Seismicity and Shallow Structure at the Mariana Subduction Zone

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Seismicity and Shallow Structure at the Mariana Subduction Zone

by

Melody Eimer

A dissertation presented to

The Graduate School

of Washington University in

partial fulfillment of the

requirements for the degree

of Doctor of Philosophy

May 2020

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Washington University in St. Louis

May 2020
Dedicated to Bill & Hiromi Eimer
ABSTRACT OF THE DISSERTATION

Seismicity and Shallow Structure at the Mariana Subduction Zone

by

Melody Eimer

Doctor of Philosophy in Earth and Planetary Sciences

Washington University in St. Louis, 2020

Professor Douglas A. Wiens, Chair

This dissertation examines the incoming Pacific plate and the overriding forearc at the Mariana Subduction Zone using passive and active source seismology. The incoming plate, with water bound in the plate sediment, crust and mantle, is of interest to help constrain the global water cycle. The seismogenic zone at Mariana is noted for being an aseismic end-member, while extensional faults and active serpentine mud volcanoes characterize the forearc. I use an ocean bottom seismograph (OBS) deployment spanning both the incoming Pacific Plate and the forearc to study the seismicity and shallow structure. The passive source component of the deployment consists of 20 broadband OBSs, 5 hydrophones, and 7 land stations on the volcanic arc deployed for one year. The active source component used in this study consists of two ~400 km transects that span both the outer forearc and Pacific Plate, with co-located multi-channel seismic (MCS) reflection and wide-angle refraction data collected. 59 short period OBSs and hydrophones were deployed for the first month to collect the refraction data, while a 646 channel, 8-km long streamer was used to collect the reflection data.

Using all of the deployed OBSs, land stations, and hydrophones, 1,692 earthquakes were detected and located in the study area. Of these, the largest 17 were used to invert for earthquake focal mechanisms, and clusters of earthquakes were relocated using a double-difference method.
Using the wide-angle refraction and MCS reflection data, P-wave velocity and reflection profiles were constructed for the two transects.

Earthquakes in the incoming plate occur to ~35 km below the seafloor, and focal mechanisms indicate normal faulting due to the plate bending. Most of the seismicity occurs within 70 km of the trench, indicating the greatest deformation due to bending occurs near the trench axis. This is consistent with the normal faults identified in the reflection profiles, with increasing fault offset and occurrence near the trench. The P-wave velocity in the mantle also reflects this change, with increasing reduction of velocity nearing the trench. The mantle velocity reduction is indicative of hydration by serpentinization and corresponds to 23 vol% bulk serpentinization. However, considering the case in which hydration is limited to the normal faults identified in the reflection profiles, the estimate of water bound in the incoming plate mantle is greatly reduced.

Earthquakes occurring arc-ward of the trench reflect the seismogenic character of the megathrust. A thrust sequence at 10 km depth and 20 km west of the trench indicates the seismogenic zone extends to nearly the trench. The forearc is marked by a heterogeneous distribution of low magnitude seismicity, which may be attributed to incoming plate roughness and/or serpentinization. The reflection profiles highlight the variable forearc morphology. Compressional features and a steep trench slope are observed in the northern line, while the southern line hosts a serpentine mud volcano and gradually slopes towards the trench. The difference may be due to the influence of subducted topography. The serpentine mud volcano, Turquoise Seamount, is underlain by a wide and deep velocity anomaly, suggesting a wide influence by seamount building. The outermost forearc is characterized by low velocities, and may indicate tectonic erosion.
Chapter 1: Introduction

Subduction zones are an essential component to Earth’s dynamic system, and are tied to processes including arc magmatism, water circulation, and recycling of tectonic plates. Despite their importance, subduction zones are traditionally more difficult to study due to their primary occurrence offshore. The development of ocean bottom seismographs (OBS) has allowed for temporary deployments aimed at studying subduction zone structure and dynamics. This dissertation covers results from an OBS deployment in Central Mariana aimed at understanding the shallow megathrust, the outer forearc, and the subducting Pacific plate.

The Mariana subduction zone is often cited as a type example of subduction zones with backarc spreading and lack of large megathrust earthquakes (Uyeda & Kanamori, 1979). The shallow seismogenic zone is characterized by a heterogeneous distribution of low magnitude earthquakes, likely reflecting variable conditions along the megathrust (Emry et al., 2011). The forearc is under extension (Kato et al., 2003) and shows evidence of extensive hydration, with active serpentine mud volcanoes (Fryer et al., 1995) and velocity anomalies observed in seismic studies (Cai et al., 2018).

The deep water cycle accounts for water circulation below the earth’s surface and has important implications for many geologic processes. Within a subduction zone, water bound in the crust and mantle of the incoming plate plays a vital role in arc magmatism, mantle wedge rheology, and seismicity. Yet estimates of water flux at subduction zones remain uncertain and particularly hindered by a lack of knowledge of the degree of hydration in the incoming plate mantle (Hacker, 2008; Emry and Wiens, 2015). Normal faults resulting from the bending of the incoming plate are thought to provide pathways for water to penetrate the subducting slab.
(Ranero et al., 2003; Naif et al., 2015). The distribution of seismicity can provide limits to the extent of hydration (Lefeldt & Grevemeyer, 2008) and seismic velocity anomalies can show areas of altered mantle (Ivandic et al., 2008).

The 2012-2013 Mariana seismic experiment was deployed to study the outer forearc and incoming plate, and included both passive and active source components. The passive source experiment consisted of a yearlong deployment of 20 broadband OBS, 5 hydrophones tethered in the water column near the trench, and 7 temporary broadband land stations. An additional 9 stations were used from the USGS Northern Mariana Island Seismograph Network. We use the passive source data to study the local seismicity for the duration of the deployment. The active source component consisted of four transects with co-located multi-channel (MCS) reflection and wide-angle refraction surveys. 44 short period stations and 15 tethered hydrophones were deployed for the refraction survey. The MCS reflection data provides information about reflective horizons in the sediment and upper-most crust, while the refraction survey allows us to determine the P-velocity structure to the upper mantle. We use two of these seismic lines in this dissertation to look at the shallow structure of the incoming plate and overriding forearc.

In Chapter 2, we compile a catalog of earthquakes for the duration of the 2012-2013 Mariana seismic experiment. The earthquakes on the forearc characterize the seismogensis of the shallow megathrust. The earthquakes on the incoming plate relate to normal faulting which is thought to allow hydration of the incoming plate. The distribution of these earthquakes is used to interpret the depth and lateral extent of mantle hydration.

In Chapter 3, we analyze the seismic reflection and refraction data for two transects on the Pacific Plate. The reflection data highlights normal faulting on the incoming plate, while the P-velocity structure highlights anomalies related to intraplate volcanic activity and hydration.
We also attempt to quantify the amount of water in the incoming plate mantle based on the observed velocity anomaly. Depending on our assumptions of the composition and distribution of the hydrous mantle materials, we obtain different estimates of water subducted. This highlights the importance of the assumptions that go into estimating the water flux at subduction zones.

In Chapter 4, we analyze the seismic reflection and refraction data for two transects on the forearc. The forearc is heterogeneous, with vastly different morphological features between the two seismic lines. One profile crosses over Turquoise Seamount, revealing the velocity structure underneath the serpentinite mud volcano. The other profile reveals compressional features in an otherwise extensional forearc. The differences may reflect variable topography of the subducted plate influencing the morphology of the overriding plate.

References


Chapter 2: Seismicity of the Incoming Plate and Forearc Near the Mariana Trench Recorded by Ocean Bottom Seismographs

2.1 Abstract

Earthquakes near oceanic trenches are important for studying incoming plate bending and updip thrust zone seismogenesis, yet are poorly constrained using seismographs on land. We use an ocean bottom seismograph (OBS) deployment spanning both the incoming Pacific Plate and the forearc to study seismicity near the Mariana Trench. The yearlong deployment in 2012-2013 consisted of 20 broadband OBSs and 5 suspended hydrophones, with an additional 59 short period OBSs and hydrophones recording for one month. We locate 1,692 earthquakes using a nonlinear method with a 3D velocity model constructed from active source profiles and surface wave tomography results. Events occurring seaward of the trench occur to depths of ~35 km below the seafloor, and focal mechanisms of the larger events indicate normal faulting corresponding to plate bending. Significant seismicity emerges about 70 km seaward from the trench and the seismicity rate increases continuously towards the trench, indicating that the largest bending deformation occurs near the trench axis. These plate-bending earthquakes occur along faults that facilitate the hydration of the subducting plate, and the lateral and depth distribution of the earthquakes is consistent with low velocity regions imaged in previous studies. The forearc is marked by a heterogeneous distribution of low magnitude (< 5 Mw) thrust zone seismicity, possibly due to the rough incoming plate topography and/or serpentinization of the forearc. A sequence of thrust earthquakes occurs at depths ~10 km below seafloor and within 20
km of the trench axis, demonstrating that the megathrust is seismically active nearly to the trench.

2.2 Introduction

Incoming plate bend-faulting and shallow megathrust seismogenesis are key components of subduction zone processes. Precise earthquake locations and source mechanisms in the incoming plate and shallow seismogenic zone are essential for understanding these two aspects. Unfortunately, small magnitude earthquakes in the trench region are generally poorly located and studied, since the nearest land-based seismic stations are several hundred kilometers away. Detailed study of seismicity and faulting processes in these regions require the deployment of ocean bottom seismographs (OBSs). However, only a few near-trench regions have been well studied using OBSs.

The Central Mariana subduction zone subducts some of the oldest oceanic crust on the planet (Nakanishi et al., 1992; Müller et al., 1997). Thus, it has long been identified as an end-member margin to study the subduction zone processes, and identified as a type example of subduction zones with backarc spreading and an absence of large megathrust earthquakes (Uyeda & Kanamori, 1979). The old subducting Pacific plate shows slow upper mantle seismic velocity anomalies, indicating pervasive hydration by water circulation along bending faults, and the slow velocity outer forearc, as well as the presence of forearc serpentine mud volcanoes (Fryer et al., 1995), suggests extensive forearc serpentinization (Cai et al., 2018). The shallow seismogenic zone is of interest with unevenly distributed, low-magnitude events that likely reflect variable conditions along the plate interface (Emry et al., 2011). Here we use records from a yearlong
OBS experiment deployed in 2012 near the Central Mariana subduction zone to study the
incoming plate and shallow seismogenic zone.

2.2.1 Bend Faulting and Incoming Plate Hydration

Subduction zones are the main locations where water can be brought back into the deep
earth. Hydrous minerals, especially serpentine, make a dominant contribution to this water
circulation process (Ulmer and Trommsdorff, 1995; Rüpke et al. 2004; van Keken et al., 2011).

The uppermost mantle is sometimes assumed to be largely anhydrous due to the
extraction of water by melting at the mid ocean ridge (Hirth & Kohlstedt, 1996). However,
normal faults resulting from the bending of the incoming plate represent pathways for water to
penetrate deep into the subducting slab and hydrate the subducting crust and upper mantle
(Ranero et al., 2003; Naif et al., 2015). Normal faulting within the incoming plate seaward of
oceanic trenches is observed globally (Craig et al., 2014; Emry & Wiens, 2015), and the bending
can produce sub-hydrostatic or even negative pressure gradients along the faults to promote fluid
flow to depth (Faccenda et al., 2009). Serpentinization of the incoming plate mantle has been
interpreted in many subduction zones globally, with the amount of hydration dependent on
several factors including plate age (Horning et al., 2016), incoming plate fabric (Shillington et
al., 2015; Fujie et al., 2018), sedimentation (Contreras-Reyes et al., 2007), and convergence rate
(Contreras-Reyes et al., 2011).

The depth of extensional faulting is limited by the neutral plane, which separates the
compressional and tensional regimes of the bending plate (Chapple & Forsyth, 1979). Hydration
of the mantle is limited to the neutral plane on the basis that water cannot penetrate into the
compressional stress regime (Lefeldt & Grevemeyer, 2008). A global average depth to the
neutral plane is 30-40 km, though regional studies show variability (Craig et al., 2014; Emry and Wiens, 2015) and there may be a correlation between the neutral plane and the 300-350 °C isotherm (Contreyas-Reyes et al., 2011).

The Pacific Plate subducting at the Mariana Trench has the potential to store a large amount of water in the form of serpentine minerals, since its age is greater than 150 Ma (Nakanishi et al., 1992). The brittle-ductile transition should occur deeper in older plates (Watts, 2001), and the 600 °C isotherm approximating the antigorite stability field occurs at > 50 km depth (McKenzie et al., 2005). Surface wave tomography indicates hydration to at least 24 km below the Moho (30 km below the seafloor) based on shear wave velocity reduction (Cai, et al., 2018). Geodynamic models matching the bathymetry of the bending plate suggest a 25 km deep neutral plane (relative to seafloor), but are subject to large uncertainties (Emry et al., 2014; Zhou & Lin, 2018). A waveform inversion study of a small number of teleseismic normal faulting earthquakes shows earthquake centroid depths down to 17 km below the seafloor (Emry et al., 2014).

In this study, we use seismograms from an OBS array deployed across the Mariana trench to precisely determine smaller magnitude earthquake source parameters and study faulting on the incoming plate associated with plate bending. Earthquake locations and depths determined by ocean bottom seismograph recording are much more accurate than those derived from teleseismic recordings, and local seismicity may be more indicative of the average stress state of the plate (Lefeldt et al., 2012). The depths and focal mechanisms of incoming plate earthquakes provide constraints on the depth extent of normal faulting, and the lateral extent places constraints on the spatial distribution of faulting, which varies between different models (e.g. Emry et al., 2015; Zhou & Lin, 2018).
2.2.2 The Seismogenic Zone

The seismogenic characteristics of subduction zone megathrust faults vary widely, with some rupturing in large megathrust earthquakes, and others showing no great (Mw > 8) earthquakes for the duration of recorded history. It has been proposed that the size of earthquakes along the megathrust may be limited or reduced by the influence of subducting topography (Wang & Bilek, 2014; Lallemand et al., 2018), limited sedimentation (Seno, 2017; Brizzi et al., 2018; Li et al., 2018), serpentinization of the mantle wedge (Reynard, 2013; Hirauchi et al., 2010), and/or extensional stress across the forearc (Heuret et al., 2011).

Subducted topography has been shown to correlate with areas of weak coupling along the megathrust (Bassett & Watts, 2015; Lellemand et al., 2018). Although some subducting seamounts have been tied to earthquake generation by locally increasing coupling (Yang et al., 2012; Bilek et al., 2003), subducting topography likely deforms the overriding plate, creating a fracture network that may promote small earthquakes and aseismic creep (Wang & Bilek, 2011; Collot et al., 2017) and reducing coupling through serpentinization of the overriding plate (Singh et al., 2011). Large megathrust earthquakes are unlikely to occur under these conditions, with small asperities rupturing along areas of stable sliding and preventing large asperities from locking and accumulating strain (Wang & Bilek, 2014; Emry et al., 2011).

The Central Mariana Subduction Zone is considered an aseismic endmember (Uyeda & Kanamori, 1979), with no historical record of megathrust earthquakes greater than Ms 7.4 (Emry et al., 2011). The seismicity along the Mariana megathrust fault is characterized by a heterogeneous distribution, which has been attributed to subducting topography and/or partial serpentinization of the forearc (Emry et al., 2011). There is strong evidence for significant mantle wedge serpentinization, on the basis of active serpentine seamounts on the outer forearc.
(Fryer, 1996) and seismic imaging of the mantle wedge (Tibi et al., 2008; Pyle et al., 2010; Barklange et al., 2015; Cai et al., 2018), though the distribution may be heterogeneous. These characteristics make it an ideal location to investigate megathrust micro-seismicity and slip properties that may help explain the absence of larger events.

One of the possible reasons for the lack of large megathrust earthquakes in the Mariana trench is a proposed narrow zone of thrust faulting (e.g. Hyndman et al., 1997). However, Emry et al. (2011) show that megathrust faulting extends between 20-60 km depth and through a width of about 100 km. This is not an unusually narrow width, as the width of the seismogenic zone ranges from 50 to 200 km globally (Herrendörfer et al., 2015). Emry et al. (2011) did find a nearly complete absence of thrust zone earthquakes at depths shallower than 20 km and within 60 km of the trench. Many previous studies have noted similar aseismic regions near the trench, which led to a widespread assumption that the near-trench region was aseismic and deformed by creep (Byrne et al., 1988; Pacheco et al., 1993). However, observations of large slip to the trench during the 2011 Tohoku earthquake (Fujiwara et al., 2011; Lay et al., 2011; Kodaira et al., 2012) and the locations of smaller earthquakes near the trench at several locations (Todd et al., 2018) provide counter-examples. Thus the processes controlling seismic slip along the megathrust at shallow depths near the trench are poorly understood.

In many subduction zones, the updip limit of seismicity appears to correlate with the 100-150 °C isotherm (Spinelli & Saffer, 2004; Oleskevich et al., 1999), leading to the thought that it is controlled by a change in frictional properties by smectite to illite clay transformation (Hyndman et al., 1997; Vrolijk, 1990). However, laboratory experiments found that illite shale is velocity-strengthening/stable (Saffer & Marone, 2003), but the fluid release during the reaction has also been considered (Spinelli & Saffer, 2004; Lauer et al., 2017). In this case, earthquakes
occur downdip of peak fluid release, due to lithification and increased effective stress by reduction in fluid pressure (Saffer, 2017; Heise et al., 2017; Lauer et al., 2017). Updip of the ~150 °C isotherm, near lithostatic overpressure prevents seismic slip and instead promotes creep and slow slip events (Ranero et al., 2008; Saffer & Tobin, 2011; Vannucchi et al., 2012). Seismic and magnetotelluric imaging confirm that areas of high water content along the decollement are inversely correlated with seismicity (Bangs et al., 2015; Saffer, 2017).

In this study, we use data collected by an ocean bottom seismic deployment centered around the Mariana Trench to precisely locate and study earthquakes near the trench. This geometry provides better coverage of the outer forearc compared to previous studies in this region and allows better characterization of the shallow seismogenic zone, including the updip limit of seismicity and heterogeneous nature of the shallow megathrust.

2.3 Methods

2.3.1 Data Set

Seismic data for this study was collected by a seismic deployment across the central Mariana Trench between late January 2012 and February 2013 (Figure 2.1). The passive source component of the experiment consisted of 20 broadband ocean bottom seismometers (OBS) that straddled the trench, 5 hydrophones tethered in the water column near the trench, and 7 temporary broadband stations deployed on islands along the Mariana arc. The OBSs included 10 Scripps Institute of Oceanography (SIO) instruments, 9 of which included Trillium T240 sensors and 1 with a Trillium T40 sensor, and 10 standard Lamont Doherty Earth Observatory (LDEO) OBSs with modified Sercel L4C 3-component sensors. All OBSs also included a differential pressure gauge. All of the OBSs were recovered and returned good data, except one LDEO
**Figure 2.1.** a. Station distribution for the 2012-2013 deployment. Stations are broadband ocean bottom seismographs (orange inverted triangles), island arc stations (green inverted triangles), and tethered hydrophones (orange stars) used for the entire yearlong experiment. White circles are short period stations and tethered hydrophones from the active source component. Red, thick line indicates location of active source profile shown in Figure 2.10. Inset shows geographical location of the study, with plate boundaries from Bird (2003). b: Map highlighting bathymetry in the study region, with earthquakes from the ISC catalog (black circles), focal mechanisms from the GCMT catalog (Dziewonski et al., 1981; Ekström et al., 2012) for the time period of the study (red focal plots), and bend-related normal faults from Emry et al., (2014) (black focal plots). The black arrow denotes the direction of apparent plate motion (Kato et al., 2003), and magnetic lineations are marked by the white lines (Nakanishi et al., 1992). Serpentine seamounts as in Fryer (2012) are denoted by white triangles. PS: Pacman Seamount, QS: Quaker Seamount, ATS: Asút Tesoru Seamount (Big Blue), TS: Turquoise Seamount, FS: Fantangisña Seamount (Celestial). An incoming plate guyot, delCanot Guyot is labeled dCG.

instrument, which did not return any data and one SIO instrument, which returned data only from the differential pressure gauge. The horizontal components from the broadband OBSs were oriented using surface waves from teleseismic earthquakes (Scholz et al., 2017). The tethered hydrophones used LDEO data-loggers and were designed to float in the water column ~5 km below sea level. Tethered instruments were required because the instruments were not rated past 6 km depth whereas the trench reaches 8.6 km below sea level in the study area. Nine island stations from the USGS Northern Mariana Islands Seismograph Network on the islands of Saipan, Anatahan, and Sarigan were also used.

The active source component consisted of an additional 44 short period OBSs and 15 tethered hydrophones that were deployed for about a month at the beginning of the experiment (Figure 2.1). These instruments were also used to determine P and S wave arrival-times for earthquakes occurring between February 1 - March 10, 2012.
2.3.2 Earthquake Location

We used a short term average to long term average amplitude ratio method (STA/LTA) implemented in the Antelope software package (BRTT Inc, Pavlis et al., 2004) to detect and associate arrivals for potential events occurring within the volume enclosed by 146-150°E, 16-19°N, and 0-150 km depth. Arrival time picks were then manually adjusted, added, and used to locate the earthquakes with a Gauss-Newton method using the program dbgenloc (Pavlis et al., 2004). Local magnitudes were also calculated in Antelope, with the median value taken over at least 5 of the incoming plate OBS and land stations. The LDEO OBS, generally located in the forearc, were not included in the magnitude calculation as the waveforms were clipped for larger events. The calculated local magnitudes are consistent with teleseismic mb values listed for the larger events in the International Seismological Centre online bulletin (Di Giacomo & Storchak, 2016).

The events are then relocated using a probabilistic nonlinear earthquake location method that can utilize a 3-D velocity model (NonLinLoc, Lomax et al., 2000). Travel times within the velocity model are calculated using an Eikonal finite-difference scheme (Podvin & Lecomte, 1991). The parameters of the earthquake location problem (latitude, longitude, depth, and time) are described as probability density functions (Tarantola & Valette, 1982; Tarantola, 1987), and determined by non-linear sampling of the model space. Each sampled point in the model space is given a probability value based on how predicted travel times compare to the observed travel times. The maximum likelihood hypocenter is determined by finding the point with minimum misfit within the probability density function. Uncertainties are estimated as ellipsoids by calculating the covariance matrix from gridded values of the non-linear probability density function. The 95% confidence intervals are taken from the covariance matrix.
An a-priori 3D velocity model was constructed to better constrain the locations of earthquakes. The velocity model includes a water layer, derived from bathymetry data interpolated at 1 km intervals (Gardner et al., 2010). The crustal structure is interpreted from P wave velocity models obtained from active source refraction profiles (Eimer et al., 2017; Takahashi et al., 2008; Feng, 2016; Calvert et al., 2008), and the mantle velocity is taken from S-wave velocity models determined from surface wave tomography (Cai et al., 2018). The model was also modified to remove artifacts due to the parameterization of the surface wave tomography model. Several models were tested with perturbations in Vp/Vs ratio, % radial anisotropy, and crustal velocity before settling on a model that minimized the RMS residuals for P and S arrivals of the events. Rather than using a constant Vp/Vs ratio, a linear relationship between Vp and Vs (Vp = 1.37 Vs + 2.02) based on experimental results (Ji et al., 2013; Christensen, 2004; Salisbury & Christensen, 1978) was implemented to determine the S velocity for the crust and P velocity for the mantle. This equation approximates the change in Vp/Vs ratios due to composition, and is roughly consistent with regional models for Vp/Vs ratios from P and S wave tomography (Barklage et al., 2015) and expected ratios for an oceanic plate (Hyndman, 1979; Kandilarov et al., 2015). The mantle shear velocities from the surface wave tomography model were corrected for radial anisotropy by applying a 1% increase to the observed Sv velocities to obtain the Voigt average S-velocity, assuming 3% radial anisotropy as is typically found for the uppermost mantle (Montagner, 2006). Changing the velocity model caused the hypocenters to shift more than the calculated 95% confidence intervals, since the confidence intervals assume the a-prior velocity model and do not include the dependence on the input velocity model. However, the events shifted less than 5 km for more than 99% of the events for the suite of plausible velocity models tested.
To further understand the earthquake distribution, localized subsets of events were relocated using a double difference relative location method (HypoDD, Waldhauser & Ellsworth, 2000; Waldhauser, 2001). This method reduces the scatter in earthquake locations due to velocity variations by assuming the ray paths for two adjacent earthquakes are similar to a given station, and thus the travel time difference between the event pair is due to their spatial offset and not velocity structure along the ray path. A weighted least squares inversion is used to minimize residuals between observed and theoretical travel time differences for pairs of earthquakes at each station. The picked absolute arrival times for P and S arrivals were used across all available stations, and differential P wave arrival times from waveform cross-correlation were added when waveforms were similar. To satisfy the condition that earthquake separation is less than event-station distance and scale length of velocity heterogeneity (Waldhauser & Ellsworth, 2000), three clusters were defined and event pairs were required to be within 10 km of each other for absolute arrival times and within 5 km for differential travel times. An additional cluster, which included 348 earthquakes around Asút Tesoru seamount (informally known as Big Blue Seamount and the largest identified mud volcano in the forearc (Oakley et al., 2007)), used absolute travel times with event pair separation of 15 km. To avoid the influence of poorly located earthquakes, only earthquakes with an RMS travel time residual of less than 1 second were included. To best approximate the velocity structure, each cluster used a different local 1-D velocity structure based on the 3-D velocity model described above.

2.3.3 Focal Mechanism Determination

Focal mechanisms and seismic moments were determined for the larger events in the catalog using regional waveform inversion. Records of larger events from OBSs located in the forearc were clipped, so we used OBSs on the incoming plate and the land stations for regional
waveform inversion. A non-linear low frequency signal on the OBS records affected some of the earthquakes, limiting the number of events that could be studied. Greens functions were calculated using wavenumber integration (Herrmann, 2013). The source time function was estimated as a 1 second parabolic pulse, sufficient for the Mw 4-5 events being studied. Because the wavenumber integration method requires a 1D velocity structure, one of two 1D velocity profiles were used depending on where the station was located (Table A2.1). A velocity model approximating the incoming plate and including a 5.6 km water layer was used for the OBS, which were all on the incoming plate, and a model approximating the forearc and including a 2 km water layer was used for island arc stations. 3% radial anisotropy was included in the mantle.

The data were filtered to 0.03-0.06 Hz to minimize the effect of local structure and to minimize long period noise on the horizontal components, and down-sampled to 1 Hz. The seismograms were rotated to the great circle path and cut to time windows extending from prior to the P-wave arrival until after the surface wave arrival. An amplitude correction was also applied to the OBS records, because of some uncertainty in the nominal OBS gain values. The station amplitude corrections were determined from surface wave amplitudes using a two-plane-wave tomography method and a dataset of 380 earthquakes (Cai, 2018; Yang & Forsyth, 2006). All vertical, radial, and transverse components with a clear earthquake signal were included in the inversion. Focal mechanisms were solved for using a grid search to find the solution, which maximizes fit between the synthetics and data (Herrmann, 2013).

Before determining source parameters for smaller events with no prior source information, we first validated the regional waveform inversion method by determining source parameters for two events during the deployment that have centroid moment tensor (GCMT)
solutions (Dziewonski et al., 1981; Ekström et al., 2012). Comparison of the double-couple source mechanisms listed in the GCMT catalog with those determined in this study indicate our procedure and 1-D structure approximation produce reliable results. The solution and waveform fits for a shallow thrust earthquake is shown in Figure 2.2, including a comparison to the solution given by the GCMT catalog.

Due to the lower seismicity levels and magnitudes of the earthquakes on the incoming plate, the signal to noise ratios at longer periods were insufficient to determine focal mechanisms using waveform inversion. In order to determine mechanisms for these earthquakes, first motion polarities were used instead. Since first motion focal mechanisms tend to be less reliable than those determined by waveform inversion, we selected the largest incoming plate events with good azimuthal coverage to ensure dependable results. Only earthquakes with at least 10 clear P polarities were considered, with SH polarities included when possible. SV polarities were not included because of the phase distortion expected for post-critical incidence angles (Snoke, 2003). As the selected earthquakes were all on the interior of the array, azimuthal coverage was acceptable for events that exceeded the minimum required number of polarities. Take off angles and station azimuths were calculated using the 3D velocity model, and a grid search over strike, dip and rake angles was used to find acceptable solutions based on the input polarities (Snoke, 2003). Final mechanisms were taken as the average of the acceptable solutions since the suite of possible solutions for each earthquake were of similar strike/dip/rake. Comparison of the first motion polarity results with two of the GCMT earthquake solutions suggests that focal mechanisms determined with this method are reliable (Figures A2.1).
GCMT
Mw 4.9
Depth 14.6 km

This study
Mw 4.9
Depth 11 km
Figure 2.2. Results from the regional focal mechanism inversion for the May 20, 2012 earthquake at 17.0 °N and 147.6 °E. All waveforms used in the inversion are shown, with data in black and synthetics in red. The focal mechanism determined by this study is in black, and the GCMT catalog solution is in grey (Dziewonski et al., 1981; Ekström et al., 2012). The P axis is marked with a filled triangle, and T axis with a filled circle for each solution.

2.4 Results

We located 1692 earthquakes within the study area using the 3-D velocity model (Figure 2.3). Of the focal mechanisms in this study, 12 were determined using regional waveform inversion in the forearc (Table 2.1), and 5 were determined using first motion polarities in the incoming plate (Table 2.2). The magnitude of completeness is about ML ~ 3 for earthquakes that met the criteria for determining magnitudes. A b-value of 1.3 is calculated by linear regression between ML 3.3 – 5.5 from the frequency-magnitude distribution (Figure A2.2).

Earthquakes in the incoming plate prior to subduction, 389 in total, are associated with the bending of the plate. While earthquakes occur at distances up to 167 km from the trench, most of the incoming plate seismicity occurs within 70 km of the axis (Figure 2.4). Earthquake frequency increases continuously towards the trench, with over half the dataset located on the inner trench slope within 30 km of the trench axis. Earthquakes on the incoming plate extend down to about 35 km below the seafloor, with a peak at 25 km depth (Figure 2.5). Although earthquake frequency increases towards the trench, the depth extent appears to be relatively constant with distance from the trench (Figure 2.6). Focal mechanisms using first motion polarities were determined for 5 earthquakes on the incoming plate. All 5 events show extensional focal mechanisms and are located at depths of 26 - 31 km below the seafloor (Table 2.2). Most are oriented as expected, with strike sub-parallel to the trench. One event shows east-
Figure 2.3. Map of seismicity from this study, color-coded by depth below the seafloor. Focal mechanisms determined by regional waveform inversion are in black and by first motion polarities are in red. Sizes are scaled to magnitude. Grey inverted triangles show seismometers used for the yearlong study.
Table 2.1. Focal mechanism solutions determined by regional waveform inversion. Depth is given in reference to sea level.

<table>
<thead>
<tr>
<th>Date</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Depth(km)</th>
<th>Strike</th>
<th>Dip</th>
<th>Rake</th>
<th>M&lt;sub&gt;W&lt;/sub&gt;</th>
<th>M&lt;sub&gt;L&lt;/sub&gt;</th>
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<td>4.4</td>
<td>4.2</td>
</tr>
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<td>2012/09/26</td>
<td>147.70</td>
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<td>4.0</td>
</tr>
<tr>
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<tr>
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<td>250</td>
<td>40</td>
<td>-150</td>
<td>4.2</td>
<td>3.8</td>
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</table>

Table 2.2. Focal mechanism solutions determined by first motion polarities. Depth is given relative to the seafloor.

<table>
<thead>
<tr>
<th>Date</th>
<th>Longitude</th>
<th>Latitude</th>
<th>Depth(km)</th>
<th>Strike</th>
<th>Dip</th>
<th>Rake</th>
<th>M&lt;sub&gt;L&lt;/sub&gt;</th>
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<td>42</td>
<td>-53</td>
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</tr>
</tbody>
</table>
Figure 2.4. a. Mantle S velocity as a function of distance from the trench at 20 km depth below seafloor, averaged over 5 profiles along the arc, perpendicular to the trench (Cai et al., 2018) b. Bathymetry with distance from the trench, averaged over the same 5 profiles used in panel a. c. Histogram of the number of events on the incoming plate as a function of distance from the trench. Blue bars include all earthquakes, orange bars show results with the two swarms on the incoming plate removed.
Figure 2.5. a. Histogram showing the depth below seafloor for events on the incoming plate. Blue bars include all earthquakes, orange bars show results with the two swarms on the incoming plate removed. Grey line indicates range of Moho depths. b. S-wave velocity profile for the trench high from surface wave tomography (Cai et al., 2018).

Figure 2.6. Plot of all earthquakes in the study, as in Figure 2.3, in cross section. Error ellipses show the 95% confidence regions for the earthquake locations.
west striking nodal planes (April 25, 2012), but the fault strike has relatively large uncertainties and a fault plane subparallel to the trench cannot be ruled out.

Two major earthquake swarms are identified on the incoming plate (Figure 2.7). Both sites show recurrent activity through the year, but have a large burst of activity over a 15-35 day interval (Figure 2.7c). Neither cluster shows a traditional mainshock-aftershock sequence, but instead show a continuous distribution of magnitudes between ML 2.5 - 4. The largest event of both clusters occurs partway through the main swarm of activity. The largest cluster, at 16.5 °N, 148 °E, is located under a large fault scarp in the bathymetry. The cloud of earthquakes is steeply dipping away from the trench and extends from 21-28 km below seafloor. There is no migration of events with time, with events occurring throughout the feature for the duration of activity. The smaller cluster, at 17.6 °N and 148.2 °E, is located just north of a major seamount in a region with several fault scarps, at depths from 17-22 km below seafloor.

The forearc seismicity consists of several prominent clusters, with the largest between 17.5 - 18.7 °N and 146.7 - 147.4 °E near Asût Tesoru (Big Blue) seamount, coincident with the largest cluster Emry et al. (2011) observed using data from the 2003-2004 OBS deployment. A smaller cluster is observed to the south at 17 °N and 146.8 °E with a relatively quiet corridor between 17.3 °N and 17.8 °N, confirming the patchy nature of the seismogenic zone in the region. This cluster also maintains a relatively constant rate of seismicity and is a dipping feature, likely delineating the plate interface. The cluster was also observed in the 2003-2004 deployment (Emry et al. 2011), indicating that both clusters are long-term features of the seismogenic zone.
Earthquakes occur throughout the year but the rate increases after the Mw 4.9 thrust event on May 20, 2012. Focal mechanisms of an earlier and three subsequent events confirm similar mechanisms, with magnitudes Mw 4.5 (May 19), 4.6 (May 26), 4.5 (May 29), and 4.1 (Sept 26). Double difference relative relocations of all the events in the cluster show earthquakes aligned in the north-south direction and deepening to the north (Figure 2.8).
Figure 2.8. Relative relocation of earthquakes at the shallow thrust sequence, color coded by depth below sea level. North-south (a.) and east-west (b.) cross sections of the relocated earthquakes, with all earthquakes from map view plotted. c. Map view of earthquakes and locations of the profiles in panels a and b.
2.5 Discussion

2.5.1 Incoming Plate Seismicity

Depth and Temporal Characteristics of Incoming Plate Earthquakes

Near the trench, focal mechanisms and locations determined in this study confirm that normal faulting occurs to depths of 31 km below the seafloor (Table 2.2), which is deeper than found for a small number of teleseismic earthquake centroids in the same area (Emry et al., 2014). The depths of all incoming plate seismicity, including events without focal mechanisms, extend to at least 35 km below the seafloor (Figure 2.5). All the earthquakes with focal mechanisms show normal faulting, and the earthquakes seem to represent a single population. If, on the other hand, the earthquake sample crossed the neutral plane and included compressional earthquakes that were too small for focal mechanism determination, an earthquake minimum and a subsequent increase at a deeper depth would be expected. Thus, it is likely that only extensional earthquakes are represented in this study. This depth is consistent with estimates of hydration from seismic tomography, which shows velocity reduction to ~30 km below seafloor (Cai et al., 2018). The correlation of earthquake depths and velocity reductions due to hydration supports the idea that bend faulting is the mechanism through which the crust and mantle are hydrated prior to subduction.

The incoming plate earthquake depths are consistent with a neutral plane about 35 km below seafloor. This is somewhat deeper than predicted by several recent flexural bending models that are constrained by the seafloor elevation profile. Zhou and Lin (2018) predict a maximum depth of normal faulting of 21 km below the seafloor, while Emry et al. (2014) predict a neutral plane at 25 km below the seafloor. However, multichannel seismic images of the outer
forearc combined with high-resolution bathymetry at the trench show that the dip angle of the incoming plate abruptly increases at the trench (Oakley et al., 2008). Emry et al. (2014) point out that including the plate interface below the outer forearc landward of the trench, rather than just the incoming plate bathymetry, predicts larger extensional stresses with a neutral plane at 38 km and brittle faulting to 30 km depth. For comparison, focal mechanisms indicate a neutral surface of about 40 km depth for the source region of the 1933 Sanriku earthquake in NE Japan (Obana et al., 2018), and the neutral surface in the incoming plate increased from 20 km depth to 40 km depth after the occurrence of the 2011 Tohoku earthquake (Obana et al., 2012). A relatively deep neutral plane may be typical for bending of old oceanic plates when the plate is not temporarily loaded by locking on the adjacent megathrust.

To further support the influence of bend faulting on plate hydration, the two clusters on the incoming plate exhibit earthquake swarm behavior, which may indicate fluid migration. The swarms are characterized by having no single principal event and a finite time of increased activity (Yamashita, 1998). Swarms observed in both intraplate settings and plate margins are thought to be generated by fluid pressure (Kurz et al., 2004). On the incoming plate off southern Chile, clusters of highly similar earthquakes have been identified and inferred to be caused by seawater infiltration to mantle depths (Tilmann et al., 2008).

Fluids are thought to cause overpressure causing earthquake rupture, which then creates permeability through which the overpressure is diffused, thus arresting further seismic activity. This limits the size of earthquakes and prevents mainshock-aftershock sequences (Yamashita, 1999). With this mechanism, the swarms indicate fluid, in this case oceanic water, along preexisting faults (Tilmann et al., 2008), and is supported by the existence of fault scarps above both locations of the swarms. Fluid discharge features observed by a Human Occupied Vehicle
(HOV) on the incoming plate in southern Mariana further support fluid cycling along these normal faults (Du et al., 2019).

**Onset and Intensity of Faulting**

Most of the seismicity is located within 70 km of the trench. The lateral extent of seismicity is consistent with where the seismic velocity starts to rapidly decrease 80 km east of the trench (Cai et al., 2018), indicating the seismicity is a good proxy for where significant hydration occurs. In addition, the seismicity rate increases with decreasing distance to the trench (Figure 2.4), with the highest seismicity rate per area occurring at the trench. If the rate of seismicity is tied to the surface faults, then increased seismicity should increase the rate of offset growth. This is generally consistent with an increase in offset (Zhou & Lin, 2018), in particular when including the large horst at the trench that is in the process of being subducted along the length of the study region (Oakley et al., 2008). In addition, a larger S-wave velocity reduction is observed closer to the trench (Cai et al., 2018), as would be expected for progressive hydration of the crust and mantle through bend faulting as the faults approach the trench. An interesting conclusion of these results is that stress and deformation in the plate are concentrated at the trench axis, instead of the outer rise as often described in the literature and in geodynamic modeling.

While the seismicity rate drops off east of 70 km from the trench, there are earthquakes occurring farther east, with three events occurring at distances larger than 130 km. Alteration of the incoming plate has been observed out to 140 km in seismic refraction profiles at the Kuril trench (Fujie et al., 2013), and has been suggested as far out as 100-500 km out on the basis of lower than average crustal velocities in the southwestern and northwestern Pacific (Grevemeyer
et al., 2018). The earthquakes observed farthest eastward from the Mariana Trench may indicate the influence of bending and the potential to alter the oceanic plate out to at least 170 km. Alternatively, these earthquakes may represent the background intraplate seismicity in the Pacific Plate.

The lateral extent of significant seismicity, although in approximate agreement with the lateral extent of fault scarps in the bathymetry, does not match regional variations observed from north to south. The onset of faulting as determined by fault scarps mapped on the seafloor appears to roughly correlate with the 6 km bathymetric contour, which varies from about 95 km from the trench axis at 18 °N to roughly 55 km at 16.5 °N (Oakley et al., 2008). Bending related seismicity occurs beyond this in the south to at least 90 km, indicating brittle deformation occurs further out than surficial expressions of faulting here. Small earthquakes likely occur on developing faults that do not yet come to the surface, or are too small to be visible in the bathymetry. In contrast, seismicity at 18 °N is largely constrained to within 30 km of the trench, while faulting in the bathymetry is observed to 95 km. This may indicate an aseismic component in fault building, or that the yearlong deployment did not capture the full extent of incoming plate seismicity.

The seismicity rate is noticeably greater on the incoming plate south of ~17.7 °N, especially at distances less than 20 km from the trench axis. This correlates with a visible change in fault scarp direction in bathymetry, with Oakley et al. (2008) identifying a change at 17.6 °N with new faults formed due to bending to the north, and coexistence of new and reactivated abyssal hill fabric to the south. Reactivation of abyssal hill faults occurs when the plate fabric strikes <25°-30° from the trench axis (Masson, 1991; Billen et al., 2007). In Alaska, alignment of incoming plate fabric at the Shumagin gap results in bend faulting and hydration
compared to the Semidi segment where the plate fabric is oblique (Shillington et al., 2015). However, in Japan, more plate hydration is observed at the Japan Trench with oblique plate fabric, compared to the Kuril trench with sub-parallel fabric. The difference is attributed to larger fault offsets at the Japan Trench compared to the reactivated faults at the Kuril trench (Fujie et al., 2018). Given the ambiguous relationship between incoming plate parameters and degree of hydration, it is unclear if the alignment in fabric would promote more hydration in the south where the seismicity rate is greater.

2.5.2 Forearc Seismicity
Heterogeneous Pattern of Seismicity

The forearc seismicity in this study is marked by a heterogeneous distribution of moderate and low magnitude events (Mw < 5), with clustering in the northern and southern parts of the study region separated by a relatively quiet gap. Relative relocation of the northern cluster near Asùt Tesoru seamount shows structure within the cluster, with small highly seismic regions located on a dipping interface (Figure 2.9). In cross section, the earthquakes appear to delineate the seismogenic zone from 20-45 km below sea level, although there are deeper events that occur within the plate and may be the updip limit of the lower plane of the double seismic zone (Shiobara et al., 2010), as also observed in Emry et al. (2011).

The sub-cluster just north of Asùt Tesoru at approximately 18°10’ N and 147°10’ E includes 3 events with focal mechanisms, the largest a Mw 4.9 on Dec 16, 2012, followed by a Mw 4.4 19.3 hours later and a Mw 4.2 three days later (Figure 2.9). The rate of seismicity also increases following the Dec 16 event, suggesting an aftershock sequence. Directly to the west at 18°15’ N and 146°50’ E, the focal mechanism of the Oct 23, 2012 event is extensional,
Figure 2.9. Relative relocations of earthquakes in the northern cluster near Asüt Tesoru (Big Blue) Seamount (labeled). Additional seamounts are denoted by white triangles. The earthquakes included in each cross section are color coded, with the horizontal lines in map view showing the boundaries for each section. Focal mechanisms for 5 larger earthquakes are also plotted.

suggesting it occurs in either the subducting slab or overriding plate. With the exception of the aftershock sequence, the earthquakes do not temporally cluster, and do not exhibit swarm-like behavior as described in Holtkamp & Brudzinski (2011). The clusters are persistent features, having been observed in previous seismic studies from ocean bottom seismic arrays located closer to the island arc (Emry et al., 2011; Shiobara et al., 2010). Larger GCMT catalog events tend to occur around the edges of the cluster (Emry et al., 2011).

Given the rough topography currently subducting in this region, topography on the downgoing plate may be responsible for the clusters. A negative correlation between large
earthquakes and rough incoming plate bathymetry has been observed (Kelleher and McCann, 1976; Bassett & Watts, 2015) and geodetic observations suggest rough bathymetry promotes creeping as the mode of subduction (Wang & Bilek, 2014). The incoming Pacific Plate at the Mariana Trench is marked by several seamounts, particularly to the north and in the south of the study region where there are loose chains of guyots. The irregularity of the trench depth and axis suggest topographic features have been subducted (Oakley et al., 2008; Fryer & Smoot, 1985), and active seamount subduction is occurring with the subduction of Dutton Ridge to the north and del Cano Guyot to the south. The irregular nature of the seamount distribution prevents identification of recently subducted seamount locations on the downgoing plate. However, past seamount subduction may have developed a fracture network downdip that would promote small earthquakes and creep while inhibiting large events (Wang & Bilek, 2014).

Globally, subducted seamounts have led to eroded frontal prisms and local seafloor uplift (Kopp, 2013). The Mariana margin does not exhibit significant forearc deformation from seamount subduction, possibly due a weak serpentinized mantle wedge and/or progressive fracturing of the incoming plate (Oakley et al., 2008), making it difficult to identify subducted features. Active seamount subduction can be observed to the north in the region east of BBS (~18.5 °N), causing a shallower trench and displaced overriding plate toe (Oakley et al., 2008). Bassett & Watts (2015) used residual bathymetry to identify the subducting anomaly continuing into the forearc, suggesting that the seamount had subducted relatively intact to depth < 20 km. The largest earthquake cluster is observed west of this feature, and seamount subduction to such downdip extent cannot be assumed given the irregular distribution of seamounts on the incoming plate. That being said, the seismicity is consistent with seamount subduction, with low magnitude events being observed as a long-term feature in this location (Collot et al., 2017).
The coincident location of seismicity clusters and forearc serpentine seamounts may further indicate the influence of subducted topography. The large cluster in the north is below and downdip of Asùt Tesoru, Quaker, and several other smaller seamounts; and the southern cluster is downdip of Turquoise and Fantangisña Seamounts. These seamounts are thought to be long-lived features, with Fantangisña estimated to be at least 10.77 Ma (Menapace et al., 2019). These serpentine seamounts are built over deep-seated extensional faults that formed due to slab rollback and increased curvature of the arc (Fryer et al., 2006; Menapace et al., 2019). Vertical tectonism caused by the subduction of plate-seamounts likely furthers the development of these faults that penetrate the forearc to allow serpentinite muds to travel to the seafloor (Oakley et al., 2007; Stern & Smoot, 1998; Fryer et al., 2000). The fact that both clusters can be associated to serpentine seamounts in the study region may lend to the idea that deformation of the overriding plate leads to conditions unable to support large megathrust earthquakes but preferentially rupture in small magnitude events.

The lack of seismicity between the two clusters could indicate two endmember scenarios: an aseismically creeping or a locked region. In Alaska, the Shumigan section of the megathrust, known to be creeping from geodetic measurements, shows a strong cluster of seismicity whereas sections that are locked are relatively aseismic (Shillington et al., 2015). This may suggest that the aseismic sections of the Mariana trench are locked and have strong earthquake potential. However, global studies have attributed decoupled subduction zones to several parameters including extensional upper plate stress (Heuret et al., 2011, 2012; Scholz & Campos, 2012), thin and heterogeneous trench fill (Heuret et al., 2011, Seno, 2017; Brizzi et al., 2018; Li et al., 2018), a serpentinized shallow mantle wedge (Reynard, 2013), and/or rough incoming plate topography (Kelleher and McCann, 1976; Wang & Bilek, 2011; Bassett & Watts, 2015). All of these
parameters are favorable in Mariana, and it is possible that the seismic gap observed in the center of the study area is aseismically creeping. However, without geodetic measurements, it is difficult to confirm that the seismogenic zone, especially in the quiescent region, is aseismically slipping rather than locked.

**Shallow Thrust Sequence and Updip Limit of Seismicity**

This study observed a shallow thrust sequence at 16.9 °N and 147.7 °E, located at just 20 km west of the trench (Figure 2.8). The plate interface at this location is expected to be about 5 km below the seafloor (11 km below sea level) from MCS reflection profiles (Oakley et al., 2008), which is consistent with depths from waveform inversion. The relative relocation results, however, locate the earthquakes at 6-14 km below the seafloor, or a few kilometers deeper than the expected slab surface. Choosing different starting locations for the earthquakes and utilizing different velocity structures in the relative relocation suggests the internal structure of the cluster is robust but the depth is dependent on the starting location and velocity structure. Given that all five events with focal mechanisms show thrust faulting, as expected for earthquakes along the megathrust, and the uncertainty in absolute location, the earthquakes are assumed to be on the plate interface.

Relative relocation shows that the events are aligned north-south, parallel to the fault strike from the focal mechanisms, with the shallowest earthquakes in the center and events deepening to the north (Figure 2.8). Thrust focal mechanisms are consistent with the sequence occurring along the seismogenic zone, so the variation in event depth reflects along-strike variation in the megathrust interface. The slab surface interpreted from previous reflection profiles show the plate is not a planar surface, but has along-strike undulations - the plate is
shallowest at del Cano Guyot, south of the shallow thrust sequence at 16 °N, due to the influence of the subducting seamount structure (Oakley et al., 2008). While the plate surface under this thrust sequence was not surveyed in the Oakley study, a general deepening of earthquakes north of the thrust sequence supports the plate deepening to the north.

Previously, the seismogenic zone was established to be at least 100 km wide, with the updip and downdip limits at 20 and 60 km, respectively (Emry et al., 2011). The thrust sequence in this study shows that the seismogenic zone is active shallower than the previous estimate, increasing the width of the seismogenic zone. While the previous limit coincided with the 150 °C isotherm that has been proposed to be the control on the updip limit (Spinelli & Saffer, 2004; Oleskevich et al., 1999), the new results show that the megathrust can rupture at colder temperatures in this location. This increases the total width of the seismically active thrust fault to 140 km by expanding the updip limit from Emry et al. (2011), further reinforcing the conclusion that the lack of large earthquakes in the Mariana subduction zone is not due to a narrow seismogenic zone (Emry et al., 2011).

Previous studies estimating the temperature along the Mariana megathrust fault, based on pore water chemistry from the serpentine mud volcanoes, suggest the 80 °C isotherm occurs at 15 km depth, with temperatures reaching 150 °C at slab depths of 17-24 km (Hulme et al., 2010). If temperature controlled, diagenetic and low-grade metamorphic reactions occurring at temperatures as low as ~60 °C have the potential to create conditions that support stick-slip behavior (Marcaillou et al., 2008; Moore & Saffer, 2001; Saffer & Tobin, 2011). Previous studies estimating the temperature along the Mariana megathrust fault, based on pore water chemistry from the serpentine mud volcanoes, suggest the 80 °C isotherm occurs at 15 km depth, with temperatures reaching 150 °C at slab depths of 17-24 km (Hulme et al., 2010). If
temperature controlled, diagenetic and low-grade metamorphic reactions occurring at
temperatures as low as ~60 °C have the potential to create conditions that support stick-slip
behavior (Marcaillou et al., 2008; Moore & Saffer, 2001; Saffer & Tobin, 2011). However,
recent estimates of temperature based on oxygen isotope thermometry suggest temperatures as
high as 180 °C under Yinazao Seamount at 13 km depth to slab (Debret et al., 2019), which is
consistent with the hypothesis that the updip limit is controlled by the 100-150 °C isotherm.

Alternatively, the shallow Mariana subduction zone may be unable to maintain high
overpressure, instead dewatering through the forearc. The topography above the thrust sequence
is of interest, with a local high above the shallowest part of the sequence (Figure 2.8). This may
indicate deformation of the overriding plate caused by increased coupling along the megathrust,
perhaps caused by the uneven plate surface and subducted topography. Structure on the plate
may allow for locking of the plate locally by enhanced drainage and higher effective stress
(Bilek et al., 2003; Tréhu et al., 2012; Saffer, 2017).

Structural control on the location of the shallowest thrust events is also indicated by
seismic structure obtained from a coincident active source survey. We compare the location of
the shallow thrust sequence with a co-located 2D P-wave velocity profile collected during the
active source component of the 2012-2013 seismic experiment (Eimer et al., 2017). Despite
significant depth uncertainties in the relocations, we assume that the shallow thrust earthquakes
are located along the megathrust, as indicated by their shallow thrust focal mechanisms and the
shallower depths indicated by the waveform inversion. The lateral location of the seismicity
cluster correlates with an increase in seismic velocity in the forearc above the plate interface
(Figure 2.10). The increase in seismic velocity may provide additional evidence for dewatering
and/or existence of more competent material that is allowing for locking of the interface at shallow depth.

![P-velocity profile inverted from wide-angle refraction data (Eimer et al., 2017) along a profile that is co-located with the shallow thrust sequence (Figure 2.1a). Dark shaded areas indicate model space that is not sampled by data in the tomographic inversion. The black bar shows the lateral position and extent of the shallow thrust sequence, which is assumed to lie along the megathrust due to the thrust faulting mechanisms and the waveform inversion depths.](image)

**Figure 2.10.** P-velocity profile inverted from wide-angle refraction data (Eimer et al., 2017) along a profile that is co-located with the shallow thrust sequence (Figure 2.1a). Dark shaded areas indicate model space that is not sampled by data in the tomographic inversion. The black bar shows the lateral position and extent of the shallow thrust sequence, which is assumed to lie along the megathrust due to the thrust faulting mechanisms and the waveform inversion depths.

### 2.6 Conclusions

1. The incoming plate seismicity indicates a neutral plane of about 35 km depth, which may be the controlling factor for mantle hydration of the incoming plate. The distribution of seismicity is consistent with the depth and lateral extent of the velocity reduction observed from seismic tomography (Cai et al., 2018), supporting bend faulting as the mechanism by which the plate is hydrated.

2. Although earthquakes are observed out to 167 km from the trench, significant seismicity begins about 70 km from the trench and the rate increases continuously towards the trench. This indicates that the largest bending deformation occurs at the trench axis, rather than along the outer rise as sometimes found in modeling studies.
3. The heterogeneous distribution of seismicity may reflect the incoming plate roughness and related forearc serpentinization. The rough incoming plate may also encourage the rupture of small earthquakes and discourage large megathrust earthquakes.

4. The shallow thrust sequence suggests the updip limit of seismicity can occur at temperatures less than 80 °C, and may be tied to dewatering, diagenesis and/or low-grade metamorphism.

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Figure A2.1. Results from the first motion polarity focal mechanism determination for the May 20, 2012 earthquake at 17.0 °N and 147.6 °E. All first motion polarity picks used are shown, using the convention of Snoke (2003). All possible solutions allowing for 1 error in SH polarity (marked red) are shown by thin black lines. The averaged solution is the thick black line, the solution determined by waveform inversion as in Figure 2 is in dashed blue, and the GCMT catalog solution is in grey (Dziewonski et al., 1981; Ekström et al., 2012).
Figure A2.2. Frequency-magnitude plot of all earthquakes in the study for which magnitude could be determined. Magnitude of completeness is $M_L \sim 3$, $n$ is the total number of events plotted, and blue circles indicate data used to calculate $b$-value by linear regression.
Table A2.1. Velocity structure used for instruments on the forearc and incoming plate, used in the regional waveform inversion. The structure is given in uniform layers, with the depth column denoting the depth below sea level of the bottom of the layer.

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Chapter 3: P-velocity Structure of the Incoming Plate at the Mariana Subduction Zone and Implications for Mantle Hydration

3.1 Abstract

Subduction zones are a key part of the global water cycle, subducting water that is tied up in the incoming plate to later be released underneath the forearc or recycled back into the mantle. The amount of water subducted in the incoming plate mantle is still relatively unconstrained, but is thought to reach mantle depths by traveling along normal faults on the incoming plate. We study the incoming Pacific Plate at the central Mariana subduction zone using multi-channel seismic reflection and wide-angle refraction data along two transects perpendicular to the trench axis. The reflection profiles reveal normal faulting out to 95 km from the trench. A reduced velocity in the mantle, interpreted as the result of hydration, is observed in both seismic lines, with a greater anomaly to the north. Interpreting the velocity anomaly as bulk serpentinization, we get an estimate of 23 vol% serpentinization or 3 wt% water based on the minimum mantle velocity observed of 7.3 km/s. We also consider the case in which hydration, either through serpentinization or as porous fault gouge, is limited to the normal faults identified in the reflection profiles. If hydration is limited to these faults and we consider water saturated porous damage zones rather than serpentine filled cracks, the estimate of water in the plate is reduced by over half.

3.2 Introduction

Estimates of the water flux at subduction zones remain uncertain, in particular the amount of water brought into the trench by the subducting plate. Water input is better constrained for the
incoming plate’s sediments and upper crust from samples gathered by the Ocean Drilling Project (Staudigel et al., 1996; Kerrick and Connolly, 2001; Plank and Langmuir, 1998), but is less understood for the incoming upper mantle (Rüpke et al., 2004; Hacker, 2008). Geophysical methods have provided estimates for hydration of incoming plates for a global assortment of subduction zones, finding hydration is dependent on many factors, including temperature (van Keken et al., 2011), plate age and convergence rate (Iyer et al., 2012; Rüpke et al., 2004), and orientation and style of remnant fabrics (Shillington et al. 2015, Fujie et al., 2018).

Bending of the incoming plate causes normal faulting, which provides a pathway for water to infiltrate the crust and mantle (Naif et al., 2015; Ranero et al., 2003). Bend-related sub-hydrostatic or even negative pressure gradients along these faults promote fluid flow to depth (Faccenda et al., 2009). The ability for water to permeate these faults is also dependent on the fault zone architecture, where a wide damage zone can act as a conduit but a high percentage of fault core material may act as a barrier (Caine et al., 1996).

Once the water is in the plate, it reacts with the surrounding rock to form various hydrous phases. In the crust, alteration products derived from oceanic crust include amphibole, lawsonite, clays, chlorite, and talc (Carlson & Miller, 2003; Morris & Ryan, 2003). Samples of gabbro with typical lower crustal velocities have been found to have 5-15% alteration products, including amphibole and phyllosilicates (Carlson & Miller, 2003). Increase in porosity and further alteration is likely at the bend region of the plate by normal faulting (Naif et al., 2015; Grevemeyer et al., 2018).

A favored mechanism for water storage in the upper mantle is through serpentinization of peridotite, which can hold up to 13% water (Ulmer & Trommsdorff, 1995). The three serpentine
minerals of interest are antigorite, with a P-wave velocity of 6.7 km/s at 1GPa, lizardite, 5.2 km/s, and chrysotile, with a velocity similar to lizardite (Christensen 1966; 1978). Lizardite and chrysotile are stable at lower temperatures than antigorite, although lizardite is likely the product of peridotite hydration while chrysotile forms in cracks and voids after hydration (Evans, 2004). Lizardite is stable to 320 °C at 1 GPa (Evans, 2004; Guillot et al., 2015), and likely is the dominant hydrous phase to ~20 km depth below the seafloor at Mariana (Cai, 2018). Antigorite, the high temperature and pressure serpentine mineral, is stable to 620°C at 1 GPA (Ulmer & Tromsdorff, 1995). Incorporating experimental results to relate seismic velocity to serpentinization and water content in ultramafic rocks, Carlson and Miller (2003) calculate that a 1% reduction in P wave velocity corresponds to 2.4 vol% serpentinization and 0.3 wt% water content.

There is uncertainty as to whether hydration causes widespread serpentinization in the mantle or is localized to the normal faults through which the water is brought down. Korenaga (2017) suggests the necessary negative pressure gradients to pump water to depth require shear stresses that exceed the yield strength of rocks, and the volume expansion of serpentinization may impede hydration by limiting access to mineral surfaces (Klein et al., 2015). This suggests rather than widespread serpentinization in the upper mantle, serpentinization is limited to joints or cracks (Miller & Lizarralde, 2016).

The Mariana Subduction Zone provides an ideal location to study bend-related hydration, with the subduction of an old (~150 Mya, Muller et al., 2008) plate and evidence of extensive forearc serpentinization (Tibi et al., 2008; Pyle et al., 2010; Barklage et al., 2015). Recently published Sv velocity results from surface wave tomography in the study region show a velocity reduction in the bend-region of the plate, interpreted as a combination of pore water and
serpentinization (Cai et al., 2018). A temperature dependence on the serpentinization of the mantle (Iyer et al., 2012) predicts the Pacific Plate at the Mariana margin should be more hydrated than younger subducting plates, such as at the Middle America Trench (Ivandic et al., 2008, others).

Two lines of co-located wide-angle refraction and multi-channel seismic (MCS) reflection data are analyzed in this study to study the effect of bend-faulting on plate hydration. The reflection data provides detailed information about the sedimentary structure, including horst and grabens created by normal faulting. The P-wave velocity provided by the wide-angle refraction data identifies areas of hydration via a velocity reduction compared to unaltered plate.

### 3.3 Methods

#### 3.3.1 Data Set

The 2012 Mariana seismic experiment consisted of 4 active source profiles centered on the central Mariana Trench. The two lines analyzed in this study are perpendicular to the trench, covering both the incoming plate and outer forearc (Figure 3.1). The experiment was carried out using the 36-element, 6600 cubic inch airgun on the R/V Marcus G. Langseth (Cruise MGL1204), towed at 9 m depth. Co-located wide-angle refraction and multi-channel seismic (MCS) reflection data were collected for the active source lines.

For the refraction survey, airgun shots were fired every 500 meters and recorded on instruments spaced ~ 20 km. Twelve 4-component short-period ocean bottom seismographs (OBS) and 8 suspended hydrophones were deployed on the North Line, and 10 short-period OBS and 9 suspended hydrophones were deployed on the South Line. The suspended hydrophones, necessary due to instrument pressure ratings, were tethered to the seafloor and stayed in the
Figure 3.1. Overview of the Mariana seismic experiment used in this study. Blue box in the inset shows the map region. The North Line stations (N8 – N20) and South Line stations (S7 – S19) used in the wide-angle refraction survey in this study are denoted by white circles. Black lines indicate ship paths during air-gun shooting for both the refraction and reflection surveys. Focal mechanisms indicating normal faulting of the incoming plate are from Emry et al. (2014). Deep Sea Drilling Project site 452 is denoted by an orange star. The trench is outlined in a thick black line, with magnetic lineations outlined in white (Nakanishi et al., 1992). Apparent plate motion (APM) is labeled on the incoming plate (Kato et al., 2003).
water column for the duration of the deployment. Of the deployed instruments, 2 OBSs and 2 hydrophones on the North Line, and 1 OBS on the South Line did not recover usable data. Most shot locations were revisited 1-2 times in order to allow for data stacking to increase signal-to-noise.

For the reflection survey, airgun shots were fired every 12.5 m and recorded by a 646 channel, 8 km long streamer.

3.3.2 Reflection Dataset

The commercial software Paradigm Echos was used to initially process the MCS reflection data. Using common depth point (CDP) gathers at 6.25 m, velocity profiles were picked for every 50th CDP. Channels with high noise levels were removed. Data was deconvolved and filtered to 20 - 100 Hz, and processed with a normal move out correction, CDP stacking, and time migration using a Time-Space Kirchoff migration.

Once the refraction profiles were completed, the P-wave velocity models were used to do a depth migration using an extended split-step migration algorithm (Kessinger and Stoffa, 1992; Stoffa et al., 1990; Lizarralde & Holbrook, 1997). The velocity profiles were sampled 6.25 m in the E-W direction and 5 m in depth, then smoothed with 10 passes of a 3 point boxcar filter. The depth migration was applied to stacks exported from Paradigm Echos, filtered 5 - 50 Hz, and padded to 30 seconds. The stacks included a pre-stack normal move out correction using the picked velocity functions on data filtered 20 - 100 Hz.

3.3.3 Refraction Dataset

The P-wave velocity tomographic inversion is done iteratively by tracing rays through a velocity model, and then solving the damped least squares problem to invert for the preferred
slowness perturbations based on the travel time residuals (VMtomo: Van Avendonk et al., 1998; 2004, modified by A. Harding at SIO). The ray tracing is done using the graph method, calculating the total travel time between the source and all points and taking the global minimum path time by Fermat’s principle (Moser et al., 1992; Toomey et al., 1994; Van Avendonk et al., 1998). The chi-squared value, the normalized sum of squared deviations between travel time observations and predictions, is used to assess the model fit. A value of 1 indicates the model is sufficient given the error associated with the travel time picks. The target chi-squared value is slowly reduced to 1, alternating between ray tracing and inversion.

Data was arranged as common instrument gathers, co-located shots were stacked, deconvolved, and filtered to 4-14 Hz. First arrivals were picked on either the vertical or hydrophone component for refractions through the forearc sediment, forearc crust, plate sediment (Ps), plate crust (Pg), and plate mantle (Pn). Reflections off the Moho (PmP) were also picked when possible for the incoming plate. The arrivals were assumed to have an uncertainty of 50 ms for all picks.

For data used to constrain the incoming plate structure, the most straightforward instruments to pick were located further out on the plate (Figure 3.2). Many stations, including three component OBSs and suspended hydrophones, had clear arrivals for Pg, PmP, and Pn. Some stations on the South Line, however, had weak Pn arrivals, most notably to the eastward direction on S15 and for all shots fired > 170 km east of the trench (Figures A3.14-A3.18).

Closer to the trench, the bathymetry drastically effects the arrivals times. Figure 3.3 shows the topographic effect on S12, which is located 66 km from the trench. The arrivals west of the station are slowed due to the dip of the plate. In order to be confident in our phases, we
Figure 3.2. Data from ocean bottom seismograph N16 on the North Line, 154 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off of the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure 3.3. Data from tethered hydrophone S12 on the South Line, 66 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off of the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure 3.4. Data from tethered hydrophone S07 on the South Line, 34 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off of the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

picked the first arrivals after shifting the data to account for the change in bathymetry. Also of note, shots that traveled through the forearc are highly attenuated (west of -70 km from S12 on Figure 3.3), and this is observed on all of the stations near the trench (Figures A3.2, A3.3, A3.4, A3.12 & A3.13).

Energy is also lost for arrivals recorded on the hydrophones suspended in the deepest region of the study area. Instruments N08, N10, and S09, all suspended greater than 2 km above the seafloor, recorded Pg and PmP, but Pn is not observed (Figures A3.1, A3.2, & A.311).

Instrument S07 (Figure 3.4) located 34 km west of the trench on the forearc, was able to record Pg, PmP, and Pn from the Pacific Plate. S08, also on the forearc, only recorded a few arrivals from the forearc but not the plate. S10 did not record any first arrivals and was not used for the inversion.
In order to best determine the instrument locations, instruments were relocated to the first arrival shot in distance along the line. Instrument depths were determined by using the two-way travel time of the first arrival shot and water velocity as in the starting model. For suspended instruments, the reflection off the seafloor was also used to constrain the depth.

The starting models for the inversions were constructed for each line based on the MCS profiles and previous studies of the area. The MCS profiles were used for bathymetry, water

Figure 3.5. Starting models for the North and South Lines. Short period stations and hydrophones are denoted by inverted, red triangles.
velocity, sediment thickness, and sediment velocity. The subducting plate interface was taken from Oakley et al. (2008) and oceanic plate thickness set to match the East Line (Feng, 2016). The starting velocity structure for the crust and mantle was estimated and adjusted to roughly match first arrivals. The starting models started with chi-squared values of 4.08 and 11.22 for the North and South lines respectively, and are shown in Figure 3.5.

The model was inverted in steps, starting with the shallowest layer and working deeper. Initially the forearc was inverted, then the plate sediment, crust, and mantle. The final models have a chi-squared value of 1.04 and 1.02 for the North and South lines respectively. The ray coverage and data fits for the final models are shown in Figures 3.6 and 3.7.

3.4 Results

3.4.1 North and South Line Profiles

The MCS profiles (Figure 3.8) show the shallow sediment structure and confirm bend related faulting near the trench. The sediments, in particular the western side of the lines, is characterized by high amplitude reflections at ~ 200 m depth, overlain by weaker reflections. Below ~300 m depth, not much is seen. Larger seamounts on the plate disrupt the sediment layering. In the bending region of the plate, horst and graben structures are seen within 60 km and 50 km of the trench, and smaller offset faulting is seen out to 95 km and 80 km for the North Line and South lines respectively.

For the P-wave velocity models (Figure 3.9), the two lines have similar structure far away from the trench, but have noticeable differences as the plate bends. Away from the trench for both lines, the crust reaches a maximum of ~7.2 km/s at the base and mantle velocities are ~8
km/s. The crustal velocities are relatively constant across the lines, with a gradual reduction in the lower crustal velocities with decreasing distance to the trench. There is a noticeable reduction in the upper crust within 40 km of the trench for the North Line. The South Line has a large crustal anomaly between 45 and 80 km east of the trench.

The mantle velocities also decrease closer to the trench, with a larger velocity reduction for the North Line. The North Line has a minimum velocity of 7.28 km/s 50 km east of the trench, but the South Line does not reach its minimum velocity of 7.32 km/s until 10 km west of

Figure 3.6. Ray coverage of the North Line plotted over the final P-velocity model. Bottom panel shows data fit, with travel time predictions plotted in black over first arrival picks in red. Time is reduced to 8 km/s.
Figure 3.7. Ray coverage of the South Line plotted over the final P-velocity model. Bottom panel shows data fit, with travel time predictions plotted in black over first arrival picks in red. Time is reduced to 8 km/s.

the trench. The South Line has better ray coverage crossing the trench (Figure 3.7), allowing us to see more of the subducted plate.

To address model uncertainties and robustness of velocity anomalies in the refraction profiles, we traced rays for models after smoothing over the velocity anomalies to compare with observations. For the plate mantle, the velocity in the bending region was changed to match the velocity at the eastern end of the line where the plate is assumed to be largely unaltered. This model with no mantle slow anomaly is shown in Figures 3.10a and 3.11a for the North and South Lines respectively. Tracing rays through the models resulted in Pn first arrival times that are too early to match the observed data, and are illustrated by the yellow dots in Figures 3.10 and 3.11.
Not only are the arrivals too fast, the slopes of the first arrivals are also too flat, indicating that the velocity is too fast. The North Line has a greater discrepancy between the predicted travel times for the fast mantle model (yellow dots) and the observed data (red dots). This is consistent with the greater velocity anomaly in the North Line tomographic model.

The South Line has a large reduction in velocity in the crust 45 and 80 km from the trench. A model without this anomaly is shown in Figure 3.12a. The two stations on either side of the anomaly are shown in Figure 3.12b and 3.12c. To the east of S11 and west of S12, predicted first arrival times are too early for Pg in the model without the anomaly (yellow dots). The slopes for the Pg arrival times are also too flat, again signifying that the data requires a
Figure 3.9. Final P-velocity models for the North and South lines. Short period stations and hydrophones are denoted by red, inverted triangles. Areas with no ray coverage are shaded out. Slower velocity structure. First arrivals for Pn and PmP are also shifted earlier, although the slopes relatively stay the same. This indicates that while the velocity structure for the mantle is reasonable, the crust is too fast.

3.4.2 Modeling Faults in the Plate Mantle

The P-velocity profiles provide evidence for serpentinization of the mantle as the incoming plate bends. However, the tomographic inversion imposes smoothing of the model, and our results cannot distinguish between a truly smooth structure in the mantle and finer
Figure 3.10. (a) Model without a velocity reduction in the mantle for the North Line. Seismic data from stations N12 and N13 are shown (b) and (c). Data is plotted by distance relative to each station, with time reduced to 8 km/s. Picked first arrival times are plotted in red, predicted travel times of the final P-velocity model in blue, and predicted travel times of the modified model without the mantle velocity anomaly in yellow.
Figure 3.11. (a) Model without a velocity reduction in the mantle for the South Line. Seismic data from stations S12 and S15 are shown in (b) and (c). Data is plotted by distance relative to each station, with time reduced to 8 km/s. Picked first arrival times are plotted in red, predicted travel times of the final P-velocity model in blue, and predicted travel times of the modified model without the mantle velocity anomaly in yellow.
Figure 3.12. (a) Model without a crustal velocity anomaly for the South Line. Seismic data from stations S11 and S12 are shown (b) and (c). Data is plotted by distance relative to each station, with time reduced to 8 km/s. Picked first arrival times are plotted in red, predicted travel times of the final P-velocity model in blue, and predicted travel times of the modified model without the crustal velocity anomaly in yellow.
structure that is averaged over the ray paths. Water is assumed to travel to mantle depths along normal faults (e.g. Ivandic et al., 2008; Faccenda et al, 2009), so it is reasonable to consider a case in which hydration and serpentinization are limited in extent to these normal faults (Miller and Lizarralde, 2016). Considering the permeability of serpentine, constrained by rock samples, and the fluid transport rate through serpentine to the reaction front, Hatakeyama et al. (2017) model the rate and extent of serpentinization of the bend-related normal faults. These results suggest serpentinization can occur to 100s of meters width prior to reaching the trench, but do not cause widespread serpentinization of the mantle.

In order to provide bounds for the amount of water bound in the plate, we consider the case in which hydration and serpentinization are limited to cross-cutting normal faults. We model this case as cracks of a slower velocity material in a background mantle velocity of 8 km/s. We then calculate first-arrival travel times through this model with cracks, and compare to observed data. For both lines, the crack locations are determined using the MCS reflection results (Figure 3.8) and corroborated with fault scarps from bathymetry.

The faults were determined to be seaward or landward facing, but assumed to have a dip of 60° as expected of Andersonian faulting (Anderson, 1951) and consistent with focal mechanisms of teleseismic normal faulting (Emry et al., 2014). The cracks taper to a point at 24 km depth, which is more than sufficient to cover ray paths which extend to 6 km below the Moho. For simplicity, initially all of the cracks were set to the same width, although the damage zone of the faults likely grow with decreasing distance to the trench (Hatakeyama et al., 2017) and also may be dependent on fault throw (Mitchell & Faulkner, 2012; Boneh et al., 2019). An example of the North Line with 600 m wide cracks in background mantle is shown in Figure 3.13.
Figure 3.13. Example of a velocity model with background mantle at 8 km/s and 600 m cracks of a lower velocity material used to estimate water content.

Figure 3.14. Width of serpentine crack as a function of time based on Hatakeyama et al. (2017). Lines represent width of crack at 0.2 My increments, with width at 1 My and 2 My bolded and labeled. This is for the South Line, with crack width growing from where the first normal fault is observed in the MCS profile at 86 km from the trench to 700 km wide at the trench, after about 2.4 My.
We also model cracks assuming a thickness progression with decreasing distance to the trench. Assuming the serpentinization is controlled by the water flux delivered to the reaction front, the lateral extent of serpentinization can be expressed as a function of permeability, pressure difference, porosity, fluid viscosity, and time (Hatakeyama et al., 2017). Varying only time and assuming the other parameters stay constant in the study area, the width of serpentinization increases by the square root of time. Assuming constant convergence of 4 cm/yr throughout recent history and that all observed faults initially formed at the same distance from the trench, the width of the mantle cracks can be varied as a function of time. We assume the furthest fault observed in the MCS is 50 m wide, based on the model resolution, 102 km from the trench for the North Line and 86 km from the trench on the South Line. The fault widths are then increased as the square root of distance multiplied by a constant so that they reach a set maximum fault width at the trench. We do not consider change in permeability with depth, as in Hatakeyama et al. (2017). An example of the rate of growth is shown in Figure 3.14.

To consider the case in which the cracks are filled with serpentine, the velocity within the cracks are taken to be lizardite/chrysotile (5 km/s) to be consistent with modeling of Miller and Lizarralde (2016) or antigorite (6.3 km/s, Wang et al., 2019). Adding more complexity and to consider the effect of water, the cracks were also modeled as a linear velocity gradient from 8.0 km/s at the edge to 2.2 km/s at the center. The velocity in the center is based on serpentine gouge recovered by Ocean Drilling Project Leg 180 with 30% porosity and calculated seismic velocity ~2 km/s (Floyd et al., 2001; Kopf, 2001; Wyllie et al., 1958).

Travel times were calculated for rays traveling through the cracks by assuming ray paths between the shot and receivers and calculating the travel times piecemeal. The ray paths were determined by tracing rays through a model with background mantle of 8 km/s near the trench.
(same model as shown in Figures 3.10 & 3.11), and were fixed in this way because the graph method for ray tracing preferentially avoids tracing rays through these cracks. For the South Line, rays traveling through the subducted portion of the plate were not included since the fault distribution can only be determined on the plate prior to subduction. Different crack widths were sampled to find the width that minimized the chi-squared between calculated travel times and first arrival picks for mantle refractions (Pn). Figure 3.15 summarizes the preferred crack models for the various scenarios tested.

In order to calculate serpentinization percent or water content of the various models, we consider the mantle between 5 - 70 km from the trench, where the velocity reduction is

![North Line](image1.png) ![South Line](image2.png)

**Figure 15.** Chi squared representing the model misfit for various crack widths for all the cases tested in this study.
concentrated. The mantle is sampled at 10 m intervals 100 m below the Moho and averaged to get a representative number. For the results of the velocity inversion (Figure 3.9), the degree of serpentinization is estimated based on a linear gradient of fraction serpentine between background mantle (0% serpentine, 8 km/s) and lizardite (100% serpentine, 5 km/s) (Christensen, 2004; Ji et al., 2013). Serpentine is taken to have 13 wt% water (Carlson and Miller, 2003) to calculate water content. In the case of cracks with a linear gradient, areas with velocity greater than 5 km/s are treated as percent serpentinization, as above. Velocities less than 5 km/s are assumed to be lizardite with water filled porosity. The porosity was calculated from velocity using the Wyllie time-averaged equation (Wyllie et al., 1958). This relationship was chosen so no assumptions would need to be made for the aspect ratio of the porosity (Mavko et al., 2009; Korenaga, 2017).

Given the many assumptions in the travel time calculation, synthetic seismograms were calculated for an instrument above the low velocity zone. A pseudo-spectral synthetic acoustic wave propagation code (Kosloff & Baysal, 1982; Cerjan et al., 1985) was used to calculate synthetics for S12 on the South Line. Synthetics were first calculated for the inverted P-velocity model of the South Line as a baseline. Synthetics were then calculated for: a model with unaltered mantle with 600 m wide lizardite cracks (5 km/s); a model with unaltered mantle with 500 m wide cracks that linearly change in velocity from 8.0 km at the edges to 2.2 km/s in the center; a model with lizardite cracks (5 km/s) that increase from 50 m wide at 86 km east of the trench to 900 m wide at the trench; and a model with cracks with a linear velocity to 2.2 km/s at the center, with widths growing from 50 m wide at 86 km east of the trench to 700 m wide at the trench. Figure 3.16 shows synthetics for the first 2 models with constant crack width compared with the observed first arrival time picks. To first order, the synthetic seismograms show that the
Figure 3.16. Synthetic seismograms for S12 to determine accuracy of travel time calculation for models with cracks. Red dots are first arrival picks from real data. (a) Real data, (b) synthetics for final P-velocity model, (c) model with background mantle 8 km/s and linear velocity gradient to 2.2 km/s in 500 m wide crack, (d) model with background mantle 8 km/s and 600 m wide cracks at 5 km/s.
Figure 3.17. Zoom to westward Pn arrivals for synthetic seismograms for S12 to determine accuracy of travel time calculation for models with cracks. Turquoise lines represent first arrivals calculated from ray paths. (a) Synthetics for model with mantle at 8 km/s, (b) for model with background mantle 8 km/s and linear velocity gradient to 2.2 km/s in 500 m wide crack, and (c) model with background mantle 8 km/s and 600 m wide cracks at 5 km/s. Calculated travel times match synthetic first arrivals in (a), but are late for (b) and (c).
models with cracks are able to replicate observed travel times and produce first arrivals that are not much different from the observed data. The coda is different for the models, but given the attenuation of the coda in the observed data, it is difficult to argue for one scenario over another. It is worth noting, that Pn arrivals for the model with velocity of 2.2 km/s in the center of the crack (Figure 3.16c) are noticeably attenuated compared to the other models.

We did find, however, discrepancies between the calculated travel times using the assumed ray configuration and the synthetic first arrivals for the models with cracks. Figure 3.17a shows that for the background model (mantle at 8 km/s) the synthetics and calculated predictions match. However, after introducing cracks, the calculated travel times arrive too late compared to the synthetics (Figure 3.17b and 3.17c). For the 4 models synthetics were calculated for instrument S12, the difference between the calculated travel times using rays and the first arrivals for the synthetics ranges from root mean squared of 0.04-0.07 seconds, with increasing error with larger cracks. Considering the worst case of 0.07 seconds RMS error and just the westward Pn picks on S12, this correlates to a change in calculated chi squared of 0.98. When considering all Pn picks from S12, the change in chi squared is reduced to 0.32, since the error is buffered by the largely unchanged eastward picks.

We attribute this difference to finite-frequency affects, where longer wavelengths are not as sensitive to the slowing in the cracks. What this indicates is that our estimates for preferred crack widths underestimate the crack width necessary, and the chi-squared plots (Figure 3.15) should be shifted towards greater crack widths. However, to fully address this issue we would need to calculate synthetics for all instruments and crack scenarios, with a method that is fully elastic and accounts for attenuation, and is beyond the scope of this paper.
3.5 Discussion

3.5.1 Background Pacific Plate
Sedimentary Structure from MCS Reflection Data

Both lines have a similar sedimentary structure in the MCS reflection profiles, with a strong reflector observed at ~ 200 m depth, overlain by a weaker reflective layer. Without drilling cores to confirm the layers, it is difficult to attribute reflectors to particular sediments. Ocean drilling east of the study area in the Pigafetta and East Mariana Basins paired seismic stratigraphy with drill cores (Ocean Drilling Project (ODP) Leg 129, Sites 800, 801, 802: Lancelot et al., 1990). Reflective layers correlate with both chert layers and volcaniclastic turbidites, and a bright reflector at ~500 m below the seafloor is identified as the basement. This basement is interpreted as mid-Cretaceous volcanic sills and flows when observed as a smooth reflector, and as the actual Jurassic oceanic crust when rough (Abrams et al., 1992).

The basement is not obvious in the MCS profiles from this study, and not much is seen beyond 300 m depth. An intermittent reflector is observed at ~500 m, and may correlate with the Jurassic oceanic crust observed in drilling and seismic stratigraphy in the East Mariana Basin. Given how smooth, flat lying, and bright the reflector is at 200 m depth, this may be a Cretaceous volcanic sill or flow, overlain by sediments including chert and volcaniclastic turbidites. Drilling south of instrument N13 (Figure 3.1, site 452, Deep Sea Drilling Project Leg 60) only penetrated a few tens of meters of pelagic sediment before encountering an impenetrable chert layer, confirming the presence of chert at shallow depths (Hussong et al., 1982).
Crustal Structure

The general crustal structure is consistent with typical oceanic crust. The oceanic Layer 2, composed of extruded lava overlaying sheeted dikes, is characterized by velocities ~3 to ~6.7 km/s and a steep velocity gradient. The gabbroic Layer 3 is marked by a change in the velocity gradient and velocities ~6.7 to ~7 km/s (Grevemeyer et al., 2018, others). For the seismic lines in this study, the Layer 2-3 boundary can be estimated by the 6.6 km/s contour for the background crustal structure based on the change in gradient.

The South Line is characterized by a thicker crust, at 8 km thick compared to 7 km for the North Line. Layer 2 is also noticeably thicker in the South Line. The western Pacific underwent widespread intraplate volcanism during the Cretaceous that resulted in the formation of several seamounts (Koppers et al., 2003; Kaneda et al., 2010). In the context of this volcanism and the numerous seamounts on the plate where the South and East lines intersect, the thickening of the crust is likely a product of intraplate volcanism (Feng, 2016).

The South Line is marked by a large crustal velocity reduction 45 - 70 km east of the trench. The anomaly does not correlate to large horst and grabens, and is likely not associated with bend-related faulting and subsequent hydration. Similar oceanic crustal velocity reductions have been observed globally, and are indicative of petit spot volcanism (Ohira et al., 2018). Petit spot volcanism occurs when subduction related bending of the plate causes magma ascension from the asthenosphere (Hirano et al., 2006). The volcanism is also characterized by opaque and disrupted structure at the top of the oceanic crust (Fujiwara et al., 2007) and unclear Moho in MCS data (Ohira et al., 2018). The sedimentary structure on the South Line is disrupted 55 - 70 km from the trench, with a seamount sitting at 60 km (Figure 3.18b).
The North line shows a velocity decrease across the crust within 40 km of the trench, although a similar reduction is not observed to the south. One interpretation is that the velocity reduction is due to hydration and alteration of the crust since it occurs closest to the trench where hydration is expected to be greatest. However, a similar anomaly is not observed in the South Line. The reduction also correlates with irregular sedimentary structure in the MCS reflection.

**Figure 3.18.** MCS profiles showing areas on North and South Lines that have disrupted sedimentary layers, which may be indicative of petit-spot volcanism. (b) A seamount is imaged at 62 km east of the trench on the South Line, at the center of the disrupted layering between 55-70 km.
profile (Figure 3.18a). This may suggest that this anomaly is also related to petit spot volcanism similar to the anomaly at 45-70 km on the South Line.

**Background Mantle Velocities**

The background upper mantle velocity observed in this study is \( \sim 8 \) km/s, lower than expected for typical Pacific Plate. The direction at which seismic velocities propagate fastest should be aligned with the paleo-spreading direction (Shearer & Orcutt, 1985), perpendicular to the magnetic lineations which strike 38° from North (Nakanishi et al., 1992). The South Line is oriented 28° from the expected fast direction, and the North Line 37°. Based on unaltered Pacific Plate with azimuthal anisotropy in this orientation, the expected upper mantle velocity is closer to 8.2 km/s (Mark et al., 2019).

Intraplate igneous activity can cause underplating and thickening of the crust (Kaneda et al., 2010) and velocity reduction in the lower crust and mantle (Farnetani et al., 1996). The East Line, which runs perpendicular to the North and South lines, similarly observes slower mantle velocities and thickening of the crust due to Cretaceous intraplate volcanism (Feng, 2016). We interpret the slow mantle velocities as remnant from this volcanism.

**3.5.2 The Hydrated Mantle**

**Bulk Hydration of the North and South Lines**

Interpreting the mantle velocity reduction as bulk serpentinization yields an estimate of 21 vol% and 17 vol% serpentinization for the North and South lines. The \( S_v \) velocity model from surface wave tomography in the same region shows a reduction in the mantle to \( \sim 3.8 \) km/s under the bending region of the plate from a background mantle value of 4.51 km/s. P velocities
from this study suggest the minimum velocity observed is 7.3-7.5 km/sec. Increasing the Sv velocity by 1% to get the isotropic S-velocity (assuming 4% radial anisotropy as in PREM (Dziewonski & Anderson, 1981) and a Voigt average shear velocity (Montagner, 2006)), the Vp/Vs ratio is ~1.9. This value is consistent with Vp/Vs of 1.9 for 20 vol% serpentinized peridotite (Ji et al., 2013). This Vp/Vs comparison does not include the effect of free water, as interpreted in the un-subducted bending-region of the plate in Cai et al. (2018) and in this study where the velocity reduction is represented by porous fault gouge.

The estimates for mantle bound water at Mariana are on the higher end of the global spectrum. A similar upper mantle velocity of 7.3 km/s is observed at Tonga, interpreted as ~30 vol% serpentinization, where the subducting plate is younger at ~ 80 Ma but the convergence rate is much higher 20-25 cm/yr and a graben on the incoming plate is 1.5 km deep (Contreras-Reyes et al., 2011). Serpentinization estimates for the subducting 120-130 Ma Pacific plate are 25 vol% (~3 wt% water) at the Japan Trench and < 20 vol% (~2 wt% water) at the Kuril Trench, with the difference attributed to incoming plate fabric orientation (Fujie et al., 2018). For the much younger plate subducting at the Middle America Trench, mantle serpentinization ranges from 10 – 25 vol% where 24 Ma plate is subducted (Van Avendonk et al., 2011; Grevemeyer et al., 2007; Ivandic et al., 2008; Naif et al., 2015). Offshore Northern Chile, an estimate of 17% serpentinization (2.5 wt% water) is observed in 48 Ma plate (Ranero & Sallarès, 2004), while offshore southern Chile, up to 9 vol% serpentinization (~1 wt% water) is observed in 9 Ma plate (Contreras-Reyes et al., 2007). The Cascadia subduction zone, subducting < 10 Ma plate, has a dry mantle with estimated water content < 0.3 wt% (Canales et al., 2017; Horning et al., 2016).

The North Line has a greater velocity anomaly than the South Line, indicating a greater degree of hydration. This cannot be simply explained by the greater number of faults to the
north since greater crack widths are required to match observations in the North Line for all tested scenarios (Figure 3.15). A notable difference between the lines is that the faulting starts much farther out in the North Line, as seen in the reflection profile (Figure 3.8) and noted by other studies (Oakley, et al., 2008; Zhou and Lin, 2018). The velocity anomalies reflect this difference, where the North Line reaches 7.8 km/s at 110 km from the trench, but the South Line reaches the same contour at 80 km. Since distance from the trench is a proxy for time, a fault on the North Line is longer exposed to seawater and may cause more serpentinization of the mantle.

Other differences between the two lines include crustal thickness and faulting geometry. The South Line is more affected by intraplate volcanism, with a thicker crust and evidence of petit-spot volcanism in the crust. The increase in crustal thickness may impede water reaching the mantle. The fault scarp geometry also changes between the two lines, with a transition occurring at 17.6 °N, with both new faults and reactivated abyssal hill fabric forming to the south, and only new faults formed due to bending to the north (Oakley et al., 2008). This is somewhat consistent with observations in Japan, where more hydration is observed where new faults are formed at the Japan Trench as opposed to reactivated plate fabric at the Kuril Trench (Fujie et al., 2018). The difference, however, is that the fault scarps subducted at the Japan Trench are significantly greater than at the Kuril trench, whereas the fault scarps on the South Line have a greater maximum magnitude when compared to the North Line.

**Estimating Hydration by Modeling Faults**

We have estimated the amount of water in the downgoing plate for the case in which hydration and serpentinization are limited to where we see normal faults in the reflection profiles (Table 3.1 & 3.2). The chi-squared values for different crack widths are shown in Figure 3.15.
### Table 3.1. North Line plate hydration estimates

<table>
<thead>
<tr>
<th>North Line Models</th>
<th>Serpentinite vol%</th>
<th>Water wt%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inverted model, bulk serpentinization</td>
<td>21.0</td>
<td>2.7</td>
</tr>
<tr>
<td>Antigorite cracks (6.3 km/s), constant crack width 1800 m</td>
<td>31.6</td>
<td>4.1</td>
</tr>
<tr>
<td>Lizardite cracks (5 km/s), constant crack width 800 m</td>
<td>13.1</td>
<td>1.7</td>
</tr>
<tr>
<td>Lizardite cracks (5 km/s), 50m at 102 km &amp; 1100 m at trench</td>
<td>15.7</td>
<td>2.0</td>
</tr>
<tr>
<td>Linear gradient to 2.2 km/s in crack, constant crack width 600 m</td>
<td>6.8</td>
<td>0.6</td>
</tr>
<tr>
<td>Linear gradient to 2.2 km/s in crack, 50 m at 102 km &amp; 800 m at trench</td>
<td>8.2</td>
<td>0.7</td>
</tr>
</tbody>
</table>

### Table 3.2. South Line plate hydration estimates

<table>
<thead>
<tr>
<th>South Line Models</th>
<th>Serpentinite vol%</th>
<th>Water wt%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Inverted model, bulk serpentinization</td>
<td>16.8</td>
<td>2.2</td>
</tr>
<tr>
<td>Antigorite cracks (6.3 km/s), constant crack width 1200 m</td>
<td>15.0</td>
<td>1.9</td>
</tr>
<tr>
<td>Lizardite cracks (5 km/s), constant crack width 600 m</td>
<td>7.0</td>
<td>0.9</td>
</tr>
<tr>
<td>Lizardite cracks (5 km/s), 50m at 86 km &amp; 900 m at trench</td>
<td>8.8</td>
<td>1.1</td>
</tr>
<tr>
<td>Linear gradient to 2.2 km/s in crack, constant crack width 500 m</td>
<td>4.0</td>
<td>0.3</td>
</tr>
<tr>
<td>Linear gradient to 2.2 km/s in crack, 50 m at 86 km &amp; 700 m at trench</td>
<td>4.8</td>
<td>0.4</td>
</tr>
</tbody>
</table>
The crack widths that minimize chi-squared are used to calculate the serpentine vol% and water wt%. Compared to the inverted models which have chi-squared values ~1, the models with cracks in background mantle have a larger minimum chi-squared ranging from 1.2 - 1.5. While the crack models are unable to fully represent the observed data, this exercise highlights how various assumptions affect calculation of water percent.

The calculations of serpentine vol% and water wt% include a lot of assumptions, including crack distribution, velocity of the cracks, area over which the calculation is averaged, and the effect of porosity and serpentinization on velocity. What we are able to show, however, is how different interpretations of the nature of the mantle serpentinization can affect the calculated water content of the subducting mantle. Assuming the case of lizardite or porous serpentine gouge, the calculated water content for a given velocity reduction is less than interpreting bulk hydration of the mantle. In fact, the necessary water is reduced by up to 5 times in the case we assume porous fault gouge at the center of the crack.

The crack widths that minimize chi-squared range from 500 to 1100 m for the case of lizardite and porous serpentine gouge. These fault widths are roughly the same order of magnitude as modeled in Hatakeyama et al. (2017), where permeability in the fault damage zone may allow growth of cracks 100s of meters wide. These modeled widths of the fault damage zone are on the high end of empirically gathered data, and imply fault displacements of > 1 km (Savage & Brodsky, 2011; Torabi et al., 2019). The fault displacements observed near the trench are ~300 m, with the largest displacement observed on the South Line at 520 m (Figure 3.8). The average subsurface displacement is thought to be somewhere between the average and maximum surface displacement (Wells & Coppersmith, 1994), suggesting there is a discrepancy between the fault displacement and damage zones required to explain the velocity reduction.
However, the empirical fault damage zone data does not account for the growth of serpentinization, and study of microtextures of partially serpentinized peridotites suggests serpentinization can produce new fluid pathways and sustain progressive serpentinization (Schwarzenbach, 2016; Tutolo et al., 2016). Therefore, these crack widths are likely not representative of the fault motion and initial damage zone widths.

3.6 Conclusions

The two seismic profiles at the central Mariana trench show evidence for alteration of the crust and mantle near the trench. The crustal structure is altered by petit-spot volcanism, reducing the observed velocity and disrupting layering of the sediments. The mantle structure is altered by progressive serpentinization as the plate bends, reaching velocities as low as 7.3 km/s, reduced from background mantle values of ~8 km/s. The MCS reflection data shows progressive normal faulting with decreasing distance from the trench, and is thought to provide the pathway through which water can reach mantle depths. Assuming bulk serpentinization, we get a maximum estimate of 23 vol% serpentine. However, calculating the travel times for a scenario in which hydration and serpentinization is limited in extent from the normal faults produces lower estimates for serpentinization. This highlights the importance of the assumptions that go into estimating the amount of water bound in the mantle at subduction zones.

References


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Appendix

Figure A3.1. Data from tethered hydrophone N08 on the North Line, 6 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.2. Data from tethered hydrophone N10 on the North Line, 34 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.3. Data from tethered hydrophone N12 on the North Line, 74 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.4. Data from tethered hydrophone N13 on the North Line, 94 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.5. Data from tethered hydrophone N14 on the North Line, 114 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.6. Data from tethered hydrophone N15 on the North Line, 134 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.7. Data from ocean bottom seismograph N18 on the North Line, 194 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.8. Data from ocean bottom seismograph N19 on the North Line, 214 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.9. Data from ocean bottom seismograph N20 on the North Line, 234 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.10. Data from tethered hydrophone S08 on the South Line, 14 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc)
Figure A3.11. Data from tethered hydrophone S09 on the South Line, 7 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.12. Data from tethered hydrophone S11 on the South Line, 46 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.13. Data from ocean bottom seismograph S14 on the South Line, 106 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.14. Data from ocean bottom seismograph S15 on the South Line, 126 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.15. Data from ocean bottom seismograph S16 on the South Line, 147 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.16. Data from ocean bottom seismograph S17 on the South Line, 166 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Figure A3.17. Data from tethered hydrophone S18 on the South Line, 186 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).

Figure A3.18. Data from ocean bottom seismograph S19 on the South Line, 206 km east of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps), crust (Pg), and mantle (Pn); reflections off the Moho (PmP); and first arrival times associated with just the forearc (Forearc).
Chapter 4: Seismic Constraints on the Forearc Structure and Turquoise Seamount, Mariana Subduction Zone

4.1 Abstract

The Mariana subduction zone is noted for being a non-accretionary margin, having an extensional forearc with back-arc spreading, and hosting active serpentine mud volcanoes. We investigate the sedimentary structure and P-wave velocity of the outer forearc in central Mariana using multi-channel seismic reflection and wide-angle refraction data. Using two parallel 2-D seismic surveys separated by ~100 km, we image two distinctly different slices of the forearc. To the north, we observe a sedimentary basin with localized compressional features and a steep trench slope. To the south, we observe a gradual deepening of bathymetry from the volcanic arc to trench that is disrupted by a serpentine seamount. The differences between the two lines may reflect the variable incoming plate topography. We image under the serpentine mud volcano, Turquoise Seamount, revealing a low velocity anomaly that extends to at least 10 km below the seamount. The anomaly suggests a wide area affected by the building of the seamount, and may reflect diffusive water flow and serpentine rock/mud. A fast anomaly underneath the summit of the Turquoise Seamount delineates the headwall of the extensional fault that cuts through the seamount. The outermost forearc for both seismic lines have low velocities, which may reflect tectonic erosion.

4.2 Introduction

The Mariana convergent margin is an end-member subduction zone in many regards, with a lack of large (Mw > 8) megathrust earthquakes (Uyeda and Kanamori, 1979; Emry et al.,
2011), lack of accretionary prism (Hussong and Uyeda, 1982), and active serpentine mud volcanoes near the trench (Fryer et al., 1995).

The occurrence of active serpentine mud volcanoes at Mariana suggests the forearc is largely hydrated from the incoming plate (Stern and Smoot, 1998). There is seismic evidence for serpentinization of the forearc (Tibi et al., 2008; Pozgay et al., 2009; Pyle et al., 2010), although the spread may not be homogeneous. The many ultramafic samples recovered from dredging suggest the mantle is not fully serpentinized but controlled by extensional faults in the outer forearc (Oakley et al., 2007). While there have been many reflection seismology campaigns, drill cores and dredges aimed at understanding these seamounts, the deeper structure has yet to be imaged by geophysical methods.

Globally, subsidence and alteration of the forearc has been attributed to tectonic erosion, the process by which forearc material is subducted. Frontal erosion occurs along the trench slope of the margin, while basal erosion occurs at the base of the upper plate (von Huene & Lallemand, 1990). The Mariana system is considered an erosive margin (Clift & Vannucchi, 2004), with no build up of an accretionary prism and exposed basement on the inner trench slope (Bloomer, 1983). Evidence for subsidence of the forearc is given by Ocean Drilling Project (ODP) drill site 460, where calcareous sediment was found well below the carbonate compensation depth (Hussong & Uyeda, 1982). However, subsidence in northeast Japan that had previously been attributed to tectonic erosion can be explained by changes in major plate kinematics, and similar conclusions can be drawn for Mariana (Regalla et al., 2013).

This study investigates the central region of the Mariana forearc using wide-angle refraction and multi-channel reflection seismic data. The survey includes a deeper look at
Turquoise Seamount, a serpentine mud volcano located 70 km from the trench, while comparison of the two lines highlights across arc variation.

### 4.3 Geologic Setting

Subduction initiation for the Mariana arc occurred ~50 Ma (Taylor, 1992), with initial supra-subduction volcanism forming the igneous basin of the forearc (Cosca et al., 1998; Ishizuka et al., 2006). The volcanic arc localized in the Late Eocene or Early Oligocene, while proper subduction occurred ~43 Ma (Stern et al., 2003). Sags and half grabens were formed in the forearc and backarc during protracted Oligocene rifting (Taylor, 1992). Two backarc rifting events occurred, the first starting ~31 Ma to form the Parece Vela Basin (Taylor, 1992). The second starting ~10 Ma to form the Mariana Trough, which remains today as a seafloor spreading center, and the currently active volcanic arc formed ~7 Ma (Stern et al., 2003).

The Mariana forearc today can be characterized into 2 provinces, with thick, stratified sedimentary basins near the volcanic arc and a thinly sedimented and extensively faulted outer forearc (Mrozowski & Hayes, 1980). The entire forearc is under tension (Kato et al., 2003), with normal faults cutting through the sediment and into the basement (Hussong & Uyeda, 1982). Arc-wide, recent extension appears to have reactivated faults created during the Eo-Oligocene rifting, and affects the outer forearc basement and sediment more than the inner forearc (Chapp et al., 2008). The extensional stress regime is caused in part by radial extension related to the increasing curvature of the arc (Stern and Smoot, 1998; Heeszel et al., 2008). Faulting is further assisted by vertical uplift and subsidence caused by subduction of the many guyots and seamounts on the incoming Pacific Plate (Fryer & Hussong, 1985; Fryer & Smoot, 1985).
The Mariana subduction zone uniquely hosts serpentine mud volcanoes, with 17 serpentine seamounts identified along the margin (Fryer, 2012). The seamounts are limited to the outer 100 km of the forearc (Fryer, 1992), forming wide, low relief edifices that undergo gravitational deformation (Oakley et al., 2007). The serpentine seamounts correlate with cross cutting normal faults, suggesting that slab derived fluids and fault gouge buoyantly rise along fault planes to protrude mud and rock at the surface (Fryer et al., 1995; Fryer et al., 2000). Fluids are derived from subducted sediments (Haggerty, 1991), and show systematic trends in pore water chemistry consistent with increasing pressure and temperature along the subduction interface (Hulme et al., 2010). Rock clasts included in the serpentine mud mixture are sourced from the slab interface and subducting crust, originating from at least 15 km depth (Maekawa et al., 1993; Pabst et al., 2012; Albers et al., 2019).

These seamounts are thought to be long-lived features, with serpentine bearing sediment found just above Eocene basement drilled south of Asut Tesoru at Deep Sea Drilling Project Site 459 (Desprairies, 1982), informally known as Big Blue Seamount and the largest serpentine seamount identified at Mariana. The seamounts are comprised of overlapping mudflows, indicating eruptions are episodic, perhaps triggered by earthquake activity (Fryer et al., 2000). The morphology and physical properties of the muds suggest the seamounts behave like mud volcanoes erupting highly liquid serpentine muds that rise due to their low density and strength (Phipps and Ballotti, 1992).
Figure 4.1. Figure of the Mariana seismic experiment, with the blue box in the inset showing the survey region. The North Line stations (N1 – N8) and South Line stations (S1 – S8) used in the wide-angle refraction study are white circles. Black lines indicate ship paths during air-gun shooting. The trench is outlined in a thick black line, with apparent plate motion (APM) labeled on the incoming plate (Kato et al., 2003). Serpentine seamounts are denoted by navy triangles, with Turquoise Seamount labeled (TS). Normal faults near Turquoise seamount, as identified in Oakley et al. (2007), are denoted by hashed black lines. Deep Sea Drilling Project Leg 60 (Hussong et al., 1982) forearc sites are denoted by orange stars and labeled with site number.
4.4 Methods

4.4.1 Data Set

The data set used in this study is from the 2012 Mariana seismic experiment (Figure 4.1), and includes two of the four active source profiles surveyed. The two investigated transects cross both the incoming plate and outer forearc, striking perpendicular to the trench axis. Co-located wide-angle refraction and multi-channel seismic (MCS) reflection was collected using the 36-element, 6600 cubic inch airgun on the R/V Marcus G. Langseth (Cruise MGL 1204).

For the refraction survey, airgun shots were fired every 500 meters and recorded on instruments spaced ~ 20 km. The airgun was towed at 9 m depth, and most shot locations were revisited 1-2 times in order to improve signal-to-noise by stacking the data. For the North Line, twelve 4-component short-period ocean bottom seismographs (OBS) and 8 suspended hydrophones were deployed. The suspended hydrophones were tethered to the seafloor and stayed in the water column and were required at depths > 6 km due to the instrument pressure ratings. 10 short-period OBS and 9 suspended hydrophones were deployed for the South Line. 2 OBSs and 2 hydrophones on the North Line and 1 OBS on the South Line did not recover usable data. For the reflection survey, the airgun was fired every 12.5 m and recorded by an 8-km streamer with 646 channels.

4.4.2 Reflection Dataset

The multi-channel data was initially processed using the commercial software Paradigm Echos. Noisy channels were first removed, and velocity profiles were picked for every 50th common depth point (CDP) gathers. Data was deconvolved and filtered 20 - 100 Hz, then
processed with a normal move out correction, CDP stacking, and a time-space Kirchoff migration.

A depth migration was performed once the P-wave velocity models were completed using the refraction data, using an extended split-step migration algorithm (Kessinger and Stoffa, 1992; Stoffa et al., 1990; Lizarralde & Holbrook, 1997). The P-wave models were sampled 6.25 m in the E-W direction and 5 m in depth, smoothed with 10 passes of a 3 point boxcar filter. The stacks were exported from Paradigm Echos, filtered 5 - 50 Hz, and padded to 30 seconds. The stacks included a pre-stack normal move out correction with initially picked velocity functions.

4.4.3 Refraction Dataset

The wide-angle refraction data was arranged as common instrument gathers, with co-located shots stacked. The data was then deconvolved and filtered to 4-14 Hz. First arrivals for refracted rays through the forearc sediment, forearc crust, incoming plate sediment, plate crust, and plate mantle, as well as reflections off the plate Moho, were picked on either the hydrophone or vertical component of the instruments. The arrivals were assumed to have an uncertainty of 50 ms for all picks.

In order to constrain the forearc structure, first arrival times for refracted rays through the forearc sediment (Ps) and forearc crust (Pg) were used. Arrivals associated with the forearc mantle could not be conclusively identified. As in Figure 4.2, showing data from instrument N03 on the North Line, first arrivals are interpreted as Pg and do not show an obvious change in phase that could be correlated across instruments. For instruments closer to the trench, the first arrivals eastward are delayed, as in Figures A4.4, A4.5, A4.10 and A4.11, indicating the outer forearc has a slower velocity structure.
Figure 4.2. Data from ocean bottom seismograph N03 on the North Line, 107 km west of the trench. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the subducting plate (Plate).

Figure 4.3. Data from ocean bottom seismograph S05 on the South Line, 70 km west of the trench. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the subducting plate (Plate).
Instrument S05 was deployed on Turquoise Seamount on the South Line. As shown in Figure 4.3, the arrivals within -15 and 20 km offset have a steep slope, indicating very slow material underneath the seamount. The same slowing is also observed for the westward arrivals on S06, which is located on the eastern flank of Turquoise Seamount (Figure A4.10).

Instrument depths were determined by using the two-way travel time and water velocity from the starting model. For suspended instruments, the reflection off the seafloor was also used to constrain the depth. Instruments were relocated to the first arrival shot location in distance along the line.

The P-wave tomographic inversion was done iteratively, by tracing rays through a velocity model, and then solving the damped least squares problem to invert for the preferred slowness perturbations based on the travel time residuals (VMtomo: Van Avendonk et al., 1998, 2004, modified by A. Harding at SIO). The graph method was used to trace the rays by calculating the total travel time between the source and all points and taking the global minimum path time by Fermat’s principle (Moser et al., 1992; Toomey et al., 1994; Van Avendonk et al., 1998). The model fit was assessed by the chi-squared value, the normalized sum of squared deviations between travel time observations and predictions. A value of 1 indicates the model is sufficient given the error associated with the travel time picks. The target chi-squared value was slowly reduced to 1, alternating between ray tracing and inversion.

The starting velocity models for both lines were constructed based on previous refraction studies in the Mariana region and the MCS profiles from this study. The bathymetry, water velocity, sediment thickness, and sediment velocity were set using the MCS profiles. The subducting plate interface was taken from Oakley et al. (2008), while the forearc velocity
structure was initially based on P-velocity results from Takahashi et al. (2007) and Calvert et al. (2008). This structure was then modified to roughly match first arrivals, especially in the outer forearc where previous studies had not surveyed. The chi-squared values for the forearc in the starting models are 55.27 and 21.99 for the North and South lines respectively, and are shown in Figure 4.4. The model was inverted progressively by starting with the shallowest layer and working deeper, from forearc to incoming plate. The final models of the forearc have a chi-
squared value of 1.06 and 1.04 for the North and South lines respectively. Ray coverage and data fit are shown in Figures 4.5 and 4.6.

**Figure 4.5.** Ray coverage of the North Line plotted over the final P-velocity model. Bottom panel shows data fit, with travel time predictions plotted in black over first arrival time picks in red. Time is reduced to 8 km/s.
Figure 4.6. Ray coverage of the South Line plotted over the final P-velocity model. Bottom panel shows data fit, with travel time predictions plotted in black over first arrival time picks in red. Time is reduced to 8 km/s.

4.5 Results

4.5.1 Sedimentary Structure

The reflection profiles image the sedimentary structure across the two lines, highlighting the two provinces of the forearc (Figure 4.7). Deep sedimentary basins, reaching a maximum thickness of 3 km at both lines, characterize the western ends of both lines. The basement below these basins is extensionally faulted, resulting in rotated blocks. Closer to the trench, the South Line is dominated by Turquoise Seamount, although there is a small sedimentary basin at 25-35
Figure 4.7. Multi-channel reflection profiles of the North and South Lines. Sedimentary basis are colored teal, with faults outlined with black lines. The Deep Sea Drilling Project sites are labeled in blue. The outlined box on the North Line shows zoomed-in area for Figure 4.8.

Figure 4.8. Zoom-in on outer forearc sedimentary basin on the North Line. Site 459 from the Deep Sea Drilling Project is labeled in blue. Faults are shown in black lines, white lines highlight sedimentary reflectors. Purple lines indicate compressional features on the surface.
km from the trench. The North Line has sedimentary basins continuing to the slope break, although the depth to bedrock is shallower than in the west. Neither line was able to image the Moho or the subducted plate at depth.

Turquoise Seamount is imaged on the South Line, although only the shallowest sediment is seen as coherent reflectors. Sediment can only be identified under the flanks and cannot be imaged under most of the seamount. There is also no evidence of a conduit that may have fed a vent on the seamount. The eastern flank of the seamount in particular appears to be slumping. There is a small sedimentary basin to the east of Turquoise, followed by a smaller mound of unknown origin. The bathymetry then drops into the trench, with a sharp increase in the slope ~5 km from the trench.

The North Line in comparison is characterized by a very gradual decrease in topography until 32 km from the trench, at which point there is a dramatic slope change to the trench. A shallower sedimentary basin sits between 35-85 km from the trench, with a topographic feature disrupting the basin at ~70 km from the trench. In map view, this feature is a knob, and does not extend north-south. The sedimentary basin is heavily faulted, and it appears these faults have been reactivated in compression, based on the steeply dipping stratigraphic layers and crenulated surface (Figure 4.8).

4.5.2 P-velocity Structure

The P-wave velocities of the two lines are shown in Figure 4.9. Far from the trench (> 90 km west), the two lines have a similar structure. Underneath the sedimentary layer (2 - 3 km/sec), the crust starts at 4-4.5 km/sec and reaches 6 km/sec at 3-4 km below the sedimentary
Figure 4.9. Final P-velocity models for the North and South Lines. Short period stations and hydrophones are denoted by red, inverted triangles. Shaded out regions indicate areas of no ray coverage.

basins in the west. While the ray coverage is not as deep in the South Line, both lines reach 7 km/sec at 9 km below sea floor at 100 km from the trench.

The models start to diverge within 90 km of the trench. The North Line maintains a relatively constant structure until 50 km from the trench. The outermost forearc is drastically reduced in velocity, with much of the wedge (within 30 km of the trench) at velocities <5 km/s.
Figure 4.10. (a) Model without fast velocity anomaly at 75 km west of the trench and 10 km depth. Seismic data from instruments S04 and S05 are shown in (b) and (c). Time reduced to 8 km/s. Picked first arrival times from observed data are plotted in red, predicted travel times of the final P-velocity model in blue, and predicted travel times of the modified model without the fast anomaly in yellow.

The South Line shows much more variation, with a fast anomaly at 75-90 km from the trench that abuts a slow anomaly related to Turquoise Seamount to the east. The seamount is capped with velocities < 4 km/sec, and a wide low velocity anomaly underneath and to the east of the peak down to ~10 km. There is a slight increase in velocity east of the seamount, but the outermost wedge (within 25 km of the trench) again consists of velocities < 5 km/sec.

To address model uncertainties and robustness of velocity anomalies in the refraction profiles, we traced rays for models without the velocity anomalies to compare with observations.
Anomalies of interest include the fast anomaly underneath Turquoise Seamount, the slow velocity ‘root’ of the seamount, and the slow outermost forearc in both lines. Figure 4.10a shows a model with the fast velocity anomaly removed from 70-80 km west of the trench and about 10 km depth. The expected travel times for this model on instruments S04 and S05 are too slow (yellow dots in Figure 4.10b and c, compared to data in red), indicating a faster velocity is needed. Figure 4.11 shows a model without the deeper slow velocity anomaly under Turquoise Seamount, with predicted travel times that arrive much too early to match the data. It is of note
that the ray coverage is not as good for the seamount root (Figure 4.6), reducing the resolution of the feature. Modeling the arrivals, however, clearly demonstrates that a significant slow velocity anomaly is required to match first arrivals, although the width of the feature cannot be uniquely determined. Lastly, Figure 4.12 shows the North and South Lines with faster velocities in the outermost forearc. Predicted travel times for these models also arrive too early, demonstrating that a slow outer forearc is required to match the observed data.

**Figure 4.12.** Top panels show models for the North (a) and South Lines (b) without as much of a velocity reduction in the outermost forearc. Seismic data from stations N07 (c) and S07 (d) are shown in the bottom panels. Time reduced to 8 km/s. Picked first arrival times from observed data are plotted in red, predicted travel times of the final P-velocity model in blue, and predicted travel times of the modified model without the fast anomaly in yellow.
4.6 Discussion

4.6.1 Forearc Structure Near the Volcanic Arc

The western end of the forearc near the volcanic arc is characterized by deep sedimentary basins, as can be seen in both the MCS and refraction profiles (Figures 4.7 and 4.9). The sedimentary basins are clearly stratified, and sit atop rotated blocks and half grabens. The faulted basement likely formed during rifting in the late Eocene to early Oligocene, followed by sediment accumulation (Chapp et al., 2008).

Localized compressional features are observed at the base of these large sedimentary basins (Figure 4.7). These features were first observed in the Chapp et al. (2008) study, and interpreted as an arc-wide structural inversion event that occurred after the Eo-Oligocene rifting ended at ~29 Ma. Structural inversions are reversal of extensional features, typically occurring after a period of quiescence following rifting. The inversion identified in the Mariana sedimentary basins may be related to backarc spreading in the Parece Vela Basin (Chapp et al., 2008).

The crustal velocity structure under these basins (>90 km west of the trench) is consistent with previously published models focused on the volcanic arc and backarc (Takahashi et al., 2007, 2008; Calvert et al., 2008). The seismic velocities are interpreted as basaltic or boninitic upper crust, tonalitic middle crust, and gabbroic lower crust near the volcanic arc (Takahashi et al., 2008).

Neither profile in the study included arrivals that could be conclusively interpreted as reflections off the Moho. Similarly, the Moho disappears underneath the forearc in a refraction profile across the Izu-Bonin subduction zone (Takahashi et al., 1998). Previous results imaging
the volcanic arc at Mariana show the Moho should be ~15 km depth below sea level at ~120 km west of the trench, with low sub-Moho velocities of ~7.7 km/s compared to normal mantle of 8-8.1 km/s (Takahashi et al., 2007). This indicates that near the volcanic arc, the mantle at Mariana is characterized by lower velocities, related to crustal growth processes. Closer to the trench, however, the lower velocities are likely tied to fluid release and serpentinization of the forearc mantle. Velocity reduction of the outer forearc mantle is also observed in the surface wave S-wave velocity model of the study region, and is indicative of a highly serpentinized forearc mantle (Cai et al., 2018; Cai, 2018).

4.6.2 Turquoise Seamount

Turquoise seamount is the most pronounced feature on the South Line, with a width of 40 km and height of ~2 km. Although the interior of the seamount does not have coherent reflectors, sediment packages continuing under the seamount at its flanks are imaged, supporting the interpretation of a mud volcano constructed on top of arc crust.

Turquoise Seamount is considered currently inactive, indicated by a lack of new mud flows in side-scan sonar images and a drill core (Fryer et al., 2000). Rather it grows laterally through gravitational collapse and by incorporating forearc sediments in large basal thrusts at the base of its flanks (Oakley et al., 2007). The sediment package upslope and under the west flank of the seamount is thrust faulted, while the downslope package is not. This relates to the gravitational growth of the seamount, as it abuts higher topography to the west but is able to slump downslope to the east (Oakley et al., 2007).

The seamount itself consists of material slower than 4 km/s (Figure 4.9). This is consistent with results from Fantangisña seamount (informally known as Celestial seamount),
which determined shallow P-wave velocities from inverting MCS data; velocities greater than 4 km/s were found only under the flanks of Fantangisña seamount, while velocities reached 3.5 km/s within 1 km depth below the summit (Asafuah & Calvert, 2019). The velocities of the mound are lower than serpentine rocks (Christensen, 1966; Ballotti et al., 1992), indicating they likely represent less consolidated, younger sediment including that from mud flows (Asafuah & Calvert, 2019). Muds from the flanks of Conical and Torishima Seamounts suggest the serpentine debris compacts and dewaters with maturity (Phipps and Ballotti et al., 1992), although compaction cannot explain all changes in porosity and density on seamount flanks (Fryer et al., 2018). Based on increasing velocity with depth in the mount, Turquoise Seamount is likely constructed of mud flows and sediment that are consolidated with depth and gravitationally deformed.

The deeper structure (> 5 km below sea level) of Turquoise Seamount highlights the heterogeneity of the seamount. This may be expected based on the velocity model of the shallow structure at Fantangisña Seamount, which showed faster velocities on the western flank (Asafuah & Calvert, 2019). While the shallow structure does not have the resolution of the Fantangisña study, the P-wave model of Turquoise Seamount in this study shows asymmetry in the velocity structure throughout the seamount and root.

Turquoise seamount was built atop a large NE trending normal fault that projects through the summit and northeast flank of the edifice (Fryer et al., 2000). The fault is considered active, and the northeast flank of Turquoise appears to be disturbed by fault movement (Fryer et al., 2000; Oakley et al., 2007). The serpentine seamounts occur as protrusions in highly faulted regions, although they can also associate with breaks in slopes of the inner trench or correlate with major joints (Fryer et al., 2000). The boundary at 75 km west of the trench between the fast
velocity anomaly and slow serpentine root occurs under the summit, where the large normal fault is projected through. The change in velocity may be related to the fault, with cohesive crustal rock to the west and altered/fractured rock to the east. Similarly, South Chamarro seamount in Southern Mariana is assumed to sit atop a headwall fracture inferred from surrounding bathymetry, with the mound slumping to the southeast of the headwall (Wheat et al., 2008).

The slow velocity anomaly related to the serpentine seamount extends to ~10 km depth. While the exact width of the anomaly cannot be constrained by the ray coverage, it is clear that the seamount disrupts the crustal structure to depth and has a wide region of influence. Most fluid flow is assumed to ascend vertically along faults and conduits given the systematic variation in fluid chemistry with distance to the trench and minimal overprinting of wall rock signatures (Mottl, 1992; Fryer et al., 2016). Diverting some of the slab-derived fluid to diffusive flow and water filled cracks would create a wider low-velocity anomaly, and can be inferred from pore fluid analysis that indicate fluid-rock interaction (Fryer et al., 2018).

### 4.6.3 North Line vs. South Line

Comparison of the North and South Lines show a vastly different morphology and structure within 100 km of the trench. Superimposing the basement from the South Line on top of the North Line indicates that the western sedimentary basin and relative location of the trench are similar, while the outer forearc is significantly deeper for the South Line (Figure 4.7).

The North Line has a sedimentary basin 85 - 35 km from the trench, which has been cored during Leg 60 of the Deep Sea Drilling Project (Figure 4.7). Drill site 459 reached basement, with the oldest sediments deposited during the Middle Eocene (Hussong et al., 1982). The cored material also indicates several episodes of turbidite deposits from the Eocene to
Middle Miocene, indicating tectonic activity in the vicinity of the basin (Hussong et al., 1982). Drill site 460 on the trench slope, about 20 km from the trench, included calcareous and vitric mud, indicating subsidence to below the calcite compensation depth (Hussong & Uyeda, 1982). The turbidites are not observed at site 458, 40 km to the west, although the sampled area is in a topographic high, indicating either uplift or less favorable conditions for sediment accumulation (Hussong et al., 1982). Faults cross cut the entire basement, and exhibit compressional features including tilted beds and a crinkled surface near reactivated faults (Figure 4.8). These observations indicate a more recent, localized structural inversion, compared to the arc-wide inversion that occurred after Eo-Oligocene rifting ceased (Chapp et al., 2008).

In comparison, the South Line, which appears to be slumping into the trench as a whole, hosts Turquoise Seamount. The difference between the two lines may relate to favorable conditions for which serpentine seamounts may form. Across arc, serpentine seamounts occur in association with forearc scarps (Fryer et al., 2018), and generally have shallower slopes between the mount and trench. In comparison, the North Line is characterized by a steep inner trench slope. The steepness of the inner trench is heterogeneous along strike, suggesting a locally constrained mechanism to explain the morphological difference between the two lines.

Vertical tectonism is created by subduction of incoming plate topography and is associated with the formation of serpentine seamounts on the forearc (Bassett & Watts, 2015; Fryer & Hussong, 1985; Fryer & Smoot, 1985). Subduction of seamounts on the incoming plate causes uplift of the inner trench slope and migration of the trench (Yamazaki & Okamura 1989; Oakley et al. 2008), followed by slumping of the inner slope once the seamount has fully subducted (Lallemand & Le Pichon, 1987; Von Huene & Culotta, 1989; Oakley et al., 2008). This deformation of the outer forearc, combined with the extensional nature of the outer forearc
and serpen tinization of the mantle, likely creates a low-strength environment to allow upwelling of the low-density serpentine material (D’Antonio & Kristensen, 2004).

Vertical tectonism by the incoming topography may also be supporting the topographic high at the North Line. The trench axis is shifted westward at the location of the North Line (Figure 4.1), and a seamount is just east of the trench axis, suggesting influence of past subduction as well as imminent subduction of a smaller seamount. The localized topographic high and compressional features observed on the North Line may be related to recent and ongoing subduction of Pacific plate features.

4.6.4 Outer Forearc Toe

Both the North and South Lines have low velocity (< 5 km/s) in the outer 25 km or so of the forearc. A very low velocity toe is also observed to the north in the Izu-Bonin subduction zone (Takahashi et al., 1998), suggesting it is a feature along the Izu-Bonin-Mariana subduction system. Moving westward along the South Line, the velocity starts to increase at about 30 km from the trench, before dropping again due to the influence of the serpentine seamount. This suggests that the velocity reduction in the outer toe is unrelated to the serpentine seamount, as would be expected if also observed in the North Line and in Izu-Bonin.

Similarly low velocities were observed at the Tonga subduction zone, and interpreted as evidence of basal and frontal subduction erosion accelerated by the subduction of the Louisville Ridge (Contreras-Reyes et al., 2011). Mariana has been identified as a location of tectonic erosion, with evidence of forearc subsidence (Hussong & Uyeda, 1982) and lack of accretionary prism (Bloomer, 1983). Geochemistry of the erupted Mariana arc basalts suggest at least some sediment is incorporated into magma generation (Elliott et al., 1997), and the southern margin is
actively subsiding and tilting towards the trench, either due to basal erosion or deepening of the subducting plate due to change in dip (Oakley et al., 2008). The low velocities in this study may reflect fractured and fluid saturated material caused by tectonic erosion.

Rather than tectonic erosion, the subsidence of the Mariana forearc can be tied to backarc spreading associated with the Parece Vela basin and Mariana Trough, suggesting the margin may be non-accretionary rather than erosive (Regalla et al., 2013). Reflection profiles from Oakley et al., (2008) suggest the incoming sediment is either completely subducted or ephemerally accreted, but deeper structure is unresolved to determine if sediment is accreted or eroded at depth.

Based on the low velocities in the forearc toe, lack of accretion, and highly faulted, extensional nature of the outer forearc, we assume that at least frontal erosion is occurring. While deeper structure cannot be constrained to verify or deny basal erosion, the rough nature of the subducted plate suggests some interaction with the overriding forearc (Bassett & Watts, 2015) and enhancement of tectonic erosion (Stern, 2011).

References


Appendix

Figure A4.1. Data from ocean bottom seismograph N01 on the North Line, 146 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.2. Data from ocean bottom seismograph N02 on the North Line, 127 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).
Figure A4.3. Data from ocean bottom seismograph N05 on the North Line, 67 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.4. Data from ocean bottom seismograph N06 on the North Line, 47 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).
Figure A4.5. Data from ocean bottom seismograph N07 on the North Line, 25 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.6. Data from ocean bottom seismograph S01 on the South Line, 148 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).
Figure A4.7. Data from ocean bottom seismograph S02 on the South Line, 130 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.8. Data from ocean bottom seismograph S03 on the South Line, 112 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).
Figure A4.9. Data from ocean bottom seismograph S04 on the South Line, 94 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.10. Data from ocean bottom seismograph S06 on the South Line, 54 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).
Figure A4.11. Data from tethered hydrophone S07 on the South Line, 34 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).

Figure A4.12. Data from tethered hydrophone S08 on the South Line, 14 km west of the trench along the profile. First arrival time picks are labeled: refraction through sediment (Ps) and crust (Pg); and first arrival times associated with just the incoming plate (Plate).