Figure 2.2a), after removing instrument
response to obtain the displacement records. The addition of local stations enabled us to
characterize the vertical rupture component better. Local records were low pass filtered at 0.5 Hz
to provide frequency content that is similar to teleseismic records attenuated along long paths.
We exclude stations near nodal planes to avoid waveform complexity. Waveform polarities were
modified to have positive first motions only, and the P-wave first arrival times were manually
picked as reference times for the following analysis.

To investigate the mainshock rupture details, we first conduct a qualitative directivity
analysis [Silver et al., 1995; Ammon et al., 2005] to determine the major rupture directions,
number of sub-events, and their origin times and locations. Then with this starting model, we
predict the arrival times of each sub-event at each station. By assuming Gaussian-shaped source
time functions centered at the predicted arrival times, we then determine the best fitting duration
and amplitude for each event at each station, and take the average values as representing the
source [Tsai et al., 2005; Zhan et al., 2014]. We then determine the best fitting sub-event times
and locations using a damped iterative Gauss-Newton algorithm that minimizes the misfit between the predicted and observed waveforms (Figure 2.2b).

2.3.2 Relative Relocation of Aftershocks and Background Seismicity

We combined P and S phases from the temporary regional network (Figure 2.1a) and teleseismic P, pP, and PKP phases from the International Seismological Center (ISC) to constrain the relative locations of the aftershocks with a hypocentroidal decomposition algorithm that minimizes the effect of velocity heterogeneities along the ray paths [Jordan and Sverdrup, 1981]. At least 10 more aftershocks in addition to those in the ISC catalogue can be identified on waveforms from the temporary network, however they are small and their locations cannot be precisely determined with the limited number of regional arrivals. 4 out of 15 ISC identified aftershocks were also omitted during the relocation process due to insufficient number of arrivals.

Background deep earthquakes (depth >300 km and north of 22°S) that occurred between 1964 and 2012 were relocated using regional P and S phases as well as teleseismic P, pP, and PKP arrival times from ISC. We discarded background events with average 95% confidence semiaxis lengths larger than 10 km to ensure that all events were accurately located.

2.3.3 Coulomb Stress Change Calculation

We calculate the static Coulomb stress changes around the mainshock rupture area in an elastic half-space assuming a finite-fault source model for the mainshock [Lin and Stein, 2004; Toda et al., 2005]. According to the body wave inversion result, the fault model consists of five sub-events (Table A2.1) with an approximate 30 km by 30 km overall rupture dimension along the NE striking steeply dipping nodal plane. For each sub-event, a uniform slip distribution of 2 m is used and the rupture dimension is estimated from their seismic moment. In the calculations,
we assumed a shear modulus of $11.6 \times 10^{11}$ dyn cm$^2$, appropriate for a depth of 590 km, and a Poisson’s ratio of 0.24 for olivine, the dominant mineral in mantle, and an apparent coefficient of friction of 0.7. The orientation of the receiving faults is set to be the same as that of the mainshock rupture fault.

### 2.4 Results and Discussions

#### 2.4.1 Rupture Characteristics and Aftershock Locations

Body wave inversion results show that the mainshock faulted with a bilateral rupture pattern in the northeast-southwest (NE-SW) direction (Figure 2.1b and 1c, Table A2.1). In the NE direction, the rupture propagated somewhat downward with an average apparent velocity of $\sim 4.6$ km/s, while in the SW direction, the rupture propagated roughly horizontal with an average apparent velocity of $\sim 1$ km/s. The overall mechanism of the mainshock consists of an east-northeast striking steeply dipping nodal plane, and a south striking shallower dipping nodal plane (Figure 2.1b). The NE-SW distribution and significant depth variation of the rupture sub-events suggests that the faulting occurred along the steeply dipping nodal plane. The overall rupture dimension is about 30 km by 30 km, giving an estimated stress drop of 13 MPa assuming a circular rupture. This stress drop is comparable to results for other deep earthquakes [Wiens, 2001; Tibi et al., 2003a; Frohlich, 2006a; Zhan et al., 2014; Houston, 2015].

The mainshock and the 11 well-located aftershocks are distributed along an east-southeast (ESE) striking trend (Figure 2.1b, Table A2.2), at a large angle to the NE-SW striking rupture direction. This distribution does not agree with either fault plane of the mainshock, which has
been noted previously for some deep earthquakes [Frohlich and Willemann, 1987] but not others [Wiens et al., 1994]. Between 1964 and 2012, more than 360 deep background earthquakes not directly associated with a subducted slab were identified (Figure 2.1a). These earthquakes include both earthquakes in the region immediately west of the Tonga slab and a concentration of deep earthquakes between the Vanuatu and Vitiaz Trench known as the Vitiaz cluster [Okal and Kirby, 1998]. Relocation results show the mainshock and most of its aftershocks extend a west-northwest (WNW) trending zone of past seismicity into a previously aseismic region at the northern boundary of the flat slab.

2.4.2 Relationship to Tonga and Vitiaz Slabs

This feature, denoted here as the northern seismicity band, has an intriguing relationship with the seismicity of the deepest Tonga slab (Figure 2.3). The northern band splits into two features at its eastern terminus; one section curves southward and deeper, forming a linear north-south feature denoted here as the eastern band that contains many of the deepest earthquakes (654 km – 687 km) in the Tonga region. A semicircle shaped seismicity gap between this zone and the main Tonga WBZ indicates that the eastern band is separated by a slab tear from the currently subducting Tonga slab. Another zone of seismicity, termed the hinge cluster in Figure 2.3b, extends from the northern band to the northernmost termination of the active Tonga slab. The hinge cluster is dominated by down-dip compressional stresses that are roughly orthogonal to the strike of the active slab to the east (Figure 2.4a), suggesting this cluster is being compressed by the subduction of the active slab. The northern band, which is composed by the aftershock sequence and adjacent background earthquakes, defines a sharp northern boundary for the Tonga outboard earthquakes. The northern band has an approximate thickness of 30 km and
is dipping to the east (Figure 2.4a), which may be an outcome of the compression and induced mantle flow by the rapidly subducting present day Tonga slab.

The newly defined northern band of earthquakes at the boundary of the Tonga outboard region is in line with the long axis of the Vitiaz cluster (Figure 2.1a and 3a), suggesting a connection between these two seismic zones. We further analyzed focal mechanisms for earthquakes along the Tonga to Vitiaz lineaments [Dziewonski et al., 1981; Ekström et al., 2012]. The Vitiaz cluster can be divided into two groups, a compact northwestern cluster and a more diffuse southeastern group (Figure 2.1a). Earthquakes within the southeastern part of the Vitiaz cluster and the northern Tonga outboard lineament show similar compression axis orientations, parallel to the lineament strike (Figure 2.1a and 4). This suggests that the subducted slab defined by the linear boundaries may be continuous, with stresses resulting from the compression of the main slab transferred along the detached Vitiaz slab, even though no earthquakes have occurred along the central part during the past half-century.

Further insight is provided by 3-D seismic velocity models of the region [Amaru, 2007; Simmons et al., 2012]. Both the southeastern part of the Vitiaz cluster and the northern Tonga band of earthquakes are located near the edge of a continuous fast velocity anomaly striking northwest at around 550 km depth (Figure 2.3a). The coincidence of the Tonga to Vitiaz lineament with high velocity material in the transition zone suggests they both mark the location of a recently subducted slab. Apparently isolated deep earthquakes worldwide are located in high velocity regions representing recently subducted, but largely aseismic, slabs [Bezada et al., 2013]. Additional evidence to link the Vitiaz cluster to the northern band of Tonga earthquakes and to the Tonga main slab comes from the slope of the earthquake magnitude frequency relationship (b-value). The Tonga subduction zone is distinctive for showing the highest b-value of deep
subduction zones worldwide [Wiens and Gilbert, 1996], and both the Vitiaz cluster and the northern band of outboard earthquakes show statistically identical b-values to that of the Tonga main slab (Figure A2.1).

The location and orientation of the Tonga to Vitiaz lineaments are similar to that of the reconstructed Vitiaz subduction zone at 10 Ma [Schellart et al., 2006] (Figure 2.3a). Southwest-directed Pacific plate subduction at the Vitiaz Trench ceased about 8-10 Ma due to the impingement of the Ontong Java Plateau, after which northeast directed subduction began along the present-day Vanuatu Trench [Hamburger and Isacks, 1987; Schellart et al., 2006]. We suggest that the earthquakes extending along the northern band to the Vitiaz cluster represent the last portion of the slab subducted along the fossil Vitiaz trench. The detached Vitiaz slab followed a dipping trajectory to its present day depth under the effect of gravity. The curved zone of earthquakes between the northern band and the eastern band (Figure 2.3b) is the junction of the ancient Vitiaz subduction slab and the Tonga subduction slab. The eastern band and the hinge cluster (Figure 2.3b) represent fragments of the ancient Tonga slab which became detached from the present-day Tonga slab due to the rapid eastward motion of the central and northern Tonga trench caused by slab rollback.

2.4.3 Dynamic Triggering within a Fossil Slab

Thus, we propose the 9 November 2009 earthquake occurred within the fossil Vitiaz slab, with many of the aftershock locations controlled by the availability of cold or seismogenic slab material in the surrounding region rather than by the rupture plane. Five aftershocks within 25 km of the mainshock hypocenter are localized near the fault plane defined by the mainshock sub-events (Figure 2.1b). We make a Coulomb stress change calculation to investigate which aftershocks likely resulted from static stress changes during the mainshock. The results show that
the five aftershocks near the fault plane are located within or very close to an area where the Coulomb stress increased by more than 0.2 MPa (Figure 2.5). These results suggest these adjacent aftershocks may have been triggered by static stress change caused by the mainshock rupture.

Six other aftershocks are more distant from the mainshock, and are located in regions with near-zero or negative Coulomb stress changes (Figure 2.5), suggesting they were not induced by the static stress increase from the mainshock. Instead, they were triggered by the dynamic strains associated with the propagating seismic body waves [Hill et al., 1993; Tibi et al., 2003b]. Synthetic seismograms calculated for the mainshock at the location of the aftershocks using orthonormal propagator algorithm [Wang, 1999] show that the strain experienced during the passage of the shear waves is on the order of $10^{-6}$, comparable to the dynamic strains known to trigger shallow earthquakes [van der Elst and Brodsky, 2010; Johnson et al., 2015]. The locations of these dynamically-triggered earthquakes were likely in a stressed state prior to the arrival of strain-inducing seismic waves, but perhaps the normal seismic nucleation process is inhibited [Tibi et al., 2003b]. Similar phenomenon are also observed in the warmer slab regions surrounding the slab core [McGuire et al., 1997] and in the normally aseismic extensions of slabs [Ye et al., 2016].

2.4.4 Implications for the Mechanism of Deep Earthquakes

There is still considerable debate about the physical mechanisms of deep earthquakes, with the most widely accepted ideas being some variation of transformational faulting [Green and Burnley, 1989; Kirby et al., 1991] or runaway ductile shear instabilities [Ogawa, 1987; Karato et al., 2001]. The dynamically triggered aftershocks occurred thirty minutes to twenty-one hours after the passage of P and S waves from the mainshock, indicating some short-term nonlinear
delay mechanisms have been involved [Freed, 2005]. These delay times are similar to those observed for other dynamically triggered deep earthquakes [Tibi et al., 2003b], and may be characteristic of deep earthquakes. We estimate the thermal parameter of the detached Vitiaz slab, the product of the lithospheric age and downward velocity [Kirby et al., 1991], is about 7000 km, larger than that of the seismically active deep South American slab. The estimated minimum temperature of the slab interior is about 690°C at 600 km [Frohlich, 2006b]. Thus the temperature in the Vitiaz slab should be near but not above the limiting temperature observed for deep earthquakes in other slabs, and within the region of olivine metastability but close to the kinetic phase boundary [Chien-Min and Burns, 1976; Kirby et al., 1991]. Both the transformational faulting and ductile instability models could accommodate the presence of regions near the criticality for shear failure that are unable to nucleate failure in slabs that are marginally able to support seismicity. In this case, the passage of seismic waves can begin a process under which failure occurs within a period of minutes to hours. Further observational and laboratory studies of these processes may play a key role in clarifying the physical mechanisms of deep earthquakes.

2.5 Conclusions

The 2009 Mw 7.3 Fiji deep earthquake faulted with a bilateral rupture pattern in NE-SW direction, with very fast rupture velocity in NE direction (~4.6 km/s) and slow rupture velocity in SW direction (~1 km/s). We infer that the mainshock statically triggered the adjacent aftershocks (< 25 km) and dynamically triggered the others. The mainshock and its aftershock sequence were located within a normally aseismic region, and their distribution does not agree with either fault plane of the mainshock. The background seismicity, the transition zone P wave velocity as well
as the tectonic reconstruction result suggest the mainshock and its aftershock sequence occurred in the final portion of the slab subducted at the now inactive Vitiaz trench. Material within the normally aseismic fossil slab appears particularly susceptible to dynamic triggering by nearby deep earthquakes, with delay times of minutes to hours. The presence of fossil slabs near critical shear failure conditions that are unable to nucleate slip without a triggering stress perturbation is compatible with either the transformational faulting or runaway shear instability models for producing deep earthquakes.

References


Figure 2.1. Maps of the study region and regional seismicity.

(a) Map showing hypocenters of relocated deep earthquakes (>300 km) beneath the Lau and North Fiji Basins between 1964-2012, with colors denoting focal depth. Contours mark depths to the Waditi-Benioff Zone (WBZ) from Slab 1.0 model [Hayes et al., 2012]. The filled star is the 9 November 2009 Mw7.3 mainshock; Filled triangles are earthquakes within the Tonga WBZ; Filled circles are outboard earthquakes discussed in this study. Black dashed box marks the area shown in Figure 2.1b. Inset shows the mainshock location and the distribution of a temporary seismic network (black filled triangles) as well as a Global Seismic Network station (gray filled triangle) deployed on the Fiji islands. (b) Enlarged map view showing the mainshock (yellow star), the mainshock sub-events inferred from waveform analysis (red open diamonds), the aftershock sequence (gray filled circles), background seismicity from 1964-2012 (black open circles), as well as an earthquake that occurred well after the aftershock sequence (black open hexagon). Focal mechanisms are plotted as lower focal hemisphere projections for the mainshock and larger aftershocks. (c) Cross section showing the locations of the mainshock sub-events along AA’.
Figure 2.2. Station distribution and waveform fitting.

(a) Map showing location of the mainshock with a lower focal hemisphere projection of its focal mechanism and teleseismic stations (red triangles) used for waveform analysis; (b) Waveform fits for the sub-event model. The data is plotted in black and the synthetics are plotted in red. Records with directivity parameters greater than 0 are from local stations; (c) Source time functions for each sub-event (black) and the overall rupture (gray).
Figure 2.3. Maps of velocity anomaly and earthquake distribution.

(a) Map showing the relationship between the deepest outboard earthquakes (>570 km) and regional P velocity anomalies at 550 km depth from LLNL_G3Dv3 model [Simmons et al., 2012]. Black dashed box indicates the area shown in Figure 2.3b. White dashed lines indicate locations of the cross sections in Figure 2.4. Inset is a tectonic reconstruction showing the location of the Vitiaz trench at 10 Ma [Schellart et al., 2006], red dashed box marks the area shown in the velocity anomaly map. (b) Enlarged map view showing hypocenters of the deepest earthquakes (>570 km) within the Tonga WBZ (filled triangles) and within the detached Vitiaz slab (filled circles). Thick gray dashed line sketches the fossil Vitiaz and Tonga slab dating from the time of active Vitiaz subduction.
Figure 2.4. Cross-sections B-B’ and C-C’ showing earthquake distribution and focal mechanisms.

(a) Cross section showing locations and focal mechanisms of earthquakes along the Tonga northern seismicity band. Symbols are defined as in Figure 2.1b, and open squares represent earthquakes within the hinge cluster. Focal mechanisms from the GCMT catalog are shown as horizontal hemisphere projections; White/gray focal mechanisms are solutions for three larger aftershocks obtained in this study. Inset is a lower hemisphere stereographic plot showing the compression axis directions for events within the northern band (black open circles) and within the hinge cluster (open squares); The solid line represents the strike of the cross section (290°);

(b) Cross section showing locations and focal mechanisms of earthquakes distributed along the northern boundary of the Vitiaz cluster. Stereographic plot shows the compression axis directions. The solid line represents the strike of this profile (290°). Dashed lines are inferred boundaries of the slab from the earthquake distribution.
Figure 2.5. Map showing the location of the aftershock sequence, and the calculated static Coulomb stress changes.

(a) at 580 km depth and (b) at 610 km depth. Open circles are aftershocks shallower than 595 km, and black dots are aftershocks deeper than 595 km.
Figure A2.1. B-value plot.

Frequency–magnitude distribution for deep events (> 300 km) between 1964 and 2012 within the active Tonga slab (black), the Vitiaz cluster (cyan), as well as the Tonga outboard earthquakes (magenta). Magnitude is body wave magnitude from ISC catalogue, and earthquakes are from relocation results in this study. Dots are data points, circles are data points used to estimate b-value, dashed lines are linear regression results.
## Table A2.1. Sub-event model for the main shock

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## Table A2.2. Aftershocks of the November 9, 2009, Fiji deep earthquake

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*the event is eliminated during the relocation process and location is obtained from ISC catalogue

1 average of 95% confidence semiaxis
Chapter 3: Water Input into the Mariana Subduction Zone Estimated from Broadband Ocean Bottom Seismic Data

3.1 Abstract

The water cycle at subduction zones remains poorly understood, despite being the only mechanism for water to be transported deep into the Earth. Previous estimates show large variation in the amount of water subducted deeper than 100 km [Rüpke et al., 2004; Hacker, 2008; van Keken et al., 2011]. A major uncertainty in these calculations is the initial water content of the subducting uppermost mantle. Previous active source seismic studies suggest the subducting slab may be pervasively hydrated in the plate-bending region near the trench [Ranero et al., 2003; Van Avendonk et al., 2011; Fujie et al., 2013; Shillington et al., 2015; Grevemeyer et al., 2018]. However, these studies do not constrain the depth extent of hydration and most investigate young incoming plates, thus leaving subduction zone water budgets for old subducting plates uncertain. In this study, we present new seismic images of the crust and uppermost mantle around the central Mariana Trench derived from Rayleigh wave analysis of ocean bottom seismic data, showing that slow velocities resulting from mantle hydration extend 24±5 km beneath the Moho. Combined with estimates of subducting crustal water, these results indicate that at least 4.3±0.8 times more water subducts than previously calculated for this region [van Keken et al., 2011]. If other old, cold subducting slabs contain correspondingly thick layers of hydrous mantle, as suggested by the similarity of incoming plate faulting, estimates of the global water flux into the sub-arc mantle must be increased by about a factor of three over
previous estimates [van Keken et al., 2011]. Since a long-term net influx of water to the deep interior is inconsistent with the geological record [Parai and Mukhopadhyay, 2012], it is likely that estimates of water expelled at volcanic arcs and back arc basins also must be revised upward [Grove et al., 2012].

### 3.2 Introduction

The Mariana subduction zone has long been cited as a water-rich system due to the prevalence of forearc serpentinite mud volcanoes [Fryer, 2012], a serpentinized mantle wedge [Barklage et al., 2015], and hydrous arc and backarc lavas [Shaw et al., 2008; Kelley et al., 2010], yet the initial amount of water within the subducting mantle is unknown. The subducting Pacific Plate in this region is among the oldest sections of oceanic lithosphere worldwide (~150 Ma) [Müller et al., 2008], and the incoming plate displays large plate-bending normal-fault scarps and earthquakes [Oakley et al., 2008; Emry et al., 2014], thus making it an excellent place to investigate the depth extent of faulting-induced hydration in an old oceanic plate. Potential hydration of old, cold oceanic plates are particularly important for the water cycle since the plates’ thermal structure permits temperature-sensitive hydrous minerals to occur throughout a thicker region [Rüpke et al., 2004].

We used the data collected by ocean bottom seismographs (OBSs) and island seismographs deployed around the central Mariana trench (Figure 3.1). The deployment covers sufficient large region for investigating structure variations in the subducting plate both prior to and following subduction. We derived local Rayleigh wave group and phase velocity dispersion curves with
two surface wave methods. A Bayesian Monte-Carlo algorithm was then used to determine a three-dimensional shear-wave velocity image of the crust and uppermost mantle. Compared to previous active source seismic studies of other subduction zones [Van Avendonk et al., 2011; Fujie et al., 2013; Shillington et al., 2015], our study resolves the 3-D structure to greater depth and avoids biases caused by possible azimuthal anisotropy. In addition, it images shear velocities which show stronger sensitivity to hydration than compressive velocities [Christensen, 2004].

3.3 Results and Discussions

The derived azimuthally averaged velocity model of vertically polarized S waves (SV waves) shows systematic changes in the incoming plate and subducting slab along the trench normal direction (Figure 3.2). At distances far from the trench axis (>100 km seaward), the subducting plate shows a typical structure of old oceanic lithosphere [Nishimura and Forsyth, 1989] at depth greater than 20 km, with velocities faster than 4.5 km/s. A somewhat lower velocity (~4.2 km/s) layer of ~10 km thickness is observed immediately beneath the Moho, consistent with active source seismic results [Feng, 2016]. A thicker region of low velocities begins about 80 km seaward of the trench axis and deepens towards the trench, with velocities as low as 3.8 km/s. At the trench axis, the bottom of this low velocity layer reaches 30±5 km beneath the seafloor, which is 24±5 km into the upper mantle. This low velocity layer persists after the Pacific plate is subducted as a 30±5 km thick low velocity layer atop the fast slab mantle between 100-160 km west of the trench axis. Synthetic data tests show that a velocity anomaly with this large magnitude and extent cannot result from the effect of limited resolution of a ~6 km thick layer of low velocity oceanic crust (Figure A3.5).
Additionally, we also solved for phase velocity azimuthal anisotropy. These results show trench parallel fast axes in the incoming Pacific plate between periods of 12-18 s, with a maximum magnitude as large as 9% reached between 14-16 s. In contrast, at periods longer than 20 s, the fast directions rotate to be oblique to the trench strike, close to the paleo-spreading direction [Oakley et al., 2008] (Figure A.3.2).

The region of incoming plate mantle velocity reduction and large azimuthal anisotropy coincides with the plate-bending region, characterized by significant normal faulting seismicity [Emry et al., 2014] and large extensional seafloor fault scarps [Oakley et al., 2008]. There is a clear spatial association between plate-bending faulting and velocity reduction. Velocities within the incoming plate begin to decrease 80 km from the trench, at the same location where faulting begins on the seafloor [Oakley et al., 2008] and where seismicity also begins. In this region, the fault scarps and earthquake fault planes are approximately north-south, parallel to the trench axis. This is consistent with the phase velocity azimuthal anisotropy observations at 12-18 s, which primarily sample depths down to ~25 km below the seafloor (Figure A3.4i). These results show trench-parallel fast directions as would be expected for pervasive trench-parallel water-filled faults or zones of alteration. The flexure model that best fits the Pacific plate bathymetry seaward of the trench axis predicts a neutral plane at ~30 km [Emry et al., 2014], suggesting that brittle normal faulting can extend nearly 30 km into the plate. This prediction agrees well with the maximum depth extent (30±5 km) of the low velocity zone (LVZ) we observed on the incoming plate near the trench in our study (Figure 3.2), suggesting this LVZ can be directly related to brittle normal faults.

Many previous studies at other subduction zones have attributed upper mantle low velocity zones associated with plate bending to the hydration of mantle peridotite to form low velocity
serpentine minerals. Extensional deformation within the shallow part of the incoming plate produces a pressure gradient that may enable water to penetrate deep into the slab along normal faults [Faccenda et al., 2012]. The serpentinization rate is geologically fast if water delivery to the serpentinization front is efficient [Reynard, 2013; Nakatani and Nakamura, 2016]. Alternatively, other studies attribute the mantle velocity reductions to water-filled porosity and cracks [Korenaga, 2017]. However, the potential velocity effect of water-filled cracks is usually difficult to estimate directly as it depends critically on the crack aspect ratio and density, which are largely unknown. In this study we use the increase in velocity as the plate subducts and porosity is reduced to distinguish effects of water-filled cracks and serpentinization.

The Mariana Trench seismic images show that the velocity of the LVZ increases as the top of the slab subducts past ~30 km depth (Figure 3.2). Since porosity reductions due to increased pressure [David et al., 1994] will reduce or eliminate the velocity effect of water-filled cracks at depth, whereas hydrous minerals will remain stable at the cold slab temperatures, this provides a means to separate the complementary velocity effects of porosity and altered minerals and to estimate the concentration of hydrous minerals. Compared to the LVZ within the slab prior to subduction, the LVZ within the subducted slab mantle at depths of ~40 km preserves the original thickness (30±5 km) but shows a smaller velocity reduction (~4.1 km/s) (Figure 3.2). We use this smaller velocity reduction to estimate the degree of mantle serpentinization in the downgoing slab. Therefore, we base our estimates of the water content due to serpentinization of the subducting slab on the initial thickness of the low velocity layer at the trench, and the shear wave velocities of around 4.1 km/s observed at depths of 30-50 km in the subducting plate mantle, after most of the pore water will be expelled. The additional velocity reduction (~0.3 km/s) within the LVZ in the slab prior to subduction can be attributed to water-filled cracks and
porosity.

Calibrating the seismic velocity change to the degree of serpentinization requires knowledge of the seismic velocity of serpentine. We select lizardite, the form of serpentine expected to predominate at lower temperatures [Nakatani and Nakamura, 2016], to interpret the observed velocity reduction. The nominal temperature predicted by plate cooling models [Stein and Stein, 1992] at ~30 km beneath the seafloor is ~470°C, possibly higher than the temperature of lizardite breakdown (~320°C from Schwartz et al. [2013], although Nakatani and Nakamura [2016] finds lizardite at temperatures as high as 580°C). Water circulation in cracks may lower the slab mantle temperature in the plate-bending region into the lizardite stability field. In addition, selection of lizardite will provide a lower bound estimate of water input. Using the experimental relationship between shear velocity and serpentine volume fraction for lizardite [Ji et al., 2013], the observed 4.1 km/s shear velocity within the subducted slab corresponds to ~19 vol% serpentinization (~2 wt% water). It is possible that the anisotropic effects of serpentine distributed along bending faults could cause additional velocity reduction, thus leading to an overestimate of serpentinization percentage [Miller and Lizarralde, 2016]. However, according to our calculation with a realistic dipping fault geometry and the frequencies used in this study, the anisotropic effects are not significant for serpentinization percentage estimation (Figure A3.7).

Thus we interpret the seismic images as strong evidence for a 24±5 km thick partially serpentinized (2 wt% water) slab mantle layer. Applying a convergence rate of 50 mm/yr, the amount of water input into the Mariana subduction zone through mantle serpentinization would be ~79±17 Tg/Myr/m, and the total water flux is ~94±17 Tg/Myr/m if water in the sediment and crust is also included from previous estimates [van Keken et al., 2011]. This new estimate of
Mariana trench water flux is 4.3±0.8 times larger than the estimation from van Keken et al. [2011], which assumed only a 2 km thick partially serpenitized slab mantle (2 wt% water). All uncertainties are estimated based on the uncertainty of the serpentinized slab mantle thickness.

Our interpretation of serpentinization extending to depths of ~24 km below the Moho in the incoming plate at the Mariana Trench has significant implications for water flux into subduction zones globally. This depth is somewhat greater than the maximum observed depth of large normal faulting earthquakes and close to the estimated depth of the neutral plane [Emry et al., 2014]. The maximum depth of serpentinization near trenches has not been well determined for other older incoming plates, as the depth extent may be too great to be well constrained with active source seismic studies [Ranero et al., 2003; Van Avendonk et al., 2011; Fujie et al., 2013; Shillington et al., 2015; Grevemeyer et al., 2018] and surface wave investigations have not been performed elsewhere. The bending and faulting features of the incoming Pacific plate in Mariana is similar to what is observed at other old subduction plates, and the maximum depth of normal faulting and depth of the neutral plane is generally about the same depth [Emry and Wiens, 2015], so it is possible that serpentinization extends to similar depths of 20-25 km below the Moho at other sites where old lithosphere subducts. Modifying previous global subduction zone water flux calculations for this increased thickness of mantle hydrous alteration increases the estimated flux to $3.0 \times 10^9$ Tg/Myr, which is about three times greater than previous calculations [Rüpke et al., 2004; Hacker, 2008; van Keken et al., 2011].

This new larger estimate of the input water flux from subduction zones is much greater than current estimates of water output from the mantle. Since a large long-term net influx of water to the deep interior is inconsistent with the stability of sea level in the geological record [Rüpke et al., 2004; Parai and Mukhopadhyay, 2012], one possible implication of this result is that the
thick layer of serpentinized mantle we find in Mariana is not characteristic of other old, cold subducting slabs, and the Mariana slab carries much more water than other subduction zones. However, there is little indication that the Mariana incoming plate-bending region is significantly different in terms of morphology and intensity of faulting compared to the corresponding regions of other old subduction zones. Thus the most likely interpretation is that previous estimates of water output from the mantle are also underestimated. Estimates of mantle water output from mid-ocean ridges and ocean islands may be relatively well constrained, but estimates for volcanic arcs and backarcs rely on the melt flux and the water content, which are poorly constrained [Parai and Mukhopadhyay, 2012; Grove et al., 2012].

3.4 Supplementary Materials

3.4.1 Data Processing and Group/Phase Velocity Tomography.

The data used in the group and phase velocity tomography were mainly collected by 19 ocean bottom seismographs (OBSs) and 7 temporary island-based seismic stations deployed from January 2012 to February 2013 (Figure 3.1). The OBS distribution covers both the trench-outer rise region and the Mariana forearc region. We also used data from 3 island stations from the USGS Northern Mariana Islands Seismograph Network active over the same time period as addition to our deployment.

We carried out ambient noise tomography (ANT) following the procedures described by Bensen et al. [2007] and Lin et al. [2008]. The daily vertical-component seismograms were corrected for instrument responses and clock errors, and down-sampled to 2 samples per second
We applied running-average time-domain normalization and spectral whitening to minimize the effects of large earthquakes. Seismograms from all station pairs were then cross-correlated and stacked over the entire time period of the deployment. Frequency-time analysis [Bensen et al., 2007; Levshin et al., 2001] was applied to the symmetric components of the stacked cross correlations to measure Rayleigh wave group and phase velocities between periods of 8 and 25 second. For each frequency, only station pairs with distance larger than twice the wavelength were kept. All dispersion curves were screened to exclude those with inconsistent measurements at adjacent periods. For each period, a ray theory based tomography method [Barmin et al., 2001] was applied to dispersion measurements with signal-to-noise (SNR) ratios greater than 5 to produce Rayleigh group and phase velocity maps on a grid of nodes spaced at 0.2°. The tomographic inversion returns both isotropic and azimuthal anisotropic components of the Rayleigh wave group and phase velocity (Figure A3.3).

We utilized the Helmholtz tomography (HT) method described by Jin and Gaherty [2015] to teleseismic Rayleigh waveforms to determine phase velocities at longer periods. On the basis of the International Seismological Centre (ISC) catalogue, we selected seismograms from 380 earthquakes with surface-wave magnitudes ($M_s$) larger than 4.5 and epicentral distances between 25° and 150°, which occurred during the time when the stations were operating (Figure A3.1). The raw seismogram of each event was cut from the origin time of the earthquake to 12000 s after. Prior to any further analysis, the vertical-component seismograms were down-sampled to 1 sps and instrument responses were corrected. Noise in seismograms at long periods (>50 s) due to ocean swell and associated water pressure variations, as well as tilt caused by local currents, were removed by correcting the vertical channel using horizontal and pressure channels [Webb and Crawford, 1999; Crawford and Webb, 2000; Bell et al., 2015].
This implementation of the Helmholtz tomography method recovers frequency-dependent phase and amplitude information via the narrow-band filtering of the broadband cross-correlations between the vertical component seismogram from a given station and time-windowed seismograms from all other nearby stations. The phase delays and amplitude information were determined by fitting the narrow-band filtered cross correlations with a Gaussian wavelet [Jin and Gaherty, 2015]. To eliminate the influence of poor-quality records, we estimated the coherence between waveforms from nearby stations for a series of periods from 21 to 53 second, and only included those measurements with coherence larger than 0.5. For each earthquake and each period, we inverted the phase delays for spatial variations in dynamic phase velocity via the Eikonal equation [Lin et al., 2009]. We then further corrected the propagation effect via Helmholtz tomography [Lin et al., 2011], producing maps of structure phase velocity with spacing of 0.2°. This tomographic method returns both the azimuthal isotropic phase velocity and the azimuthal anisotropic component at each node simultaneously.

3.4.2 Bayesian Monte-Carlo Inversion

We combined the ANT and HT results to provide more complete measurements of phase velocity for the SV-velocity inversion (Figure A3.4). The two sets of dispersion curves were combined in the geographical region well resolved by both methods. Phase velocities were interpolated onto a uniform grid of nodes with a spacing of 0.2° before being combined at each node. For phase velocities from ANT (8-25 s), the uncertainties were normalized at each period so that the uncertainty of the best-resolved node equals to 0.075 km/s. The uncertainties of the group velocity (8-21 s) were normalized that the best-resolved node has uncertainty of 0.188 km/s, 2.5 times the value for phase velocities [Shen et al., 2016]. For phase velocities from HT
(21-53 s), the uncertainties were normalized at each period that the best-resolved node has uncertainty equal to the standard deviation of velocity differences between HT results and results from a two-plane-wave tomography method [Yang and Forsyth, 2006] (Figure A3.6). We utilized a linear weighting average method to combine phase velocity measurements and uncertainty estimates for overlapping periods (22-25 s). A running average was then applied to make the resulting dispersion curve smoother. Group velocity results from ANT (8-21 s) were also included for the SV-velocity inversion to better constrain water thickness and to improve resolution for shallower structure.

We use a Bayesian Monte-Carlo algorithm [Shen et al., 2013] to invert the azimuthal averaged SV-wave velocity at each node. This approach allows us to apply prior constraints on crustal thickness and other parameters in a systematic way, avoid any potential bias of the starting model [Wei et al., 2015], and to derive formal estimates of velocity uncertainty.

The Bayesian Monte-Carlo method constructs an a priori distribution of SV velocity models at each node, defined by perturbations relative to the starting model and model constraints. Each model consists of four layers on top of a half-space: (1) water with starting thickness from bathymetry [Lindquist et al., 2004] that has been smoothed with a Gaussian filter (at 125 km length) and an allowed perturbation of ±1.5 km, (2) sediments, (3) crust, and (4) upper mantle from the Moho to 180 km depth. The sedimentary layer is described by two parameters: a layer thickness of 0.5 km with an allowed perturbation of ± 0.5 km and a constant $V_{SV}$ of 2.0 km/s with a perturbation of ± 1.0 km/s. The crust is assumed to have linearly increasing velocity with depth, and is described by three parameters: a layer thickness and $V_{SV}$ at the top and bottom of the layer. For the incoming plate east of the trench, the crustal thickness is allowed to vary ±1.5 km around the starting value 6.5 km. The forearc crustal thickness perturbs within 3 km with
starting values from a previous seismic refraction survey at the southern edge of the study region ranging from 19 km to 6.5 km [Takahashi et al., 2008]. The top and bottom crustal $V_{SV}$ are set at 3.0 and 3.2 km/s, respectively, with a perturbation of $\pm$ 1.0 km/s. The upper mantle $V_{SV}$ is parameterized by a B-spline, which is defined by 7 nodes with a perturbation of $\pm$ 30% for first 5 nodes and 20% for the last 2 nodes. We imposed the constraint that the jumps in $V_{SV}$ from the sediment to the crust and from the crust to the mantle are positive.

We also applied a physical dispersion correction with a reference period of 1 s [Kanamori and Anderson, 1977] using a 1-D Q model simplified from a seismic attenuation study in the same region [Pozgay et al., 2009]. Compared to the PREM, our 1-D Q model for the forearc has a high-attenuation layer in the uppermost mantle: $Q_S=60$ from Moho to 100 km depth. For the incoming plate region, the uppermost mantle is set to have a typical lithospheric attenuation: $Q_S=300$ from Moho to 100 km depth.

For each grid node, the best fitting model is identified and models are accepted if their $\chi^2$ misfit is less than 50% higher than that of the best fitting model [Shen et al., 2013, 2016]. We also exclude models with mantle velocity higher than 4.9 km/s. The posterior distribution thus provides statistical information of all possible SV-velocity models that satisfied the Rayleigh wave dispersion curves within tolerances depending on data uncertainties. An average model is then calculated from all accepted models and used for plotting and interpretations [Shen et al., 2013; Wei et al., 2016]. Examples of the SV-velocity inversion at four representative nodes are shown in Figure A3.4. The results of the Bayesian Monte-Carlo inversion fit the measured group and phase velocity dispersion curves well.

3.4.3 Robustness of the Thick Low Velocity Layer
The application of Bayesian Monte-Carlo algorithm [Shen et al., 2013] helps to avoid the potential bias of the starting models and provides better prior constraints on crustal thickness and other parameters. However, it uses a B-spline method to parameterize the upper mantle $V_{SV}$, thus may smooth a thin low velocity layer over a wider depth range. Here we run simulations to test whether the observed thick low velocity region above and below the subducting slab surface can be caused by the smoothing of the 6 km thick lower velocity crust due to the inversion parameterization and lack of precise depth resolution.

For the target node, we set up a 1-D shear velocity model without a low velocity serpentinized slab mantle layer based on our prior knowledge of the Mariana subduction zone geometry (Figure A3.5a), and calculate the synthetic phase/group dispersion curves [Herrmann, 2013]. We then apply the Bayesian Monte-Carlo inversion with the same parameterizations as in our study to the synthetic phase/group data and get a 1-D reconstructed shear velocity structure. Examples for two nodes are shown in Figure A3.5b and c, suggesting the thick low velocity layer observed in this study cannot be purely caused by the smearing of the subducting oceanic crust.

### 3.4.4 Serpentine and Water-filled Cracks

Previous studies at various convergent margins generally attribute the observed upper mantle slow velocity anomalies to the presence of serpentine without distinguishing the serpentine minerals involved [DeShon and Schwartz, 2004; Contreras Reyes et al., 2007, 2011; Shillington et al., 2015]. Although the three main serpentine minerals, lizardite, chrysotile and antigorite, have the same water content (~13 wt%), their physical properties including seismic velocity [Christensen, 2004; Ji et al., 2013] and stability field [Evans, 2004; Guillot et al., 2015] are quite different due to the different crystal structure. Lizardite and chrysotile are the more
abundant serpentine minerals in hydrated mantle rocks formed at low temperatures and are stable up to ~320°C at 1 GPa. When the temperature reaches between 320°C to 390°C, lizardite will be progressively replaced by antigorite at the grain boundaries and in the core of the lizardite meshes [Schwartz et al., 2013]. Antigorite is the main stable serpentine mineral at higher temperature (up to ~620°C at 1GPa) [Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997; Perrillat et al., 2005; Schwartz et al., 2013; Guillot et al., 2015]. Lizardite and chrysotile have much lower shear velocities (~2.3 km/s) compared to antigorite (~3.7 km/s) at 600 MPa [Ji et al., 2013]. For the same amount of shear velocity reduction, the serpentinization percentage, and thus the water content, estimated with antigorite component will be ~2.5 times of that with lizardite and chrysotile component assumption [Ji et al., 2013]. This feature makes it imperative to decide which serpentine minerals are present before making any further interpretation of observed velocity reductions. In addition to causing large seismic velocity reduction, serpentinization along the exposed normal faults faces will also generate significant azimuthal anisotropy (Figure A3.2) [Miller and Lizarralde, 2016].

Korenaga [2017] argues that the same amount of velocity reduction can also be caused by water-filled porosity, without involving substantial bulk hydration. This argument resembles a non-fractured isotropic media, which is incompatible with the field observations in the Mariana subduction zone where numerous normal faults and normal fault earthquakes have been detected [Oakley et al., 2008; Emry et al., 2014]. Instead, a porous media with aligned cracks should be the more appropriate assumption [Hudson et al., 1996; Gurevich, 2003]. On the other hand, this argument can only be applied to the slab prior to subduction. When the slab starts to subduct, the confining pressure increases with increasing depth, causing closures of cracks and expulsion of
free water within these cracks and/or porosity [David et al., 1994]. Thus this hypothesis is not applicable to the velocity reduction observed within the slab at greater depth after subduction.

### 3.4.5 Estimation of Global Subduction Zone Water Flux

We recalculated the global subduction zone water flux based on a previous estimation from van Keken et al. [2011], by re-evaluating the water content in the slab mantle. As serpentine minerals are stable up to 620°C, young and warm subducting plates have less potential to be serpentinized to significant depth. According to thermal models for oceanic plates [Stein and Stein, 1992], only plates older than about 40 Ma have 600°C isotherm deeper than 30 km beneath the seafloor, thus we set 40 Ma as a threshold for subducting plate age to be affected by deeper serpentinization. For subduction zones with subducting plate younger than 40 Ma, we simply take the water flux estimations from van Keken et al. [2011]. For subduction zones with incoming plate older than 40 Ma, we assume the slab mantle is partially serpentinized (2 wt% water) to 20 km below the Moho and keep the water volume in the sediment and crust unchanged as van Keken et al. [2011]. This rough estimation suggests the global subduction zone water flux should increase to $3.0 \times 10^9$ Tg/Myr.

### 3.4.6 Anisotropy Effect of Serpentine along Bending Faults

The anisotropic effects of serpentine distributed along bending faults could lead to an overestimate of serpentinization percentage [Miller and Lizarralde, 2016]. We calculated the long wavelength azimuthal anisotropy produced by evenly distributed serpentine layers [Backus, 1962]. According to our estimate of serpentinization percentage (~19 vol%), we assumed pure serpentine layers (450 m thick) were evenly distributed within isotropic peridotite with 2 km
spacing. We show results for two layering geometries, vertical layering and 45° dipping layering (Figure A3.7). Serpentine layers were set to have following properties: \( V_p = 5.1 \text{ km/s}, \ V_S = 2.32 \text{ km/s}, \ \rho = 2.52 \text{ g/cm}^3 \). Peridotite layers were set to have following properties: \( V_p = 8.1 \text{ km/s}, \ V_S = 4.51 \text{ km/s}, \ \rho = 3.32 \text{ g/cm}^3 \). The azimuthally-averaged quasi-SV velocity for the 45° dipping layering case is 4.08 km/s, very close to the SV velocity (4.1 km/s) we use to directly estimate serpentinization percentage. This suggests the anisotropy effect of serpentine distributed along bending faults may be less important when estimating serpentinization percentage [Miller and Lizarralde, 2016], especially when the bending faults are dipping.

References


Figure 3.1. Map of seismic stations used in this study.

Red circles are ocean bottom seismographs (OBSs) deployed from January 2012 to February 2013. White squares represent the temporary island-based stations. Red squares are stations from USGS Northern Mariana Islands Seismograph Network used in the study. Open triangles are locations of three large serpentine seamounts within the study area. White dashed line is the trench axis. White solid lines show the cross-sections in Figure 3.2. APM arrow shows the absolute plate motion direction.
Figure 3.2. Cross sections A-A’, B-B’ and C-C’ showing the azimuthally averaged SV-wave velocity.

White dashed lines are the forearc moho location [Takahashi et al., 2008]. Thick white lines are projected 6-km thick slab crust. Thin white lines are contours of 3.6 and 3.8 km/s, and thin black lines are contours of 4.1 and 4.2 km/s. Black circles are relocated earthquakes in the subducting plate around each profile [Emry et al., 2011].
Figure A3.1. Earthquakes used in this study.

Blue dots represent ISC earthquake locations. Red star show the location of the Mariana trench.
Figure A3.2. Azimuthal anisotropy results at various periods.

At 12 s (a), 14 s (b), 16 s (c), and 18 s (d), only results from the ANT method are plotted (red bars); at 21 s (e), results from both the ANT method and the HT method (yellow bars) are plotted; at 27 s (f), only results from the HT method are plotted. Trench axis and serpentine seamounts are labeled as in Figure 3.1.
Figure A3.3. Maps of azimuthally averaged group and phase velocity.

(a) and (b) show group velocity at periods of 10 s and 21 s inverted by ANT. (c) and (d) show phase velocity at periods of 10 s and 21 s from ANT. (e) and (f) are phase velocity maps for periods of 25 s and 40 s inverted by HT. 3km, 4km and 5km bathymetry contours are shown as thin grey lines. Trench axis and serpentine seamounts are labeled as in Figure 3.1.
Figure A3.4. Examples of Monte-Carlo inversion and phase velocity sensitivity kernel.

(a-d) The joint Rayleigh phase and group dispersion curves and one standard deviation error bars for four locations (a - Inner forearc, b - Outer forearc, c - Trench high, d - Pacific plate) and the computed phase (red solid) and group (blue solid) dispersion curves from the Bayesian Monte-Carlo averaged model (e-h). (i) Phase velocity sensitivity kernels at example periods, calculated base on the average velocity model in (g).
Figure A3.5. Robustness test of the low velocity zone (LVZ).

(a) The assumed subduction zone geometry according to our prior knowledge. Simulation results for nodes 80 km (b) and 110 km (c) landward from the trench. Black dashed line is the input 1-D model; Blue dashed and solid lines are the best fitting and average model from the Monte-Carlo inversion of the synthetic dispersion curves respectively; Red dashed and solid lines are the best fitting and average model from the Monte-Carlo inversion of the real data.
Figure A3.6. Comparison between Rayleigh wave isotropic phase velocities from teleseismic tomography using HT and two-plane-wave method. (a) at 27 s. (b) at 36 s.
Figure A3.7. Azimuthal anisotropy from evenly distributed serpentine layers (450 m thick with 2 km spacing).

(a) Result for vertical layering. (b) Result for 45° dipping layering. Numbers in the parenthesis are the mean velocity for quasi-P, quasi-SV and quasi-SH respectively. Incidence angle is defined relative to the layer’s strike, thus 0° is parallel and 90° is normal to the strike of the layer.
Chapter 4: Shear Velocity Structure near the Central Mariana Trench from Rayleigh Wave Tomography

4.1 Abstract

We investigate the crustal and uppermost mantle shear velocity structure across the Mariana trench by inverting Rayleigh wave phase and group velocities from ambient noise tomography along with longer period phase velocities from Helmholtz tomography of teleseismic waveforms. We use data from a temporary deployment in 2012-2013, consisting of 7 island-based stations and 19 broadband ocean bottom seismographs, as well as data from the USGS Northern Mariana Islands Seismograph Network. To avoid any potential bias from the starting model, we use a Bayesian Monte-Carlo algorithm to invert for the azimuthally-averaged SV-wave velocity at each node. This method also allows us to apply prior constraints on crustal thickness and other parameters in a systematic way. The results show the development of a low velocity zone within the incoming plate at least 120 km seaward of the trench axis, consistent with the earthquake locations and previous studies. The maximum depth of the velocity anomaly increases towards the trench, and extends to about 30±5 km below the seafloor. The low velocities persist after the plate is subducted, as a low velocity layer of similar thickness but a smaller velocity reduction is imaged along the top of the subducting mantle beneath the forearc. An extremely low velocity zone is observed beneath the serpentine seamounts in the outer forearc. Azimuthal anisotropy results show trench parallel fast axis within the incoming plate at depths immediately beneath the Moho. These observations suggest the velocity reduction in the incoming plate prior to
subduction results from both serpentinized normal faults and water-filled fault zones or cracks. Water is expelled from the cracks early in subduction, causing a modest increase in the velocity of the subducting mantle, and the water moves upward and causes serpentinization of the outer forearc mantle. Assuming the velocity anomaly remaining in the subducting plate mantle is caused by serpentinization, calculations suggest the top \(~24\pm5\) km of the slab mantle retains 18.4 vol\% serpentinization (\(~2\) wt\% water) beyond the outer forearc. The amount of water carried into the deep mantle by this layer (\(~94\pm17\) Tg/Myr/m) is about four times greater than previous estimates for the entire slab. If other old, cold subducting slabs contain correspondingly thick layers of hydrous mantle, as suggested by the similarity of incoming plate faulting, estimates of the global water flux into the sub-arc mantle must be increased by about a factor of three over previous estimates.

### 4.2 Introduction

Water is essential for many subduction zone processes, including magma production in arcs [Davies and Stevenson, 1992; Iwamori, 1998; Pearce and Peate, 1995; Schmidt and Poli, 1998] and overpressuring and weakening the thrust interface, allowing the downgoing plate to descend [Wang et al., 1995]. Elements transported by hydrous phases dominate many aspects of the geochemistry of igneous rocks formed in volcanic arcs [Tatsumi, 1989] and backarc [Kelley et al., 2006]. Water from the subducting slab may result in serpentinization of the forearc mantle [Bostock et al., 2002; Brocher et al., 2003; DeShon and Schwartz, 2004]. Intermediate and deep earthquakes are thought to be produced by processes associated with mineral dehydration in the subducting slab [Kirby et al., 2013; Meade and Jeanloz, 1991].
It is also increasingly clear that water is an essential factor in the geodynamics of the entire planet, and that subduction zones are the primary locations where water can be transported deep into the mantle [Hacker, 2008; Hirschmann, 2006; Kendrick et al., 2017; Rüpke et al., 2004]. Plate tectonics may be impossible, and “stagnant lid” convection the dominant mode of heat loss [Solomatov and Moresi, 1996], without the rheological weakening and fault zone “lubrication” provided by the widespread hydration of the surface lithospheric plates [Bercovici, 2003; Hirth and Kohlstedt, 1995; Korenaga, 2007]. Recently it has been realized that the earth’s transition zone could provide a repository for huge amounts of water, if the water can be delivered to significant depths by hydrous minerals in subducting slabs [Bercovici and Karato, 2003; Ohtani et al., 2004; Smyth and Jacobsen, 2013].

Despite its obvious importance, the water cycle at subduction zones remains poorly constrained. Previous modeling studies show that most of the water carried by the subducting slab is expelled at very shallow depth, and only a small and poorly-constrained fraction of the water can penetrate deeper than 150 km [Hacker, 2008; van Keken et al., 2011]. The large uncertainties in the amount of water subducted into the deeper mantle result from a lack of constraints on the water content of the downgoing mantle. The main purpose of this study is to constrain how much water was initially carried by the subducting slab mantle, and how much of this water can be brought deeper into the subduction zone.

It has often been assumed, for lack of other evidence, that the oceanic upper mantle is nearly anhydrous due to efficient water extraction by melt formation at the ridge. However, fluids flowing through normal faults formed when the plate bends into the trench have the potential to cause significant hydration of the downgoing plate [Peacock, 2001; Seno and Yamanaka, 2013]. Upper mantle peridotite reacts strongly with ocean water to form a variety of hydrous minerals
including serpentine, talc and brucite depending on temperature, pressure and bulk composition (e.g., SiO2 content). Serpentine (Mg3Si2O5(OH)4), hydrous phases containing 13 wt% water [Ulmer and Trommsdorff, 1995], are expected to be the most abundant hydrous minerals in the upper most mantle settings [Manning, 1995; Peacock and Hyndman, 1999; Wunder and Schreyer, 1997]. Lizardite and chrysotile are the more abundant serpentine minerals in hydrated mantle rocks formed at low temperatures and are stable up to ~320°C at 1 GPa [Evans, 2004; Guillot et al., 2015], whereas antigorite is the serpentine mineral stable at higher temperature (up to ~620°C at 1Gpa) [Guillot et al., 2015; Perrillat et al., 2005; Schwartz et al., 2013; Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997]. In addition, brucite coexists with serpentine in olivine-rich compositions, talc coexists with serpentine in pyroxene-rich compositions, and chlorite forms in compositions with substantial amounts of Al [Manning, 1995; Peacock and Hyndman, 1999].

Serpentinization of upper-mantle peridotite, especially formation of lizardite and chrysotile, dramatically affects seismic observables such as isotropic velocities and Poisson’s ratio [Christensen, 2004; Ji et al., 2013], making it detectable with seismic methods. Formation of parallel normal faults in the trench outer rise, by reactivation of abyssal hills during plate bending or by opening new cracks in the bend-normal direction if the bending direction is more than ~25° from the relic spreading direction [Masson, 1991; Shillington et al., 2015], is likely to create an extrinsic anisotropy. These serpentinized and/or water filled faults tend to have fast velocity direction along the fault-parallel direction [Crampin, 1984; Hudson, 1981; Miller and Lizarralde, 2016], thus identifiable by anisotropy measurements. Seismic evidence for serpentinization have been found within the subducted plate and the override fore-arc mantle at several convergent margins [Contreras-Reyes et al., 2007, 2011; DeShon and Schwartz, 2004;
Shillington et al., 2015], with serpentinization percentage ranging from 9% for southern central Chile [Contreras-Reyes et al., 2007] to ~30% for Tonga [Contreras-Reyes et al., 2011]. However, these previous studies utilized active source seismic imaging methods and are only able to image the upper few kilometers of the mantle, thus the depth extent of incoming plate hydration is poorly constrained. In addition, most studies consist of a profile in only one direction and are thus unable to estimate the azimuthal anisotropy property, leaving some ambiguities about the interpretation. Moreover, few of previous studies cover both the incoming plate prior to subduction and the forearc area, and thus fail to provide an understanding of how serpentinization and porosity change with pressure increasing during subduction. Finally, most of these studies investigate incoming plates with young or intermediate ages, where the thickness of mantle that is cold enough to allow serpentinization is limited.

We choose the Mariana subduction zone to study the water cycle in the incoming plate and forearc for several reasons. Firstly, the presence of numerous serpentine mud volcanoes in the outer fore-arc region of the Mariana subducting zone suggest the forearc mantle wedge must be partially serpentinized [Fryer, 1996]. A previous body wave tomography study of the region reveals a low shear velocity and high Vp/Vs ratio in the outer forearc region [Barklage et al., 2015], consistent with the serpentinized mantle wedge hypothesis. Secondly, Emry et al. [2014] and Oakley et al. [2008] identified various normal faults in the trench outer rise region. These faults can provide paths for the water to be transported deep into the uppermost mantle of the incoming plate. Thirdly, the age of the Pacific plate in this region (> 150 Ma) is greater than plates in any previously studied regions [Müller et al., 2008], thus this study is able to provide information on bend faulting hydration within extremely cold oceanic lithosphere. In this study, we utilize the surface wave method to investigate the shear velocity structure across the central
Mariana trench. Compared to previous studies covering either the incoming plate or the forearc region, our results covering both the incoming plate prior to subduction and the forearc region will help us to understand the subduction zone water cycle more completely. Compared to previous active source studies, our study resolves a 3-D shear wave structure to deeper depth and provides additional azimuthal anisotropy constraints as well.

4.3 Data and Methods

4.3.1 Data Collection

The data used in the group/phase velocity tomography were collected by 19 ocean bottom seismographs (OBSs) (yellow and red circles in Figure 4.1a) and 7 temporary island-based seismic stations (white squares in Figure 4.1a) deployed from January 2012 to February 2013. The OBS distribution covers both the trench-outer rise and the Mariana forearc region. Each OBS from Lamont-Doherty Earth Observatory (LDEO) contains a modified Sercel L-4C seismometer and a LDEO-designed datalogger, while the other OBSs from Scripps Institution of Oceanography (SIO) use Nanometrics Trillium 240/40 seismometers and Quanterra Q330 dataloggers. The land stations were deployed on various islands, using Nanometrics Trillium 120PA seismometers and Quanterra Q330/Reftek 130 dataloggers. We also used data from 3 island stations from the USGS Northern Mariana Islands Seismograph Network active over the same time period (red squares in Figure 4.1a). Our study region covers three major serpentine seamounts: Big Blue, Turquoise, and Celestial.
4.3.2 Rayleigh Wave Phase and Group Velocities from Ambient Noise Tomography (ANT)

We performed the ambient noise analysis following the procedures described by Bensen et al. [2007] and Lin et al. [2008]. The daily vertical seismograms were corrected for instrument responses, down-sampled to 2 samples per second (sps), and modified to have uniform polarity for all instrument types. We applied running-average time-domain normalization and spectral whitening to minimize the effects of large earthquakes. Cross correlations were then performed daily between all station pairs. After detecting and fixing clock errors, we stacked the daily cross-correlations for the 1 year time window when the maximum number of stations were operating.

We then applied frequency-time analysis (FTAN) [Bensen et al., 2007; Levshin and Ritzwoller, 2001] to the symmetric components of the stacked cross correlations to measure Rayleigh wave group and phase velocities between periods of 8 and 25 seconds. To extract the most reasonable phase velocity dispersion curves, we utilized an iterative approach to update the reference dispersion curve between station pairs. Initially, we used structures from previous studies [Nishimura and Forsyth, 1989; Pyle et al., 2010; Takahashi et al., 2008] to calculate the reference phase velocity dispersion curves. For the forearc, we combined crustal and mantle results from a previous refraction survey [Takahashi et al., 2008] and the surface wave study [Pyle et al., 2010] by assuming a Vp/Vs ratio of 1.79. For the incoming Pacific plate, we used the results for plate older than 110 Myr from [Nishimura and Forsyth, 1989]. For each station pair, we chose the forearc structure as the reference model if the great circle path between the stations mainly samples the forearc region; otherwise we used the incoming plate structure as the
reference model. The average water depth along the great circle was used to adjust the reference model before calculating the synthetic dispersion curve. For each frequency, only station pairs with distance larger than twice of the wave-length were kept. All dispersion curves were screened to exclude those with inconsistent measurements at adjacent periods.

For each period, a ray theory based tomography method [Barmin et al., 2001] was applied to dispersion measurements with signal-to-noise (SNR) ratio greater than 5 to produce Rayleigh group and phase velocity map on a grid of nodes spaced at 0.2°. We used an iterative procedure to converge on the best solution by modifying the damping parameter and eliminating outliers between iterations. We started with larger damping factors and excluded dispersion curves with misfit larger than 2 times of the standard deviation (2σ) from the mean misfit, and then redid the tomography with smaller damping factors. After obtaining the Rayleigh wave phase velocity maps, we extracted the average dispersion curve between station pairs and took this as the reference to redo the dispersion curve measurements. This procedure helped to minimize the 2π ambiguity of phase velocity measurements [Bensen et al., 2007]. We applied the tomographic inversion to the new dispersion curve measurements to update phase velocity maps and then repeat the iterative procedure described above one more time. The tomographic inversion returns both isotropic and azimuthal anisotropic components of the Rayleigh wave group and phase velocity (Figure 4.2).

4.3.3 Telesismic Rayleigh Wave Phase Velocities from Helmholtz Tomography (HT)

We utilized the multichannel cross-correlation technique described by [Jin, 2015; Jin and Gaherty, 2015] to determine phase velocities of longer periods from teleseismic Rayleigh
waveforms. Using the International Seismological Centre (ISC) catalogue, we selected seismograms from 380 earthquakes with surface-wave magnitudes (Ms) larger than 4.5 and epicentral distance between 25° and 150°, which occurred during the time when the stations were operating (Figure 4.1b). The raw seismograms of each event were cut from the origin time of the earthquake to 12000 s after. Then the vertical seismograms were down-sampled to 1 sps and instrument responses were corrected. Noise in seismograms at long periods (> 50s) due to ocean swell and associated water pressure variations, as well as tilt caused by local currents, were removed by correcting the vertical channel using horizontal and pressure channels [Bell et al., 2015; Crawford and Webb, 2000; Webb and Crawford, 1999]. A two-plane-wave tomography (TPWT) method was applied to the same dataset, and a frequency-independent station amplification correction was obtained for each station to account for any discrepancy in the nominal instrument responses as well as site effect [Yang and Forsyth, 2006a]. These station amplifications were then applied to all records (Table A4.1).

This tomography method recovers frequency-dependent phase and amplitude information via the narrow-band filtering of the broadband cross-correlations between the vertical component seismogram from a given station and time-windowed seismograms from all other nearby stations. The phase delays and amplitude information were determined by fitting the narrow-band filtered cross correlations with a Gaussian wavelet [Jin, 2015]. To eliminate the influence of poor-quality records, we estimated the coherence between waveforms from nearby stations for a series of periods from 21 to 53s, and only included those measurements with coherence larger than 0.5. We also incorporated an averaged 1-D phase velocity dispersion curve from two-plane-wave tomography as reference model (Figure A4.1a), and excluded individual phase delay measurements deviate more than 20% away from the predictions for short period (<=25 s) and
those more than 15% away from the predictions for longer periods. The incorporation of the 1D reference model also helped avoiding erroneous phase measurements arising from cycle skipping.

For each earthquake and at each period, we inverted the phase delays for spatial variations in dynamic phase velocity via the Eikonal equation [Lin et al., 2009]. We then further corrected the propagation effect via Helmholtz tomography [Lin and Ritzwoller, 2011a], producing maps of structure phase velocity with spacing of 0.2°. For the entire dataset, the azimuthal isotropic phase velocity at each node was obtained by fitting the mean within each 15° bin to the following equation [Lin and Ritzwoller, 2011b]:

$$c(\psi) = c_{iso} \left\{1 + \frac{A_{1psi}}{2} \cos(\psi - \psi_{1psi}) + \frac{A_{2psi}}{2} \cos(2(\psi - \psi_{2psi}))\right\}$$  \( (1) \)

where \( c_{iso} \) is the isotropic velocity, \( \psi \) is the azimuthal angle measured positive clockwise from the north, \( A_{1psi} \) and \( A_{2psi} \) are the peak-to-peak relative amplitude of \( 1\psi \) and \( 2\psi \) anisotropy, and \( \psi_{1psi} \) and \( \psi_{2psi} \) define the orientation of the anisotropy fast axes for the \( 1\psi \) and \( 2\psi \) component, respectively. We observed strong \( 1\psi \) anisotropy in our study (Figure A4.1b), which is possibly caused by backward scattering [Lin and Ritzwoller, 2011b]. This tomographic method returns both the azimuthal isotropic phase velocity and the azimuthal anisotropic component at each node simultaneously.

Phase velocities were also obtained by applying a TPWT method to the same teleseismic dataset. Phase velocities from HT and TPWT are comparable although differences are still noticeable (Figure A4.2). We calculated standard deviations of the velocity differences at each period, which later were used to define phase velocity uncertainties. Because the absolute phase velocities from the TPWT varies depending upon the chosen parameters of the tomographic
inversion such as damping coefficient and smoothing length, we thus only used isotropic phase
velocity obtained from HT method for later shear velocity inversion.

4.3.4 Teleseismic Rayleigh Wave Phase Velocities from Two-Plane-Wave
Tomography (TPWT)

We also applied the Two-Plane-Wave Tomography method to the same teleseismic dataset
(Figure 4.1b). The raw seismogram of each event was cut from the origin time of the earthquake
to 12000 s after. Prior to any further analysis, the vertical seismograms were down-sampled to 1
sps and instrument responses were corrected. Noise in seismograms at long periods (> 50s) due
to ocean swell and associated water pressure variations, as well as tilt caused by local currents,
were removed by correcting the vertical channel using horizontal and pressure channels [Bell et
al., 2015; Crawford and Webb, 2000; Webb and Crawford, 1999]. For each period of interest
from 21 to 53 s, we used a narrow bandpass filter (fourth-order Butterworth, zeros-phase shift)
centered at the interested frequency to filter the seismograms. The filtered data were then
windowed manually to isolate the fundamental mode of the Rayleigh Wave. We noticed the
significant amplitude difference between records from LDEO OBSs and records from other
stations, which cannot be purely caused by structure variations. This suggested considerable
amplitude corrections should be applied at least to data from LDEO OBSs (Table A4.1).

We utilized the two-plane-wave analysis method [Forsyth and Li, 2005] with a 2-D Fréchet
kernels [Yang and Forsyth, 2006b] to invert for Rayleigh wave phase velocity at each period for
both isotropic and anisotropic components. Unlike the traditional Rayleigh wave tomography
based on ray theory, this method considers scattering effects of Rayleigh waves outside the study
region by simplifying the scattered incoming wave as the sum of two interfering plane waves
Additionally, by using 2D Fréchet kernels based on the starting model, scattering and multipathing effects within the study region can be also approximately addressed [Yang and Forsyth, 2006b]. This method can also return station amplification factors at each period [Yang and Forsyth, 2006b]. Both the calculation of the 2D Fréchet kernels and the nonlinear tomographic inversion required a good starting model of phase velocity. In the first step that determined the average phase velocity for each period for the entire study region, we chose as the starting model the anisotropic model of the Pacific [Nishimura and Forsyth, 1989] with age older than 110 Myr. In the second step, we divided the study region into four sub-regions according to tectonic settings: the incoming Pacific plate, the Mariana fore-arc, the Mariana backarc and background. Then we used the average phase velocity as the starting model to invert the 2D phase velocity map.

To quantify the amplitude corrections for all stations, we first applied the TPWT method to only temporary land stations and SIO OBSs, those have better calibrated Nanometrics seismometers, and obtained average station amplification factors for these stations (Table A4.1). Results show amplification factors for these stations didn’t deviate much from 1. We then used these amplification factors as references, and estimated the average amplification factors for all other stations by comparing amplitudes of long period seismograms (> 60 s) (Table A4.1). Using data from all stations with amplitude corrected, we then conducted a series of TPW inversions with progressively finer grids (from 1.2° down to 0.6° and then 0.3°). This scenario can largely eliminate the dependence on the starting model and the a priori information [Rau and Forsyth, 2011; Wei et al., 2015]. The tomographic inversion returns both azimuthal isotropic phase velocity and the regional averaged azimuthal anisotropic component.
4.3.5 Combining Ambient Noise and Teleseismic Dispersion Curves

We combined the ANT and HT results to provide more complete measurements of phase velocity for the SV-velocity inversion (Figure 4.2). As these two methods are based on different principles and assumptions, the results of phase velocity at common periods from two methods are not always exactly the same [Wei et al., 2016; Yao et al., 2006] (Figure 4.3). The two sets of dispersion curves were combined in the geographical region well resolved by both methods. Phase velocities were interpolated onto a uniform grid of nodes spacing at 0.2° before being combined at each node. Group velocity results from ANT (8-21 s) were also included for SV-velocity inversion to better constrain water layer thickness and to improve resolution for shallower structure.

For phase velocities from ANT (8-25 s), the uncertainties were normalized at each period so that the uncertainty of the best-resolved node is 0.075 km/s. The uncertainties of the group velocity (8-21s) were normalized so that the best-resolved node has uncertainty of 0.188 km/s, 2.5 times the value for phase velocities, as suggested by Shen et al. [2016]. For phase velocities from HT (22-53 s), the uncertainties were normalized at each period so that the best-resolved node has uncertainty equal to the standard deviation of velocity differences between HT results and TPWT results (Figure A4.2). We utilized a linear weighting average method to combine phase velocity measurements and uncertainty estimates for overlapping periods (22-25s). A running average filter was then applied to make the resulting dispersion curve smoother.

4.3.6 SV-Wave-Velocity Inversion with a Bayesian Monte-Carlo Algorithm
Previous studies [Wei et al., 2015] suggest that results using linearized inversion of SV-wave velocity from phase velocity dispersion curves [Herrmann, 2013] have some dependence on the starting model. In order to avoid the potential bias of the starting model, to better apply prior constraints on crustal thickness and other parameters in a systematic way, and to derive formal estimates of velocity uncertainty, we use a Bayesian Monte-Carlo algorithm [Shen et al., 2013] to invert the azimuthal averaged SV-wave velocity at each node.

The Bayesian Monte-Carlo method constructs a priori distribution of SV velocity models at each node, defined by perturbations relative to the starting model and model constraints. Each model consists four layers on top of a half-space: (1) water with starting thickness from Gaussian smoothed (at 125 km length) bathymetry [Amante and Eakins, 2009] and an allowed perturbation of ±1.5 km (Figure A4.3), (2) sediments, (3) crust, and (4) upper mantle from the Moho to 180 km depth. The sedimentary layer is described by two parameters: a layer thickness of 0.5 km with an allowed perturbation of ±0.5 km and a constant VSV of 2.0 km/s with a perturbation of ±1.0 km/s. The crust is assumed to have linearly increasing velocity with depth, and is described by three parameters: a layer thickness and VSV at the top and bottom of the layer. For the incoming plate east of the trench, the crustal thickness is allowed to vary by ±1.5 km around the starting value 6.5 km based on preliminary seismic refraction results [Eimer et al., 2017]. The forearc crustal thickness is allowed to vary by ±3 km with starting values from a previous seismic refraction survey at the southern edge of the study region ranging from 19 km to 6.5 km [Takahashi et al., 2008] (Figure A4.4). The top and bottom crustal VSV are set at 3.0 and 3.2 km/s, respectively, with an allowed perturbation of ±1.0 km/s. The upper mantle VSV is parameterized by a B-spline, which is defined by 7 nodes with a perturbation of ±30% for first 5 nodes and 20% for the last 2 nodes. We imposed the constraint that the jumps in Vsv from the
sediment to the crust and from the crust to the mantle are positive. We also applied a physical dispersion correction with a reference period of 1 s [Kanamori and Anderson, 1977] using a 1-D Q model simplified from a seismic attenuation study in the same region [Pozgay et al., 2009]. Compared to the PREM, our 1-D Q model for the forearc has a high-attenuation layer in the uppermost mantle: Qs=60 from moho to 100 km depth. For the incoming plate region, the uppermost mantle is set to have a more regular attenuation: Qs=300 from Moho to 100 km depth.

For each grid node, the best fitting model is identified and models are accepted if their \( \chi^2 \) misfit is less than 50% higher than that of the best fitting model [Shen et al., 2013, 2016]. Following Shen et al. [2016], we also exclude models with mantle velocity higher than 4.9 km/s. The posterior distribution thus provides statistical information of all possible SV-velocity models that satisfied the Rayleigh wave dispersion curves within tolerances depending on data uncertainties. An average model is then calculated from all accepted models and used for plotting and interpretations [Shen et al., 2013; Wei et al., 2016]. Examples of the SV-velocity inversion at four representative nodes are shown in Figure 4.3. The results of the Bayesian Monte-Carlo inversion fit the measured group and phase velocity dispersion curves well.

4.4. Results

4.4.1 Group/Phase Velocity Maps

2-D maps of azimuthally averaged isotropic phase velocity for representative periods from ANT and HT are shown in Figure 4.2. At the shortest periods (10 s), phase velocities are more sensitive to the overriding water layer and crustal structures, so the lower speeds indicate a
thicker water layer (the incoming plate) or slower crust (the outer forearc) (Figure 4.2c). At 21s period, extremely slow velocities are observed beneath the trench outer-rise region and the Mariana outer fore-arc (Figure 4.2d), suggesting the uppermost mantle in this region has significant shear velocity reductions. In addition, the slow velocity zone tends to have a very sharp western boundary, corresponding well with the westward extent of major serpentine seamounts and disrupted forearc terrain. At longer periods, the slow velocity zone moves somewhat westward and its amplitude decreases, but it still can be observed at long periods beneath the fore-arc close to the arc (Figure 4.2e). Starting from 25 s, the fast velocity anomalies associated with the subducting slab are observed beneath the outer forearc (Figure 4.2e and f).

The Rayleigh wave group velocity maps presented in Figure 4.2 are derived exclusively from ANT. They are similar to the phase velocity maps except group velocity at a given period is sensitive to shallower structure than phase velocity. The checkmark shape group velocity dispersion curve for locations near the trench is the result of the thick water layer and is highly sensitive to the water depth (Figure 4.3c and 4.3d). At 10 s, high group velocities correspond well with shallow water locations such as the volcanic arc and Big Blue Seamount, the largest serpentine seamount in this area (Figure 4.2a), implying our results have relative good horizontal resolution. At 21s, the group velocity map is comparable to the phase velocity map, also showing an extremely slow velocity zone beneath the outer trench slope and the Mariana outer fore-arc (Figure 4.2b). The relatively stronger heterogeneities in the group velocity maps are accommodated by assigning larger uncertainties to the group velocity dispersion curves (Figure 4.3).

4.4.2 Azimuthally Averaged SV-Velocity
The azimuthally averaged SV-velocity structure near the Mariana trench obtained from the average model of the Monte-Carlo inversion at each node is shown in Figure 4.4 and 4.5. The model fits to the data are good, and the square root of the reduced \( \chi^2 \) misfit of the average model from the Bayesian Monte-Carlo inversion at each node is consistently less than 1 (Figure A4.5), suggesting the assigned uncertainties to phase and group velocity measurements are appropriate.

The group velocities, and to a lesser extent the shorter period phase velocities, are highly sensitive to the water depth. Since the dispersion curves at a given node is an average over the surrounding region, the water depth used in the inversion must represent an average depth rather than the depth at the node. To avoid biases of local sharp topography, we smoothed the bathymetry map [Amante and Eakins, 2009] with a Gaussian filter of 125 km and used the water depth from the smooth map at each node as starting depth (Figure A4.3). In addition, the water layer thickness is allowed to vary by up to 1.5 km in the inversion. After the inversion, the resulting water depth matches our expectations well: deep water concentrates to a smaller area right at the trench location and serpentine seamounts region has shallower water (Figure A4.3).

Without constraints from converted seismic waves (e.g. receiver functions), it’s difficult to determine the Moho depth precisely, however a qualitative analysis of the crustal thickness is still valuable. In the incoming plate region, the resulting crustal thickness changes little compared to the starting thickness of 6.5 km (Figure A4.4), and the changes are relatively smaller than the allowed perturbation ±1.5 km. This is compatible with active source data suggesting normal ocean crust thicknesses on the incoming plate. In the outer forearc, the uppermost mantle right beneath the Moho is characterized by ultra slow shear velocities (Figure
4.3f and A4.4d), resulting in a very small velocity contrast across the Moho. The Moho is thus no longer a distinct discontinuity, although the Monte-Carlo inversion prefers a thicker crust in the outer forearc (Figure A4.4). It is not clear how to define the moho discontinuity in the outer forearc, as discussed in a later section. In the inner forearc, the resulting crustal thickness is consistently slightly thinner than the starting model, but changes are still smaller than the allowed perturbation ±3.0 km (Figure A4.4). Shear velocity cross sections show the resulting crustal thickness agrees well with the previous P-wave refraction study [Takahashi et al., 2008] (Figure 4.5). Figure A4.4d shows the average shear velocity along the latitude at the top and bottom of the crust. The most significant feature is the ultra slow crust (2.4-2.8 km/s) in the outer forearc. The incoming crust has a faster top layer (~3.2 km/s) and a slower bottom layer (~3.6 km/s), compared to the inner forearc. No systematic change is observed in the incoming crust along the subduction direction.

The SV-velocity structure in the uppermost mantle doesn’t show prominent variations along the trench strike (Figure 4.4 and 4.5), we thus only describe the structure changes along the trench norm direction. The shear velocity cross sections consistently show a low velocity layer along the subducting slab beneath the Moho (Figure 4.5). At distance far from the trench axis (> 100 km), the incoming plate shows a typical structure of old oceanic lithosphere [Nishimura and Forsyth, 1989] at depths greater than 20 km, with velocities faster than 4.5 km/s (Figure 4.3h and 4.5). A somewhat lower velocity (~4.2 km/s) layer of ~10 km thickness is observed immediately beneath the Moho, consistent with active source seismic results [Feng, 2016]. A thicker region of low velocities begins about 80 km seaward of the trench axis and deepens towards the trench, with velocities as low as 3.8 km/s. At the trench axis, the bottom of this low velocity layer reaches 30±5 km beneath the sea floor, which is 24±5 km into the upper mantle (Figure 4.5).
This low velocity layer persists after the Pacific plate is subducted, as a 30±5 km thick low velocity layer (~4.1 km/s) is observed atop the fast slab between 100-160 km west of the trench axis. Such a large dimension velocity anomaly can’t be purely caused by the smearing of a 6 km thick crust (Figure A4.6). A distinct velocity increase from 3.8 km/s to 4.1 km/s is observed within the low velocity layer when the slab subducts to ~35 km depth, which is ~100 km west of the trench.

In the outer forearc, we observe extremely slow velocities (3.4-3.6 km/s) at depth shallower than 30 km within both the forearc mantle wedge and the subduction slab. This extreme low velocity zone has a very sharp western boundary, correlating well with the westward extent of major serpentine seamounts (Figure 4.4). This suggests the extreme low velocity zone is located beneath the serpentine seamounts. In the inner forearc region, with relatively flat bathymetry, a mantle wedge with fast velocity of 4.4-4.6 km/s was imaged between the inner fore-arc crust and the subducted Pacific plate (Figure 4.3e and 4.5).

### 4.4.3 Azimuthal Anisotropy

Seismic azimuthal anisotropy was determined simultaneously with the isotropic phase velocities by both the ANT method and the HT method. The magnitude of anisotropy from the ANT method varies depending upon the chosen inversion parameters such as smoothing length and damping coefficient, while the magnitude of anisotropy from the HT method is not significantly affected. Thanks to the good data quality, the HT and ANT methods return anisotropy results for some common periods, which enable us to compare anisotropy magnitude (Figure 4.6). At 21 s, when both the ANT method and the HT method can provide reliable anisotropy measurements, the anisotropy magnitude (3%-4%) from the ANT is very close to that
from the HT method at least in the incoming Pacific plate. This implies the magnitude of the anisotropy from the ANT method, as well as the fast directions, can be used for further analysis. Because our study area is relatively small and covers a tectonic boundary, we limit our anisotropy interpretations to more general features.

In the incoming Pacific plate east of the trench, azimuthal anisotropy results from the ANT method consistently show trench parallel fast axis between periods of 12-21 s, with a maximum magnitude as large as 9% for periods of 14-16 s. At periods longer than 27 s, the fast directions rotate to be oblique to the trench strike, close to the paleo-spreading direction [Oakley et al., 2008; Stern et al., 2004], which is the expected fast direction for anisotropy in the oceanic upper mantle [Forsyth, 1975; Lin et al., 2016]. The fast directions in the forearc at short periods show more heterogeneity and smaller magnitude compared to those observed in the incoming plate. At longer periods, the fast directions in the forearc are sub-parallel to the subduction direction (Figure 4.6f).

As the azimuthal anisotropy of the phase velocity in the incoming Pacific plate are relatively homogeneous within the well resolved region, we are able to investigate the average azimuthal anisotropy of the incoming plate close to the trench. We first obtain the average anisotropy amplitude and fast axis direction for the region between the trench and the outer-rise (within 80 km from the trench) from 10 to 36 s period (Table A4.2). Using only isotropic and $2\psi$ anisotropy terms in equation (1), we calculate phase velocities for different propagation directions (every 5° from 0° to 355°). Linear shear velocity inversions [Herrmann, 2013] are then performed to obtain the azimuthally-dependent SV velocity. We then obtain the shear velocity anisotropy from the variation of shear velocity with direction at different depths. The
inversion result (Table A4.2) shows that the Pacific plate crust has very strong azimuthal anisotropy (~18%) with fast direction parallel to the trench. The top 10 km of the uppermost mantle also shows azimuthal anisotropy with a trench parallel fast direction, but the anisotropy amplitude decreases with depth. The deeper part of the uppermost mantle turns to have fast axis along the paleo-spreading direction and with anisotropy amplitude close to previous studies for oceanic plate upper mantle [Forsyth, 1975; Kawasaki and Konno, 1984].

4.5 Discussions

4.5.1 Serpentinization vs Water-filled Cracks

Previous studies at various convergent margins attribute the slow velocity anomalies in the incoming plate or the forearc to the presence of serpentine without distinguishing the minerals involved [Contreras-Reyes et al., 2011; DeShon and Schwartz, 2004; Shillington et al., 2015]. Although the three dominant serpentine minerals, lizardite, chrysotile and antigorite, have same water storage capacity (~13 wt%) [Ulmer and Trommsdorff, 1995], their physical properties including seismic velocity [Christensen, 2004; Ji et al., 2013] and stability field [Evans, 2004; Guillot et al., 2015] are distinct due to the different crystal structure. Lizardite and chrysotile are the more abundant serpentine minerals in hydrated mantle rocks formed at low temperatures and are stable up to ~320°C at 1 GPa [Evans, 2004; Guillot et al., 2015]. When the temperature reaches between 320°C to 390°C, lizardite will be progressively replaced by antigorite at the grain boundaries and in the core of the lizardite meshes [Schwartz et al., 2013]. Antigorite is the sole stable serpentine mineral at higher temperature (up to ~620°C at 1GPa) [Guillot et al., 2015;
Perrillat et al., 2005; Schwartz et al., 2013; Ulmer and Trommsdorff, 1995; Wunder and Schreyer, 1997]. Lizardite and chrysotile have much lower shear velocities (~2.3 km/s) compared to antigorite (~3.7 km/s) at 600 MPa [Ji et al., 2013]. For the same amount of shear velocity reduction, the serpentinization percentage, and thus the water percentage, estimated with antigorite will be about 2.5 times larger than if lizardite and/or chrysotile are assumed [Ji et al., 2013]. This feature makes it greatly important to clarify the assumptions about which serpentine minerals are present before making any further interpretation of the observed velocity reductions. In addition to causing large seismic velocity reduction, serpentinization along the exposed normal fault faces will also generate significant azimuthal anisotropy as a result of shape-preferred orientation [Miller and Lizarralde, 2016] (Figure 4.6). Therefore, azimuthal anisotropy observations should also be considered when interpreting the velocity reductions in terms of serpentinization.

Korenaga [2017] argues that the same amount of velocity reduction can also be caused by water-filled porosity, without involving substantial bulk hydration. This argument presumes a non-fractured isotropic media, which is incompatible with the field observations in the Mariana subduction zone where numerous normal faults and normal fault earthquakes have been detected [Emry et al., 2014; Oakley et al., 2008]. Instead, a porous media with aligned cracks should be the more appropriate assumption [Gurevich, 2003; Hudson et al., 1996]. On the other hand, this argument can only be applied to the slab prior to subduction. When the slab starts to subduct, the confining pressure increases with increasing depth, causing closures of cracks and expulsion of free water within these cracks and/or porosity [David et al., 1994]. Thus this hypothesis can’t be applied to velocity reductions observed within the slab at greater depth after subduction.
Because the serpentinization rate is geologically fast at temperatures higher than 100°C [Reynard, 2013], the serpentinization rate is dominated by the rate of water delivery. Although a previous experimental study [Macdonald and Fyfe, 1985] suggests the diffusion speed of water in unfractured serpentine is as high as 1 km/Myr, this value becomes unrealistic when applied to any subduction setting. Taking the Mariana subduction zone as example, which has a subduction rate of ~ 50 mm/yr and has normal faults forming about 80 km from the trench, a 3.2 km thick serpentine layer will form around each normal faults when it reaches the trench. Some other factor must have been involved to prevent effective water transportation, and candidates can be the low permeability of the serpentine layer [Kawano et al., 2011; Reynard, 2013] and the large volume expansion (~50%) associated with serpentinization which can reduce pre-existing porosity.

Combining these facts and velocity variations observed in our study (Figure 4.4 and 4.5), we prefer to interpret the velocity reductions observed in the slab prior to subduction as a combined effect of serpentinization and water-filled cracks but attribute the remaining velocity reductions in the slab after subduction to serpentine only. This approach will give a lower bound estimation of the water volume carried by the subducting slab.

4.5.2 Pacific plate

The Pacific Plate at Central Mariana (15–18°N) has had many, recent, normal-faulting earthquakes. In this region, Oakley et al. [2008] observed that the incoming seafloor fabric is nearly parallel to the strike of the trench axis; in our study region between ~16.5 and 18°N, new faults and faults reactivating abyssal hill normal faults formed during seafloor spreading exist together. The existence of large normal earthquakes is evidence that a large amount of extension
is present within the shallow part of the incoming Pacific plate [Emry et al., 2014] which will balance the surrounding confining stress and allow the water to penetrate deeper into the slab along the normal faults [Faccenda et al., 2012]. The flexure model which best fits the Pacific plate bathymetry seaward of the trench axis predicts a neutral plane at ~30 km [Emry et al., 2014], suggesting the brittle normal faulting can continue down to ~30 km into the plate. This prediction agrees well with the maximum depth extent (30±5 km) of the low velocity zone (LVZ) in the trench outer-rise region in our study (Figure 4.5), suggesting this LVZ can be directly related to brittle normal faults. The plate flexure model in Emry et al. [2014] also predicts that the area with higher extension stress is thickening toward the trench, which means normal faults form due to increasing tensor can rupture to greater depth when the slab moves toward the trench. This is consistent with the gradually thickened LVZ geometry observed in our study. In addition, the point at which the velocity begins to decrease, 80 km from the trench, corresponds well with where the faults begin on the seafloor [Oakley et al., 2008] and where the intense seismicity begins [Eimer et al., 2018]. These correlations all suggest that the LVZ in the trench outer rise region is closely related to the new ruptured or reactivated normal faults.

The anisotropy inversion result shows only the top 10 km of uppermost mantle in the incoming plate is characterized by trench-parallel anisotropy, thinner than the thickness of the mantle LVZ (~24 km). Two factors may contribute to this discrepancy. First, this is an average anisotropy result for the region between the trench and the outer-rise, where the thickness of the mantle LVZ also changes significantly from ~15 km to ~24 km. Second, minor cracks and fractures deeper in the slab may not be well orientated, thus can’t generate significant anisotropy.

Compared to the LVZ within the slab prior to subduction, the LVZ within the subducted slab arcward of the serpentine seamounts (>100 km west of the trench) preserves the original
thickness (30±5 km) but only has reduced amounts of velocity reductions (~4.1 km/s) (Figure 4.5). At this depth (> 35 km beneath the sea floor), water-filled cracks and/or porosities can’t be the cause for the velocity reduction, because the increased confining pressure will effectively close cracks and porosities [David et al., 1994], leaving the presence of serpentine the dominant factor. In addition, the temperature of the slab mantle around the Moho is still low at this depth [van Keken et al., 2011], lower than the temperature when the lizardite starts to be unstable (~320°C) [Schwartz et al., 2013]. Thus, it’s reasonable for us to choose lizardite to interpret the velocity reduction. Using the experimental relationship $V_s = 4.51 - 2.19\Phi$ between shear velocity and serpentine volume fraction (Φ) for lizardite at 600 MPa [Ji et al., 2013], ~4.1 km/s shear velocity within the subducted slab can be caused by ~18.7 vol% serpentinization (~2 wt% water). When a relationship for higher pressure is used, the water percentage will be slightly higher [Ji et al., 2013]. Because no significant dehydration should happen within the slab mantle at this shallow depth and low temperature [Evans, 2004; Guillot et al., 2015], it’s reasonable for us to assume the slab prior to subduction should have same amount of serpentine. The additional velocity reduction (~0.3 km/s) within the LVZ in the slab prior to subduction should be attributed to other factors like anisotropy effect caused by water-filled cracks [Miller and Lizarralde, 2016].

One unexpected feature in our results is the ~10 km thick low velocity layer (~4.2 km/s) observed right beneath the Moho at the location where the slab is still far from the trench and experiencing little tension (Figure 4.5). A previous refraction survey [Feng, 2016] also imaged a layer right beneath the Moho with a slightly slower P wave velocity (7.9-8.1 km/s) compared to normal oceanic lithosphere. The layer thickness estimated from the higher resolution active source study is ~ 8km, very close to our 10 km observation, suggesting our shear velocity results
have very good vertical resolution. Fujie et al. [2013] observed sharp changes of Vp/Vs ratio, a sensitive indicator of water content, in the slab starting from 140 km away from the Kuril trench, although their results have relatively poor constraints for the uppermost mantle. In a latest seismic study using the same dataset as this study [Eimer et al., 2018], the furthest event is located more than 170 km away from the Mariana trench (Figure A4.7). In addition, we also found well relocated events more than 120 km away from the trench in the ISC bulletin (Figure A4.7). These observations suggest that faults/cracks may start to form and the incoming slab may start to be altered far away from the trench.

4.5.3 Overriding Forearc

The Mariana forearc contains more abundant evidence of mantle serpentinization than any other arc, with large serpentinite seamounts, composed primarily of serpentine mud and clasts, and partially serpentinized and metamorphosed mafic and ultramafic rocks [Fryer, 1996; Oakley et al., 2008; Stern et al., 2004]. Previous body wave study [Barklage et al., 2015] and long period surface wave study [Pyle et al., 2010] imaged a fore-arc mantle wedge characterized by slow shear velocity anomalies. Our study reveals similar structure with higher resolution and less vertical smearing (Figure 4.5). Our model shows an extremely low velocity zone (ELVZ) of velocity 3.4-3.6 km/s in the Mariana outer forearc, which is located to be beneath the serpentine seamounts only and possibly extends into the Pacific plate (Figure 4.4 and 4.5). This ELVZ doesn’t extend deeper than 30 km and its geometry shows no correlation with 320°C or 600°C isotherms [van Keken et al., 2011]. Its distinct western boundary also suggests that this feature cannot be dominated by temperature variations. Instead, the water accessibility could be the controlling factor. Boron and Chlorine isotope studies of serpentine samples from various
Mariana serpentine seamounts suggest the water in the Mariana forearc mantle should have a slab source [Barnes et al., 2008; Benton et al., 2001]. If we attribute all the velocity reduction to serpentine [Ji et al., 2013], the Mariana outer forearc mantle is about ~46 vol% serpentinized (~5 wt% water).

Within a reasonable temperature range, the serpentinization percentage is determined by how much water can be transported to the reaction front [Reynard, 2013]. The much higher percentage of serpentinization in the Mariana outer forearc than that in the incoming Pacific plate implies the Mariana outer forearc has more sufficient water supply, maybe from the subducting plate. The subducted seamounts are able to cause pervasive fracturing in the upper plate [Dominguez et al., 1998]. The arc-parallel extension in the Mariana forearc, caused by the ‘bowing-out' of the arc associated with the crescent-shaped opening of the Mariana Trough, can also cause more fracturing [Stern and Smoot, 1998]. The more intensive fracturing feature in this region may explain why serpentine seamounts are more widely distributed in the Mariana forearc than any other subduction zones.

4.5.4 Water Cycle at the Mariana Subduction Zone

According to our observations, the water cycle at the Mariana subduction zone can be generally summarized as three stages. In the first stage, when the Pacific plate moves toward the Mariana trench, abyssal hill normal faults formed during seafloor spreading are reactivated and new normal faults form as well [Emry et al., 2014; Oakley et al., 2008]. These normal faults provide paths for the seawater to move downward to the slab mantle depth. Water reacts with surrounding peridotite and serpentinizes the top 24±5 km of the slab mantle to ~18.7 vol% (~2 wt% water) when the Pacific plate reaches the Mariana trench (Figure 4.5). Although water
supply is unlimited, the slab mantle can’t be fully serpentinized due to the limited accessibility to water at the reaction front. Water-filled cracks and porosities coexist with serpentine in the slab, causing some extra velocity reduction. In the second stage, the Pacific plate starts to subduct. Confining pressure increases with subducting depth, causing closure of cracks and porosities. Due to the low permeability of the overriding plate and the permeability anisotropy of the serpentine [Kawano et al., 2011; Reynard, 2013], water expelled from the cracks moves along the slab interface instead of vertically until it reaches the Mariana outer forearc. The Mariana outer forearc is highly fractured and permeable. Water moves upward along fractures and cracks in the outer forearc, and serpentinizes the outer forearc mantle to a higher degree (Figure 4.5). In the third stage, the 24±5 km thick serpentinized (~2 wt% water) slab mantle subducts together with the whole Pacific plate to the deeper mantle.

To get a lower bound estimation of the water flux input, we assume a 24±5 km thick partially serpentinized (2 wt% water) slab mantle. Applying a convergence rate of 50 mm/yr, the amount of water input into the Mariana subduction zone through mantle serpentinization would be ~79±17 Tg/Myr/m, and the total water flux is ~94±17 Tg/Myr/m if water in the sediment and crust is also included [van Keken et al., 2011]. This new estimation is 4.3±0.8 times larger than the estimation from van Keken et al. [2011] with only 2 km thick partially serpentinized slab mantle (2 wt % water). All uncertainties are estimated based on the uncertainty of the serpentinized slab mantle thickness.

### 4.5.5 Estimation of Global Subduction Zone Water Flux

Our interpretation of serpentinization extending to depths of ~24 km below the Moho in the incoming plate at the Mariana Trench has significant implications for water flux into subduction
zones globally. This depth is somewhat greater than the maximum observed depth of large normal faulting earthquakes and close to the estimated depth of the neutral plane for the incoming plate at the Mariana Trench [Emry et al., 2014]. The maximum depth of serpentinization near trenches has not been well determined for other older incoming plates, as the depth extent may be too great to be well constrained with active source seismic studies [Fujie et al., 2013; Grevemeyer et al., 2018; Ranero et al., 2003; Shillington et al., 2015; Van Avendonk et al., 2011] and surface wave investigations have not been performed elsewhere. The bending and faulting features of the incoming Pacific plate in Mariana are similar to what is observed at other old subduction plates, and the maximum depth of normal faulting and depth of the neutral plane is generally about the same depth [Emry and Wiens, 2015], so it is possible that serpentinization extends to similar depths of 20-25 km below the Moho at other sites where old lithosphere subducts.

We recalculated the global subduction zone water flux based on a previous estimation from van Keken et al. [2011], by re-evaluating the water content in the slab mantle. As serpentine minerals are stable up to 620°C, young and warm subducting plates have less potential to be serpentinized to significant depth. According to thermal models for oceanic plates [Stein and Stein, 1992], only plates older than about 40 Ma have 600°C isotherm deeper than 30 km beneath the seafloor, thus we set 40 Ma as a threshold for subducting plate age to be affected by deeper serpentinization. For subduction zones with subducting plate younger than 40 Ma, we simply take the water flux estimations from van Keken et al. [2011]. For subduction zones with incoming plate older than 40 Ma, we assume the slab mantle is partially serpentinized (2 wt% water) to 20 km below the Moho and keep the water volume in the sediment and crust unchanged as van Keken et al. [2011]. This rough estimation suggests the global subduction zone
water flux should increase to $3.0 \times 10^9$ Tg/Myr, which is about three times greater than previous calculations [Hacker, 2008; Rüpke et al., 2004; van Keken et al., 2011].

This new larger estimate of the input water flux from subduction zones is much greater than current estimates of water output from the mantle. Since a large long-term net influx of water to the deep interior is inconsistent with the stability of sea level in the geological record [Parai and Mukhopadhyay, 2012; Rüpke et al., 2004], one possible implication of this result is that the thick layer of serpentinized mantle we find in Mariana is not characteristic of other old, cold subducting slabs, and the Mariana slab carries much more water than other subduction zones. However, there is little indication that the Mariana incoming plate-bending region is significantly different in terms of morphology and intensity of faulting compared to the corresponding regions of other old subduction zones. Thus the most likely interpretation is that previous estimates of water output from the mantle are also underestimated. Estimates of mantle water output from mid-ocean ridges and ocean islands may be relatively well constrained, but estimates for volcanic arcs and backarcs rely on the melt flux and the water content, which are poorly constrained [Grove et al., 2012; Parai and Mukhopadhyay, 2012].

### 4.6 Conclusions

We obtained a comprehensive image of the SV-wave velocity structure across the Mariana trench by jointly inverting the Rayleigh wave phase/group velocities from ANT and HT with a Bayesian Monte-Carlo method [Shen et al., 2013]. The SV velocity structure was then further analyzed to understand the water cycling at the Mariana subduction zone. A LVZ is imaged
within the incoming plate prior to the trench down to about 30±5 km, which is a result of combined water-filled cracks and mantle serpentinization along the normal faults reactivated or newly created in the trench outer rise. The LVZ preserves the thickness after the slab is subducted, but the velocity reductions become smaller. Additionally, the outer forearc shows extremely low velocities beneath the serpentine seamounts. Isotope studies suggest the outer forearc mantle was serpentinized by water with a slab source \cite{Barnes et al., 2008; Benton et al., 2001}. One possible source is the water trapped in the sediments, in the crust and within the water-filled cracks or porosity into the slab mantle. Results of azimuthal anisotropy show fast directions parallel to the trench at shallow depths in the incoming Pacific plate prior to subduction, consistent with both the serpentinized normal faults hypothesis and the water-filled cracks hypothesis. The water input into the Mariana subduction zone from our estimation is ~94±17 Tg/Myr/m, 4.3±0.8 times larger than a previous estimation with a thin partially serpentinized slab mantle \cite{van Keken et al., 2011}. If other old, cold subducting slabs contain correspondingly thick layers of hydrous mantle, as suggested by the similarity of incoming plate faulting, estimates of the global water flux into the sub-arc mantle must be increased by about a factor of three over previous estimates \cite{van Keken et al., 2011}. Since a long-term net influx of water to the deep interior is inconsistent with the geological record \cite{Parai and Mukhopadhyay, 2012}, it is likely that estimates of water expelled at volcanic arcs and back arc basins also must be revised upward \cite{Grove et al., 2012}.

References


Pozgay, S. H., D. A. Wiens, J. A. Conder, H. Shiobara, and H. Sugioka (2009), Seismic attenuation tomography of the Mariana subduction system: Implications for thermal...


Figure 4.1. Maps of station and earthquakes used in this study.

(a) Map of seismic stations used in this study. Yellow and red circles are SIO and LDEO OBSs deployed from January 2012 to February 2013, respectively. White squares represent the temporary island-based stations. Red squares are stations from USGS Northern Mariana Islands Seismograph Network active during the same time period, open triangles are locations of three main serpentine seamounts covered by our study. White dashed line is the trench axis. Magenta stars illustrate nodes used as example shown in Figure 4.3. White solid lines show the cross-sections in Figure 4.5. (b) Earthquakes (blue dots) used in this study in a plot centered on the Mariana trench (red star).

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Figure 4.2. Maps of azimuthally averaged group and phase velocity.

(a) and (b) show group velocity at periods of 10 s and 21 s inverted by ANT. (c) and (d) show phase velocity at periods of 10 s and 21 s from ANT. (e) and (f) are phase velocity maps for periods of 25 s and 40 s inverted by HT. 3km, 4km and 5km bathymetry contours are shown as thin grey lines. Trench axis and serpentine seamounts are labeled as in Figure 4.1.
Figure 4.3. Examples of Monte-Carlo inversion and phase velocity sensitivity kernel.

(a-d) The joint Rayleigh phase and group dispersion curves and one standard deviation error bars for four locations (a - Inner forearc, b - Outer forearc, c - Trench high, d - Pacific plate) and the computed phase (red solid) and group (blue solid) dispersion curves from the Bayesian Monte-Carlo averaged model (e-h). (i) Phase velocity sensitivity kernels at example periods, calculated base on the average velocity model in (g).
Figure 4.4. Maps of azimuthally averaged SV-velocity at 20, 30, 40, 50, 60, 100 km depths relative to the sea floor. Trench axis, serpentine seamounts, and bathymetry contours are labeled as in Figure 4.1.
Figure 4.5. Cross sections A-A’, B-B’ and C-C’ showing the azimuthally averaged SV-velocity.

White dashed lines are the forearc Moho location from [Takahashi et al., 2008]. Thick white lines are projected 6-km thick slab crust. Thin white lines are contours of 3.6 and 3.8 km/s, and thin black lines are contours of 4.1 and 4.2 km/s. Black circles are relocated earthquakes in the subducting plate around each profile [Emry et al., 2011].
Figure 4.6. Azimuthal anisotropy results at various periods.

At 12 s (a), 14 s (b), 16 s (c), and 18 s (d), only results from the ANT method are plotted (red bars); at 21 s (e), results from both the ANT method and the HT method (yellow bars) are plotted; at 27 s (f), only results from the HT method are plotted. Trench axis and serpentine seamounts are labeled as in Figure 4.1.
Appendix

Figure A4.1. HT phase velocity perturbation range and anisotropy at an example node.
(a) 1-D averaged phase velocity dispersion curve for the entire study area from TPTW (solid line) and perturbation allowance when used as a reference model for HT (error bars). (b) Directionally dependent phase velocity measurements (green dots with error bar) of 21 s Rayleigh wave from HT. The solid line shows the best-fitting curve based on equation 1. Red dots are the measurements after subtracting the $1\psi$ anisotropy, and blue dashed line are the best fitting curves for the $2\psi$ anisotropy.
Figure A4.2. Comparison between Rayleigh wave isotropic phase velocities from teleseismic tomography using HT and TPWT method at 27 sec (a) and 36 sec (b).
Figure A4.3. Water depth of the Monte-Carlo inversion.

(a) Starting water depth is adopted from a smoothed bathymetry map (Lindquist et al., 2004). (b) Water depth of the Monte-Carlo inversion results. (c) Changes in water depth of the results compared to the starting models.
Figure A4.4. Crustal structure of the Monte-Carlo inversion.

(a) Starting crustal thickness is adopted from a previous refraction survey [Takahashi et al., 2007] and preliminary results of the 2012 Mariana refraction survey [Eimer et al., 2017]. (b) Crustal thickness of the Monte-Carlo inversion results. (c) Changes in crustal thickness of the results compared to the starting models. (d) Vs at the top (red dots) and bottom (blue dots) of the crust and the averaged value along the latitude (red and blue line).
Figure A4.5. Misfit of the Monte-Carlo inversion.

(a) Misfit of the Bayesian Monte-Carlo inversion to the observed Rayleigh wave phase and group velocity curves at each node. The misfit is defined as the square root of the reduced $\chi^2$ misfit of the average model. (b) Histogram of the misfit. Average misfit is about 0.8.
Figure A4.6. Robustness test of the low velocity zone (LVZ).

(a) The assumed subduction zone geometry according to our prior knowledge. Simulation results for nodes 80 km (b) and 110 km (c) landward from the trench. Black dashed line is the input 1-D model; Blue dashed and solid lines are the best fitting and average model from the Monte-Carlo inversion of the synthetic dispersion curves respectively; Red dashed and solid lines are the best fitting and average model from the Monte-Carlo inversion of the real data.
Figure A4.7. Distribution of earthquakes in the Pacific plate near the Mariana trench.

(a) Map view of earthquake locations. Only earthquakes more than 50 km from the trench are plotted. Black dots are earthquakes occurred between 2012 and 2013, and are relocated with local arrivals [Eimer et al., 2018]. Red dots are earthquakes from the ISC Bulletin (1960-2015), only earthquakes recorded by more than 20 stations and with horizontal uncertainty less than 20 km are kept. (b) and (c) Histograms for ISC bulletin and local relocation result respectively.
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**Table A4.2.** Average phase velocity azimuthal anisotropy as a function of period and shear velocity azimuthal anisotropy as a function of depth for the trench outer-rise region.

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Chapter 5: Shear Velocity Structure of the Mariana Subduction System from the Arc to the Backarc

5.1 Abstract

We investigate the crustal and uppermost mantle shear velocity structures across the Mariana arc and backarc by inverting the Rayleigh wave phase and group velocities from ambient noise tomography along with longer period phase velocities from two-plane-wave tomography of teleseismic waveforms. We use data collected by two separate ocean bottom seismograph deployments in 2003-2004 and 2012-2013, to provide good coverage from the Pacific plate to the West Mariana Ridge. To avoid any potential bias from the starting model, we use a Bayesian Monte-Carlo algorithm to invert for the azimuthally-averaged SV-wave velocity at each node. This method also allows us to apply prior constraints on crustal thickness and other parameters in a systematic way. A thick deep low velocity zone (LVZ) is imaged beneath the Mariana trough with varying depth. We find a small size slow velocity anomaly beneath the volcanic arc near the top of the mantle. An inclined LVZ is observed west of the arc and connected to the deep LVZ. We also find a shallower thin slow velocity layer at the top of the mantle distributed in a narrow channel along the central part of the spreading center, where larger magma supply is suggested [Kitada et al., 2006]. The shallow slow velocity anomaly beneath the volcanic arc and the backarc spreading center may both represent a shallow melt reservoir. However, neither of them is imaged to be directly connected to the deep LVZ, suggesting the melt may be transported through a narrow zone [Forsyth, 1996; Matsuno et al.,]
In the backarc, results of azimuthal anisotropy show fast directions are consistently sub-parallel to the opening direction of the Mariana trough [Deschamps and Fujiwara, 2003], with maximum anisotropy observed between 50 km and 90 km, correlating well with the deep LVZ imaged in the upper mantle.

5.2 Introduction

Subduction zone is a key component of Earth’s dynamic system and essential for understanding the processes that control the dynamic evolution of the mantle. These processes include water circulation, arc magmatism, and the formation of new oceanic lithosphere at back-arc spreading centers. Many questions remain including spatial extent of melt production and transport beneath volcanic arcs and back-arc spreading centers, the distribution of volatiles, and the temperature structure of the mantle wedge. The study of seismic velocities, anisotropy and attenuation provides important constraints on these processes. High-resolution body wave tomography results show significant low velocity anomalies beneath the volcanic front and above the slab in several subduction zones [Barklage et al., 2015; Husen et al., 2003; Reyners et al., 2006; Wagner et al., 2005; Wiens et al., 2008]. Surface wave tomography from the Lau back-arc shows a wide low-velocity zone beneath the Lau Basin, indicating upwelling hot asthenosphere with extensive partial melting [Wei et al., 2015; 2016; Zha et al., 2014]. The direction of mantle flow is usually inferred from shear wave splitting results or surface wave azimuthal anisotropy results, which can vary significantly in different subduction zone environment [Long, 2013; Menke et al., 2015; Pozgay et al., 2007; Wei et al., 2016].
The Mariana subduction zone provides an excellent setting to evaluate subduction zone processes because it includes a wide variety of tectonic features, including an extremely old subducting plate, a highly faulted forearc with actively venting serpentinite seamounts [Fryer et al., 1985; 1999], a remnant arc, an active modern volcanic arc and back-arc spreading center [Stern et al., 2004]. The intraoceanic environment also prevents contamination of continental crust for geochemical studies of arc and backarc magmatism. At the trench, the old Pacific plate (> 150 Ma) subducts beneath the Philippine Sea plate with a direction of N80°W and a rate of ~4.5 cm/yr in the region of this study [Seno et al., 1993]. The subducted slab becomes nearly vertical below depths of about 250 km [Engdahl et al., 1998] and no slab rollback is observed [Schellart et al., 2007; Stern et al., 2004]. The spreading rates vary along the back-arc spreading center, but all are very slow (~25 mm/yr at 18°N) [Seama and Fujiwara, 1993; Kato et al., 2003]. Geochemical studies have shown slab-derived volatiles and fluid mobile elements at the arc and the backarc along with variation along the arc chain and the backarc axis [Kelley et al., 2006; Newman et al., 2000; Pearce et al., 2005; Stolper and Newman, 1994; Taylor and Martinez, 2003], suggesting complex mantle flow and localized mantle upwelling in the Mariana Trough.

A long seismic refraction survey line covering from the West Mariana Ridge to the Mariana forearc at around 17°N shows Mariana trough Pn velocities of 7.8-8.0 km/s and arc/forearc Pn velocities of 7.7 km/s [Takahashi et al., 2007; 2008]. A 2-D wide-angle survey in the region surrounding the volcanic arc suggests maximum crustal thickness of 20-23 km beneath the remnant frontal arc [Calvert et al., 2008]. A passive magnetotelluric survey using ocean bottom electro-magnetometers (OBEM) across the Mariana system at around 18°N shows a resistive region down to 60 km depth and a conductive region at greater depth beneath the volcanic arc [Matsumo et al., 2010]. Beneath the Mariana Trough, an uppermost resistive layer with a
thickness of 80-100 km is observed along with a conductive layer beneath it. There is no
evidence for a conductive feature beneath the backarc spreading center, suggesting any melt
present is poorly connected or discontinuously distributed [Matsuno et al., 2012; 2010].

Passive seismology results from this region show many interesting features related to
subduction zone processes. Surface wave result shows the top 24 km of Pacific plate upper most
mantle is hydrated to a significant amount (~2 wt% water) right before subduction [Cai et al.,
2018], suggesting the slab is carrying down much more water than we thought. Slow velocity
anomalies were also observed beneath the serpentine seamounts in the outer forearc region [Cai
et al., 2018; Pyle et al., 2010; Tibi et al., 2008]. Previous body wave tomography in this region
indicates the mantle wedge is characterized by slow velocity and high $V_p/V_S$ beneath the forearc,
an inclined zone of slow velocity underlying the volcanic front, and a strong region of slow
velocity beneath the back-arc spreading center [Barklage et al., 2015]. Body wave attenuation
tomography [Pozgay et al., 2009] shows a 75 km wide columnar-shaped high-attenuation
anomaly ($Q_p$ 43–60) beneath the spreading center that extends to 100 km depth. A separate
weaker high-attenuation region ($Q_p$ 56–70) occurs beneath the volcanic arc, and the two
anomalies are connected at depths of 75–125 km. Shear wave splitting measurements show
along-strike fast directions in the forearc/arc with fast directions rotating to convergence parallel
near the back-arc spreading center [Pozgay et al., 2007]. P wave receiver functions image
subducting oceanic crust at depths of 75–110 km indicating the basalt to eclogite phase transition
must occur at a greater depth [Tibi et al., 2008].

In this study, we combine data collected by two separate deployments extending from the
Pacific plate to the West Mariana Ridge (Figure 5.1a), and derive shear velocity structure by
inverting Rayleigh phase and group velocities. Rayleigh wave phase and group velocities at short
period are obtained by the ambient noise tomography \cite{Bensen et al., 2007; Lin et al., 2008}, and Rayleigh wave phase velocities at long period are obtained by the two-plane-wave tomography \cite{Forsyth and Li, 2005; Yang and Forsyth, 2006a}. The incoming plate structure has been determined and discussed in detail in \cite{Cai et al., 2018}. We here only study and interpret the SV-velocity structure for the Mariana forearc, the volcanic arc, and the Mariana back-arc. We then compare the results to other arc and backarc systems such as the Lau basin, and interpret the results in terms of differences in spreading rate and regional geodynamics.

## 5.3 Data and Methods

### 5.3.1 Data Collection

Most of the data used in this study was collected by the 2003-2004 Mariana Subduction Factory Imaging Experiment. This 11 month experiment consists of 20 broadband land stations and 58 semi-broadband ocean bottom seismographs (OBSs) (Figure 5.1a). The land stations were deployed on various Mariana islands, using either STS-2 or Guralp CMG-40T sensors. The OBSs were densely distributed around the island of Pagan and over the active back-arc spreading center and extended more sparsely from the forearc to the West Mariana Ridge (Figure 5.1a). Fifty OBSs were operated by Lamont Doherty Earth Observatory (LDEO) and used three component Mark Products L4 sensors with 1 Hz natural period and modified amplifiers to extend long-period performance \cite{Webb et al., 2001}; 15 of the instruments had 16 bit data loggers and 35 of the instruments has newer 24 bit data loggers. Another eight OBSs equipped with precision measuring devices (PMD) semi-broadband sensors were developed and operated by the
University of Tokyo [Shiobara and Kanazawa, 2008]. Forty-nine of the OBSs were recovered and returned useful data. Of these, 29 LDEO OBSs with the 24 bit data loggers returned data for only approximately 50 days due to a firmware defect (Figure 5.1a). The remaining OBSs returned data for the entire duration of the experiment. In order to improve the coverage for the volcanic arc and forearc, data collected by a more recent deployment from January 2012 to February 2013 was also included in the phase/group velocity inversion (Figure 5.1a) [Cai et al., 2018]. The 2012-2013 deployment consists of 7 island stations and 20 OBSs, including 10 OBSs from LDEO covering the Mariana forearc and another 10 OBSs from Scripps Institution of Oceanography (SIO) covering the trench-outer rise. OBSs from LDEO contain a modified Sercel L-4C seismometer and LDEO-designed dataloggers, while the other OBSs from SIO use Nanometrics Trillium 240/40 seismometers and Quanterra Q330 dataloggers.

5.3.2 Rayleigh wave Phase and Group Velocities from Ambient Noise Tomography (ANT)

We performed the ambient noise analysis following the procedures described by Bensen et al. [2007] and Lin et al. [2008]. The daily vertical component seismograms were corrected for instrument responses, down-sampled to 2 samples per second (sps), and corrected to have uniform polarity for all instrument types. We applied running-average time-domain normalization and spectral whitening to minimize the effects of large earthquakes. The daily seismograms were then cross correlated between all station pairs.

We applied a time-frequency domain phase-weighted stacking (tf-PWS) algorithm to enhance the coherent surface wave signals [Li et al., 2017; Schimmel and Paulssen, 1997]. Compared to the more commonly used linear stacking method [Bensen et al., 2007; Lin et al.,
2008], the tf-PWS can suppress incoherent noise more efficiently, and enables phase/group velocity extraction from cross correlations between stations with only short overlapping time (Figure A5.1). The phase-weighted stacking proved to be particularly valuable in this study since many of the OBSs only collected data for 50 days. All individual causal and acausal components of daily cross corrections were stacked to achieve the most of the phase coherence. As suggested by [Li et al., 2017], a power of 1 is chosen for the coherence weighting term to mitigate possible waveform distortions.

Frequency-time analysis (FTAN) [Bensen et al., 2007; Levshin and Ritzwoller, 2001] was then applied to the stacked cross correlations to measure Rayleigh wave group and phase velocities between periods of 6 and 30 second. Compared to previous OBS ambient noise studies [Cai et al., 2018; Wei et al., 2016; Zha et al., 2014], introducing of the tf-PWS enabled us to extend measurements to shorter and longer period (Figure A5.1). To extract the most reasonable phase velocity dispersion curves, we utilized an iterative approach by updating the reference dispersion curve between station pairs [Cai et al., 2018; Shen et al., 2013]. Initially, we used the 1-D average dispersion curve from a previous surface wave study [Pyle et al., 2010] as the reference phase velocity dispersion curve. All dispersion curves were screened to exclude those with inconsistent measurement at adjacent periods.

For each frequency, a ray theory based tomography method [Barmin et al., 2001] was applied to dispersion measurements with signal-to-noise (SNR) ratio greater than 5 to produce Rayleigh group and phase velocity map on a grid of nodes spaced at 0.2° (Figure 5.1b). Only station pairs with distance larger than twice of the wave-length were kept. The maps were constructed iteratively while discarding outliers. We started from larger damping factors and
excluded dispersion curves with misfit larger than 2 times of the standard deviation \( (2\sigma) \) from the mean misfit, and then redid tomography with smaller damping factors. After deleting all outlying measurements and obtaining the Rayleigh wave phase velocity maps, we extracted the average dispersion curve between station pairs and took this as the reference to redo the dispersion curve measurements. This procedure helped to minimize the \( 2\pi \) ambiguity of phase velocity measurements [Bensen et al., 2007]. We applied the tomographic inversions to the new dispersion curve measurements to update phase velocity maps and then repeat the iterative procedure described above one more time.

Although an independent surface wave study has already been constructed with the data collected by the 2012-2013 deployment [e.g. Cai et al., 2018], we re-processed the data following similar procedure described above except using the reference dispersion curves extracted directly from the phase velocity inversion results in Cai et al. [2018]. As the \( 2\pi \) ambiguity has already been addressed in Cai et al. [2018], we didn’t update the phase dispersion curve measurements iteratively.

Phase and group dispersion curve measurements from these two deployments were then combined, and tomographic inversions were performed to get both group and phase velocity maps on a grid of nodes spaced at 0.2° (Figure 5.1b). Outlying measurements were discarded iteratively as described earlier. Based on the L-curve, a damping factor of 200 was chosen for the final tomography inversion to get a balance between resolution and variance (Figure A5.2). The tomographic inversion returns both isotropic and azimuthal anisotropic components of the Rayleigh wave group and phase velocity.

5.3.3 Teleseismic Rayleigh Wave Phase Velocities from Two-plane-Wave
Tomography (TPWT)

Pyle et al. [2010] investigated the teleseismic Rayleigh wave phase velocities using data from the 2003-2004 deployment. In this study, we performed the tomography inversion with the combined dataset from the two independent deployments to improve the coverage for the Mariana forearc and volcanic arc. Data from the two deployments were first processed separately and then combined. We utilized the two-plane-wave tomography method [Forsyth and Li, 2005] with a 2-D Fréchet kernels [Yang and Forsyth, 2006a] to invert for Rayleigh wave phase velocity at each period for both isotropic and anisotropic components. Unlike traditional Rayleigh wave tomography based on ray theory, this method considers scattering and multipathing effects of Rayleigh waves outside the study region by simplifying the scattered incoming wave as the sum of two interfering plane waves [Forsyth and Li, 2005]. Additionally, by using 2D Fréchet kernels, scattering and multipathing effects within the study region can be also approximately addressed [Yang and Forsyth, 2006a]. This method also enables us to combine two datasets collected over different time periods to expand the study region.

For the 2003-2004 deployment, we followed a slightly different data selection criteria and data processing procedure compared with [Pyle et al., 2010]. Due to the close spacing of land stations and the better long period performance of the STS-2 sensors, we used only land stations with the STS-2 sensors for the teleseismic phase velocity inversion. We selected seismograms from 154 earthquakes during the deployment period with surface-wave magnitude (Ms) larger than 4.5 and epicentral distance between 30° and 150°, on the basis of the International Seismological Center (ISC) catalogue (Figure 5.1c). The raw seismogram of each event was cut from the origin time of the earthquake to 12000 s after. Prior to any further analysis, the vertical
seismograms were down-sampled to 1 sps and corrected for instrument responses. Noise in seismograms at long periods (> 50s) due to ocean swell and associated water pressure variations, as well as tilt caused by local currents, were removed by correcting the vertical channel using horizontal and pressure channels when data was available [Bell et al., 2015; Crawford and Webb, 2000; Webb and Crawford, 1999]. The seismograms were then modified to have uniform polarity for all instrument types. For each period of interest from 22 to 78 s, we used a 10 mHz narrow bandpass filter (fourth-order Butterworth, zeros-phase shift) centered at the interested frequency to filter the seismograms. The filtered data were then windowed manually to isolate the fundamental mode of the Rayleigh wave and records with poor SNR were discarded.

We noticed significant amplitude differences between records from different instrument types, which could not be purely caused by structure variations. We thus obtained a preliminary amplitude correction for each instrument type by comparing long period waveforms (> 40 s) from adjacent stations before any further analysis. For the 16 bit OBSs, the correction is 2.287, for the 24 bit OBSs, the correction is 7.858, and for the Japanese OBSs, the correction is 0.181. The land seismic stations, which had STS-2 sensors, all had amplitude corrections near 1.0 and served as a reference for the other instruments. Only a constant correction was applied because the amplitude difference was consistent over different frequency bands. As we used more recently revised instrument responses, we didn’t observe any systematic phase discrepancy between data recorded by Japanese OBSs and data recorded by adjacent U.S. stations [e.g. Pyle et al., 2010]. To further exclude records with high noise or with unduly complicated waveforms, we performed TPWT on a grid of nodes spaced at 0.5° starting from a 1-D average model from [Pyle et al., 2010]. We discarded records showing large residuals and inconsistent waveforms. We repeated this procedure several times until all remaining waveforms are consistent and of
For the 2012-2013 deployment, we followed similar data processing procedure as described above [Cai et al., 2018]. We selected seismograms from 263 earthquakes with surface-wave magnitude (Ms) larger than 4.5 and epicentral distance between 30° and 150° on the basis of the International Seismological Center (ISC) catalogue (Figure 5.1d). A constant amplitude correction was also applied to each instrument type [Cai et al., 2018]. Only waveforms of high SNR and showing consistent singles were kept.

Good fundamental-mode Rayleigh wave records for each period of interest from the two deployments were then combined, and TPWT is performed over an extended region (Figure 5.1b). We divided the study region into three sub-regions according to tectonic settings: the incoming Pacific plate, the Mariana fore-arc, and the Mariana trough (Figure 5.2a). In the first step, we determined the average phase velocity for each period for the entire study region and for each sub-region. 2D Fréchet kernels were calculated with the average phase velocity for the entire region. In the second step, we used the average phase velocity for each sub-region as the starting model to invert the 2D phase velocity map, with node spacing of 0.4°. A priori standard deviation of 0.25 km/s and a smoothing length of 80 km were used. This inversion also solved for station amplification corrections for each station to account for any discrepancy in the nominal instrument responses as well as site effect [Yang and Forsyth, 2006b]. In the third step, the newly obtained frequency-independent station amplification corrections were applied to all records. We then conducted a series of TPWT inversions with progressively finer grids (from 1.2° down to 0.8° and then 0.4°), using the previously inverted phase velocities as the a priori model. This scenario can largely eliminate the dependence on the starting model and the a priori
information [Rau and Forsyth, 2011; Wei et al., 2015]. For the larger node spacing of 1.2° and 0.8°, we used a smaller a priori standard deviation (0.05 km/s) and a larger smooth length (150 km and 100 km), giving results more damped to the starting model. For the finest grid of nodes with spacing of 0.4°, we allowed more variability (a priori standard deviation as 0.15 km/s and smoothing length as 60 km). The tomographic inversion returns both azimuthal isotropic phase velocity and the averaged azimuthal anisotropic component for each sub-region (Figure 5.2).

5.3.4 Combining Phase Velocity from ANT and TPWT

We combined the ANT and TPWT results to provide more complete measurements of phase velocity for the SV-velocity inversion (Figure 5.1b). As these two methods are based on different principles and assumptions, the results of phase velocity at common periods from two methods are not always exactly the same [Wei et al., 2016; Yao et al., 2006] (Figure 5.3 and Figure 5.4). We designed a region for the joint ANT-TPWT inversion so that all nodes within it have well-resolved phase velocities at more than 10 periods from the ANT (blue contour in Figure 5.1b). Phase velocities were interpolated onto a uniform grid of nodes spacing at 0.2° before being combined at each node (Figure 5.1b). In the outer regions where no good ANT results were available, we used phase velocities from the TPWT at all periods to invert SV-velocity, so the shallower structures have lower resolution in these areas.

For phase velocities from ANT (6-30 s), the uncertainties were normalized at each period so that the best-resolved node has uncertainty equal to a reference value given below. According to previous experience [Cai et al., 2018; Shen et al., 2013; Wei et al., 2016], the reference uncertainty is assigned as 0.05 km/s for period of 6-12 s, 0.04 km/s for periods of 13-23 s, and 0.07 km/s for periods of 24-30 s. The uncertainty of the ANT group velocity (6-24 s) were
assigned to be 2.8 times of the uncertainty of the ANT phase velocity at the same period, as suggested by [Shen et al., 2016]. For phase velocity from TPWT (22-78 s), the uncertainties were normalized in the same way. The reference uncertainty is assigned as 0.07 km/s for periods of 22-30 s, 0.04 km/s for periods of 33-50 s, and 0.05 for other periods. We utilized a linear weighting average method to combine phase velocities measurements and uncertainty estimates for overlapping periods (22-30 s). A running average filter was then applied to make the resulting dispersion curve smoother. Group velocity results from ANT (6-24 s) were also included for SV-velocity inversion to better constrain water thickness and to improve resolution for shallower structure.

5.3.5 SV-Wave-Velocity Inversion with a Bayesian Monte-Carlo Algorithm

Previous studies [Wei et al., 2015] suggest the linearized inversion of SV-wave velocity from phase/group velocity dispersion curves [Herrmann, 2013] depends on the starting model. In order to avoid the potential bias of the starting model, to better apply a priori constraints on crustal thickness and other parameters in a systematic way, and to derive formal estimates of velocity uncertainty, we used a Bayesian Monte-Carlo algorithm [Shen et al., 2013] to invert the azimuthal averaged SV-wave velocity at each node.

The Bayesian Monte-Carlo method constructs a priori distribution of SV velocity models at each node, defined by perturbations relative to the starting model and model constraints. Each model consists four layers on top of a half-space: (1) water with starting thickness from smoothed bathymetry [Amante and Eakins, 2009] and an allowed perturbation of ±1.5 km, (2) sediments, (3) crust, and (4) upper mantle from the Moho to 180 km depth. The sedimentary layer is described by two parameters: a layer thickness of 0.5 km with an allowed perturbation of 0.5 km,
± 0.5 km and a constant $V_{SV}$ of 2.0 km/s with a perturbation of ± 1.0 km/s. The crust is assumed to have linearly increasing velocity with depth, and is described by three parameters: a layer thickness and $V_{SV}$ at the top and bottom of the layer. We obtained an empirical relationship between water thickness and crust thickness (Figure A5.3) based on a previous seismic refraction survey [Takahashi et al., 2008], and applied this relationship to the entire study region to get a starting crustal thickness for each node. The crustal thickness was allowed to vary by ±3.5 km from the starting value. The top and bottom crustal $V_{SV}$ are set at 3.0 and 3.2 km/s, respectively, with a perturbation of ± 1.0 km/s. The upper mantle $V_{SV}$ is parameterized by a B-spline, which is defined by 7 nodes with a perturbation of ± 30% for first 5 nodes and 20% for the last 2 nodes. We imposed the constraint that the jumps in $V_{SV}$ from the sediment to the crust and from the crust to the mantle are positive. We also applied a physical dispersion correction with a reference period of 1 s [Kanamori and Anderson, 1977] using a 1-D Q model simplified from a seismic attenuation study in the same region [Pozgay et al., 2009]. Compared to the PREM, our 1-D Q model has a high-attenuation layer in the uppermost mantle: $Q_s=60$ from moho to 100 km depth.

For each grid node, the best fitting model is identified and models are accepted if their $\chi^2$ misfit is less than 50% higher than that of the best fitting model [Shen et al., 2016; 2013]. Following Shen et al. [2016], we also exclude models with mantle velocity higher than 4.9 km/s. The posterior distribution thus provides statistical information of all possible SV-velocity models that satisfied the Rayleigh wave dispersion curves within tolerances depending on data uncertainties. An average model is then calculated from all accepted models and used for plotting and interpretations [Shen et al., 2013; Wei et al., 2016]. Examples of the SV-velocity inversion at three representative nodes are shown in Figure 5.4. The results of the Bayesian Monte-Carlo inversion fit the measured group and phase velocity dispersion curves well.
The application of Bayesian Monte-Carlo algorithm helped to avoid the potential bias of the starting models and provided better constrains on crustal thickness and other parameters. However, it used a B-spline method to parameterize the upper mantle $V_{SV}$, thus may smooth a thin layer of velocity anomalies over a wider depth range. As comparison, we also inverted the azimuthal averaged SV-wave velocity at each node using a linearized method [Herrmann, 2013], with starting models based on the Monte-Carlo inversion results. The starting model includes a water layer with thickness from the Monte-Carlo inversion, a two-layered crust if the crust is thinner than 10 km, otherwise a three-layered crust is used, and mantle layers down to 180 km, with increasing thickness from 2.6 km to 8.3 km. A differential smoothing parameter is set.

5.4 Results

5.4.1 Group/Phase Velocity

2-D maps of azimuthally averaged isotropic phase velocity for representative periods from ANT and TPWT are shown in Figure 5.3. Since ANT inverts Green’s functions for propagation between station pairs, the well resolved region is confined to a smaller area. TPWT provides results encompassing larger areas, because it has some limited resolution outside the array. Results for the incoming plate and the Mariana forearc are similar to results from [Cai et al., 2018]. At the shortest period (8 s), phase velocities are more sensitive to the overriding water layer and crustal structures, so fast speeds indicate a thinner water layer (the volcanic arc). At 20 s, slower velocities are observed along the volcanic arc compared to the backarc basin, suggesting a thicker crust or slower uppermost mantle. Extremely slow velocities are observed
beneath the Mariana outer fore-arc (Figure 5.3d), suggesting the uppermost mantle in this region has significant shear velocity reductions. In addition, the slow velocity zone tends to have a very sharp western boundary, corresponding well with the westward extent of major serpentine seamounts and disrupted forearc terrain. Phase velocity along the backarc spreading center shows large variation. The phase velocity of the southern portion of the spreading center (south of 16.5°N) and its adjacent region is slower than that of the central and northern portion (north of 16.5°N). At longer periods, lateral variations are observed both along the backarc spreading center and in the forearc. The central portion of the spreading center (~18°N) is always of somewhat higher phase velocity, consistent with the observations in (Pyle et al., 2010). The forearc portion with deeper bathymetry (south of ~16°N) is imaged to be faster than the northern region between Big Blue and Celestial serpentine seamounts (Figure 5.3e and f), with phase velocity similar to that in the incoming plate. The modern volcanic arc is characterized by slow velocities, similar to the southern backarc spreading center.

The Rayleigh wave group velocity maps presented in Figure 5.3a and 5.3b are derived exclusively from ANT. They are similar to the phase velocity maps except group velocity at a given period is sensitive to shallower structure than phase velocity. At 8 s, high group velocities correspond well with shallow water locations such as the volcanic arc and serpentine seamounts, implying our results have relative good horizontal resolution. At 20 s, the group velocity map also shows significant variations along the spreading center (Figure 5.3b). The relatively stronger heterogeneities in the group velocity maps are accommodated by assigning larger uncertainties to the group velocity dispersion curves (Figure 5.4).

5.4.2 Azimuthally Averaged SV-Velocity
The azimuthally averaged SV-velocity structure of the Mariana forearc, arc and backarc obtained from the average model of the Monte-Carlo inversion at each node is shown in Figure 5.5 and Figure 5.6. The model fits to the data are good, and the square root of the reduced $\chi^2$ misfit of the average model for most nodes is less than 1 (Figure A5.4), suggesting the assigned uncertainties to phase and group velocity measurements are appropriate. Without constraints from converted seismic waves (e.g. receiver functions), it’s difficult to resolve the Moho depth precisely. However, the resulting crustal thickness changes little compared to the starting models (Figure A5.3), and the changes are much smaller than the allowed perturbations, suggesting the starting models fit the data reasonably well.

The addition of the high resolution ANT phase velocity maps at short periods improves the resolution for shallower depth, and small scale features are well resolved for shallow depths within the region of the ANT-TPWT joint inversion. But this resolution contract also introduces some artificial discontinuous features at the boundary of the joint inversion region. For instance, low velocity anomalies are imaged beneath the outer forearc at 20 km and 30 km depth, but stop in an artificial way at the northern and southern boundary of the joint-inversion region (purple contour). We thus avoid interpreting shallower structure close to the boundary and outside of the joint-inversion region.

The most prominent feature in the forearc is the extremely slow velocity anomaly (~ 3.6 km/s) beneath the outer forearc at 20 km depth. This slow velocity anomaly persists to at least 50 km depth (~4.2 km/s), with a smaller velocity reduction. Below 60 km, the forearc is mainly characterized by high velocities associated with the subducting slab. This feature is consistent with previous studies [Cai et al., 2018; Pyle et al., 2010]. Discrete slow velocity anomalies (4.1-
4.2 km/s) are observed beneath the volcanic arc starting from 20 km, and move westward toward the spreading center as depth increases. Between 16.5°N and 17.5°N, a low velocity anomaly (~4.2 km/s) is imaged beneath the backarc spreading center down to about 25 km. A high velocity layer emerges below the low velocity anomaly and covers a larger area northward to about 19°N. This high velocity layer is imaged down to 60 km, and a low velocity anomaly is observed again below it. South of 16.5°N and north of 19°N, the spreading center is characterized by a low velocity anomaly starting from 40 km. The area between the West Mariana Ridge and the spreading center is dominated by high velocity anomaly from the Moho down to ~40 km.

To better illustrate the shear velocity results, we construct a series of cross sections both perpendicular and along the spreading axis (Figure 5.6). A thick low velocity zone (LVZ) is consistently imaged in the upper mantle at depths of about 50 to 100 km extending from the West Mariana Ridge to the volcanic arc, and is always inclined upwards towards the volcanic arc beneath the eastern portion of the backarc basin. A thin layer of slow velocity (~4.2 km/s) from the Moho to about 25 km depth is also observed beneath and west of the spreading center. This shallow anomaly is separated from the deeper LVZ by relatively fast velocities extending westward from just west of the spreading center. The slow velocity anomaly at the top of the uppermost mantle is also observed at some places beneath or slightly west of the volcanic arc. (Figure 5.6). The cross section DD’ along the spreading center (Figure 5.6) shows the LVZ is at about 40-110 km depth beneath the southern portion of the spreading center, and deepens to 70-130 km beneath the middle portion of the spreading center, and then becomes shallower again beneath the northern portion of the spreading center. The LVZ beneath the southern portion of the spreading center shows most significant velocity reduction. The shallower thin low velocity
layer is most prominent beneath the middle portion of the spreading center, where the LVZ is deeper and of smaller velocity reduction (Figure 5.6). Using the TPWT method exclusively, [Pyle et al., 2010] found a inclined low shear velocity zone between the volcanic arc and the spreading center along the 18°N, which reaches about 90 km beneath the spreading center. This result is relatively similar to our results, except that our results have better resolution for shallow structure because of the inclusion of ANT phase velocities.

The azimuthally averaged SV-velocity structure obtained from the linearized method is also shown in Figure A5.5 for comparison. It images comparable features as the Monte-Carlo inversion results, except the amplitude of velocity reduction is larger. A thick LVZ is also consistently imaged in the upper mantle between 50-100 km showing similar geometry as the Monte-Carlo results. The shallow low velocity zone beneath the spreading center is more prominent, but is still separated from the deeper LVZ by relatively fast velocities (Figure A5.5).

### 5.5 Discussion

Here we discuss the implications of the new model for the structure and geodynamic processes of the Mariana arc and backarc. The forearc model is similar to that derived in Cai et al. [2018] so we will not discuss that part of the model here. We also compare the new model to previous body wave velocity and attenuation tomography [Barklage et al., 2015; Pozgay et al., 2009] and magnetotelluric [Matuno et al., 2012; 2010] studies of the region. Body wave tomography is complementary to the surface wave tomography described in this paper, in that the body wave raypaths are more vertical and provide good lateral resolution, but inferior depth
resolution. Seismic and magnetotelluric (MT) methods are sensitive to different material properties and therefore provide complementary information on the structure of back-arc basins. Seismic observations are generally more sensitive to temperature variations [Jackson and Faul, 2010] whereas MT methods are more sensitive to water content, either as a free fluid or incorporated into the crustal structure of mantle minerals [Gardés et al., 2014]. Both methods are sensitive to melt content although MT is only greatly sensitive if the melt pockets are connected [Faul et al., 2004; Pommier and Garnero, 2014; Wiens et al., 2006b].

5.5.1 Structure of the Mariana Volcanic Arc

We image two low velocity anomalies beneath the volcanic arc. A small size slow velocity anomaly is found beneath the active arc at the top of the mantle (Figure 5.6), and an inclined low velocity anomaly is observed west of the active modern arc. The inclined zone of low velocities has been previously imaged using earthquake surface wave tomography [Pyle et al., 2010] and body wave tomography [Barklage et al., 2015]. However, in this study a gap is imaged between the shallower small slow velocity anomaly and the inclined LVZ in profiles B-B’ and C-C’ (Figure 5.6). Using the Si and Mg thermobarometer developed by Lee et al. [2009], Kelley et al. [2010] found that the final equilibrium depths for Mariana arc melts is 34–87 km, which is essentially identical to the depths of the inclined LVZ.

A magnetotelluric (MT) survey across the Mariana arc at around 18°N by [Matsuno et al., 2010] images a resistive region down to 60 km depth and a conductive region beginning at 60 km depth beneath the volcanic arc. This result is consistent with our result, suggesting slab-derived water in the mantle above the slab as well as aqueous melts at the depths indicated by inclined LVZ. It also suggests the melt transport from the deep melt production region to the
crustal reservoir beneath the arc is confined to a narrow zone, so that it is invisible to images with the resolution of the surface wave or MT methods [Forsyth, 1996; Matsuno et al., 2012].

In profiles B-B’, the shallow anomaly beneath the volcanic arc shows velocity of ~4.1 km/s. This is consistent with a previous seismic refraction survey at the similar location showing sub-arc Pn velocities of 7.7 km/s [Takahashi et al., 2008]. This anomaly may be caused by downward transfer of dense crustal materials back to the mantle due to arc crustal growth [Takahashi et al., 2008; 2007]. Jull and Kelemen [2001] also suggested that the lower crustal materials might sink into the mantle because they have higher density than mantle olivine.

5.5.2 Structure of the Mariana Backarc Spreading Center

We image a thin slow velocity layer beneath the spreading center from the Moho to about 20 km depth, which is followed by a faster velocity layer from about 30-50 km and a LVZ at deeper depth. In contrast, body wave tomography finds a continuous low velocity anomaly down to 70 km beneath the backarc spreading center [Barklage et al., 2015]. Body wave attenuation tomography shows a 75 km wide columnar-shaped high-attenuation anomaly (Qp 43–60) beneath the spreading center that extends to 100 km depth [Pozgay et al., 2009]. Considering the vertical smearing of body wave tomography, this continuous feature may be an artifact. Another possible interpretation is the dynamic upwelling of melting is confined to a small area, thus can’t be imaged by the surface wave method with poorer lateral resolution. We favor the latter explanation, as other observations at slow spreading centers predict segmented melt delivery to the ridge crest [Forsyth, 1996; Lin et al., 1990].

A magnetotelluric (MT) survey across the Mariana subduction system at around 18°N by Matsuno et al. [2010] images a resistive region down to 80-100 km depth beneath the spreading
center followed by a conductive mantle, consistent with our observation of a thick higher velocity layer at similar depth range followed by a LVZ (Figure 5.6). In addition, the MT survey reveals prominent lateral variation of the resistive layer across the study area, showing a thinner resistive layer down to 50-60 km beneath the West Mariana Ridge and a thickest resistive layer right beneath the spreading center. This observation is consistent with our results regarding the thickness of the fast velocity layer, which is also greatest right beneath the spreading center and becomes thinner toward the West Mariana Ridge.

The shallower thin slow velocity layer is distributed in a narrow channel along the spreading axis, and is most prominent between 16.5°N and 17.5°N where the LVZ is deeper and of smaller velocity reduction (DD’ in Figure 5.6). This range coincides with a region of more magma supply suggested by mantle Bouguer gravity anomaly [Kitada et al., 2006]. For the region south of 16.5°N, the mantle Bouguer gravity anomaly result suggests a low magma supply, and no prominent shallow slow velocity layer is imaged in this region in our study. A reflection and refraction seismic survey across the backarc spreading center at around 17°N also shows thickening of the crust and deep reflectors near the top of the upper mantle just beneath the spreading center [Takahashi et al., 2008]. All these observations support the idea that the shallow thin low velocity layer near the top of the upper mantle representing a melting reservoir related to backarc opening. The melt transport from the deep melt production region to this reservoir may be confined to a narrow zone [Forsyth, 1996; Matsuno et al., 2012], thus is not resolved by the surface wave method. Matsuno et al. [2012] also suggests that melt transport occurs through dynamic buoyancy-driven mechanisms, in contrast to evidence from the fast spreading southern East Pacific Rise.
It is interesting to compare shear velocity results from this study with results from the Lau basin, where the spreading rate is much faster. The LVZ imaged in the Mariana backarc is of much higher shear velocity compared to the LVZ in the Lau basin [Wei et al., 2016; 2015; Wiens et al., 2006a], indicating smaller velocity reduction. This difference cannot be due to variations in mantle wedge water content, since the mantle beneath the Mariana backarc has higher average water content than the mantle beneath the Lau basin [Wiens et al., 2006a]. Instead, the mantle velocity difference is correlated well with the mantle temperature difference between these two regions. The mantle temperature in the Lau basin is ~100°C higher than that in the Mariana backarc estimated from major element [Wiens et al., 2006a]. The mantle temperature difference may result from the particular subduction and mantle flow feature in these two regions. The Lau region is characterized by extremely fast subduction [Bevis et al., 1995], slab rollback and mantle inflow from the Samoa hotspot region to the north [Smith et al., 2001; Turner and Hawkesworth, 1998; Wei et al., 2016], whereas the subduction slab in Mariana is relatively stationary. In addition, warmer backarcs may exhibit greater melt content and producing larger velocity reductions [McCarthy and Takei, 2011; Sundberg and Cooper, 2010], and this mechanism was introduced to interpret the extreme velocity anomaly in the Lau basin [Wei et al., 2016; 2015].

5.5.3 Azimuthal Anisotropy

Azimuthal seismic anisotropy is determined simultaneously with the isotropic phase velocities by both ANT and TPWT (Figure 5.2). The azimuthal anisotropy results for the incoming plate and forearc at short periods have already been discussed in Cai et al. [2018], we thus here only discuss the results from TPWT at longer periods (Figure 5.2). The absolute
magnitude of anisotropy varies depending upon the chosen parameters of the tomographic inversion such as damping coefficient and smoothing length, but the fast directions are robust. We thus confine our interpretation to the directions of anisotropy, and the relative amplitude difference between adjacent periods.

We divided the entire study region into three sub-regions, the Pacific plate, the forearc and the backarc (Figure 5.2a), and obtained the averaged azimuthal anisotropic component for each sub-region. The Pacific plate shows strong anisotropy at periods shorter than 40 s, with fast directions parallel to the trench. At periods longer than 40 s, fast directions rotate to be sub-parallel to the paleo-spreading directions (Figure 5.2e) [Oakley et al., 2008]. The fast axis orientation and stronger anisotropy at shorter periods are consistent with observations in a previous study [Cai et al., 2018], suggesting the shallower part of the Pacific plate is faulted by plate bending faults parallel to the trench.

Fast axis orientation in the forearc shows more complexity. From 25 s to 40 s, fast directions are WNW and parallel to the subduction direction of the Pacific plate. From 45 s to 60 s, fast directions rotate to be trench-parallel. At periods longer than 60 s, we find fast directions are similar to fast directions in the Pacific plate at longer periods (Figure 5.2d), both parallel to the paleo-spreading directions. The north-south orientation at longer periods is also consistent with shear wave splitting studies [Pozgay et al., 2007], suggesting that the deeper part of the forearc has north-south anisotropy.

In the backarc, fast directions are consistent over all periods longer than 22 s, which are sub-parallel to the opening direction of the Mariana trough [Deschamps and Fujiwara, 2003; Kato et al., 2003]. The magnitude of anisotropy is strongest between 33 s and 60 s, corresponding to a
most sensitive depth range from 50 km to 90 km (Figure 5.2b and c). This depth range is consistent with the deep LVZ imaged in the uppermost mantle (Figure 5.6). This result is somewhat different from previous S-wave splitting studies that show fast directions parallel to the arc or absolute plate motion direction of the Pacific plate in the backarc [Pozgay et al., 2007]. However, the S-wave splitting analysis only measures the accumulated anisotropy at all depths, thus would be unable to resolve the relatively thin layer of spreading-parallel fast directions at shallower depths.

5.6 Conclusions

We obtain a comprehensive image of the SV-wave velocity structure across the Mariana backarc spreading center, extending from the Mariana forearc to the West Mariana Ridge, by jointly inverting the Rayleigh wave phase and group velocities from ANT and TPWT with a Bayesian Monte-Carlo method [Shen et al., 2013]. Our results help constrain the melt production and transport beneath volcanic arc and back-arc spreading center. A thick deep LVZ is imaged beneath the Mariana trough with varying depth. We find a small size slow velocity anomaly beneath the volcanic arc near the top of the mantle. An inclined LVZ is observed west of the arc and connected to the deep LVZ. We also find a shallower thin slow velocity layer at the top of the mantle distributed in a narrow channel along the central part of the spreading center, where larger magma supply is suggested [Kitada et al., 2006]. The shallow slow velocity anomaly beneath the volcanic arc and the backarc spreading center may both represent a shallow melt reservoir. However, neither of them is imaged to be directly connected to the deep LVZ,
suggesting the melt may be transported through a conduit that is too narrow to be resolved by surface wave tomography [Forsyth, 1996; Matsuno et al., 2012].

Results of azimuthal anisotropy suggest the shallow part of the Pacific plate is affected by faulting associated with the bending of the plate, consistent with previous studies [Cai et al., 2018; Emry et al., 2014; Oakley et al., 2008]. In the backarc, fast directions are consistently subparallel to the opening direction of the Mariana trough [Deschamps and Fujiwara, 2003], with maximum anisotropy observed between 50 km and 90 km (Figure 5.2c), corresponding to the deep LVZ imaged in the upper mantle. In the forearc, the fast directions are parallel to the subduction direction of the Pacific plate at shallow depth, and rotate to be trench-parallel approaching the slab interface. The slab emerges at longer periods (> 60s), we find fast directions are similar to fast directions in the Pacific plate at longer periods (Figure 5.2d and e), both parallel to the paleo-spreading directions.

References


Sundberg, M., and R. F. Cooper (2010), A composite viscoelastic model for incorporating grain boundary sliding and transient diffusion creep; correlating creep and attenuation responses


Figure 5.1. Maps of the study region and earthquakes used in this study.

(a) Station distribution and the seafloor bathymetry of the Mariana subduction zone. For the 2003-2004 deployment, locations of broadband land seismic stations are plotted as blue squares, OBSs that operated throughout the observation period are plotted in red circles, and OBSs that recorded only 50 days due to a firmware issue are plotted in yellow. The OBSs from the 2012-2013 deployment are plotted as black circles. The axis of the back-arc spreading center is traced in black, and the Mariana trench axis is plotted as white dashed line. Also note the presence of Big Blue and Celestial serpentine seamounts in the fore arc. (b) Nodes for ANT and TPWT. Green squares represent the nodes for the TPWT, black dots indicate the ANT nodes, and red
dots included by the blue contour are the nodes for the ANT-TPWT joint inversion. The gray polygon outlines the region in which the TPWT inversion result achieved a reasonable resolution of phase-velocity at all periods. Yellow stars illustrate nodes used as examples shown in Figure 5.4. (c) Earthquakes (blue dots) used for the 2003-2004 deployment centered at the Mariana backarc. (d) Earthquakes (blue dots) used for the 2012-2013 deployment centered at the Mariana backarc.
Figure 5.2. Azimuthal Anisotropy determined by TPWT.

(a) The TPWT nodes are divided into three tectonic subregions: the Pacific plate (cyan dots), the forearc (magenta dots), and the backarc (green dots). The spreading center and trench are labeled as in Figure 5.1a. (b) Phase velocity sensitivity kernels at example periods including the water layer, calculated based on the average velocity model in Figure 5.4e. (c) Azimuthal anisotropy as a function of period for the backarc. Short bars through each symbol show the fast direction in map view (the arrow shows the direction of north in map view). Error bars display doubled standard deviations. (d) Anisotropy for the Forearc. (e) Anisotropy for the Pacific plate.
Figure 5.3. Maps of azimuthally averaged group and phase velocity.

(a) and (b) show group velocity at periods of 8 s and 20 s inverted by ANT. (c) and (d) show phase velocity at periods of 8 s and 20 s from ANT. (e) and (f) are phase velocity maps for periods of 25 s and 40 s inverted by TPWT. 1.5 km, 3 km and 4.5 km bathymetry contours are shown as thin grey lines. Spreading centers and trench axis are labeled as in Figure 5.1a. Big Blue and Celestial serpentine seamounts are plotted as white triangles.
Figure 5.4. Examples of Monte-Carlo inversion.

(a-c) The joint Rayleigh phase and group dispersion curves and one standard deviation error bars for three locations shown in Figure 5.1b. and the computed phase (red solid) and group (blue solid) dispersion curves from the Bayesian Monte-Carlo averaged model (d-f).
Figure 5.5. Maps of azimuthally averaged SV-velocity at 20, 30, 40, 50, 60, 100 km depths relative to the sea floor.

Purple contour encloses the region of the ANT-TPWT joint inversion. White straight lines show the cross sections in Figure 5.6. Spreading centers, trench axis, serpentine seamounts, and bathymetry contours are labeled as in Figure 5.3.
Figure 5.6. Cross sections A-A’, B-B’, C-C’ and DD’ showing the azimuthally averaged SV-velocity from Monte-Carlo inversion.

White dashed lines are the forearc Moho depth from [Takahashi et al., 2008]. Thick white lines are projected 6-km thick slab crust, constrained by results from an active source reflection survey [Oakley et al., 2008] for shallow depth (< 50 km) and data from Slab 1.0 model [Hayes et al., 2012] for greater depth (> 80km). Thin white lines are contours of 3.6 and 3.8 km/s, and thin black lines are contours of 4.1 and 4.2 km/s. Circles are relocated earthquakes in the subducting
plate within 25 km from each profile collected from different datasets. The magenta circles are from Emry et al. [2011] and white circles are from Barklage et al. [2015], both of which used the 2003-2004 data. Black circles are from the 2012-2013 OBS deployment [Eimer et al., 2018]. Main tectonic features are labeled as West Mariana Ridge (WMR), spreading center (SC), volcanic arc (VA), Peacock Seamount (PS), Turquoise Seamount (TS), and Big Blue Seamount (BBS).
Figure A5.1. Comparison between linear stacking and tf-PWS for a station pair 205 km apart. Station OBS06 only has 40 days records.

(top) Linear stacking (black) and tf-PWS (red) waveforms filtered between 10 and 30 seconds. 
(bottom left) Power spectrum of linear stacking waveform. (bottom right) Power spectrum of tf-PWS waveform.
Figure A5.2. **L-curve for ambient noise tomography.** The number beside each data point is the corresponding damping factor.
Figure A5.3. Crustal structure of the Monte-Carlo inversion.

(a) Empirical relationship between water depth and crust thickness (red line), by fitting data from [Takahashi et al., 2008] (black dots). (b) Starting crustal thickness. (c) Crustal thickness of the Monte-Carlo inversion results. (d) Changes in crustal thickness of the results compared to the starting models.
Figure A5.4. $\chi^2$ misfit for Monte-Carlo inversion.

(a) Misfit at each node. (b) Histogram of the misfit for nodes within the joint ANT-TPWT inversion region (purple contour in (a)). Average misfit is about 0.9.
Figure A5.5. Cross sections A-A’, B-B’, C-C’ and DD’ showing the azimuthally averaged SV-velocity from linear inversion. Slab crust, forearc Moho location, earthquakes and main tectonic features are labeled as Figure 5.6.