Passive seismoacoustic imaging from the seafloor to the lithosphere: Methods and applications to New Zealand and Ascension Island

by

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Passive seismoacoustic imaging from the seafloor to the lithosphere: Methods and applications to New Zealand and Ascension Island

Thesis directed by Anne F. Sheehan

Passive-source seismic methods often rely on the isolation of transient signals such as distant earthquakes from a pervasive background of ambient noise. In marine seismology, discriminating signals from noise is particularly complicated due to the efficient wave propagation characteristics of the ocean and sediments, and oceanographic noise sources including long-period ocean surface gravity waves. Furthermore, high-amplitude structural reverberations near the receiver modulate and obscure teleseismic arrivals targeted by the analyst. In this thesis I further the development of methods to both accommodate the signal-generated noise and utilize the rich ambient noise wavefield in the ocean. I apply these methods to image subsurface structure at scales from the shallow sediments to the lithospheric mantle beneath the South Island of New Zealand and Ascension Island.

I first utilize ambient noise in the form of infragravity waves and Rayleigh waves, which both sense shear structure at depth, in conjunction with reverberations in P-S wave receiver functions to model shallow sediment structure offshore New Zealand using a Markov Chain Monte Carlo algorithm.

I then turn my focus to the theory and application of Rayleigh/Scholte wave noise interferometry. First I investigate the effects of bathymetric variations on microseism-band modal propagation between two hydrophones moored off Ascension Island. I model the range-dependent dispersion observed in the noise correlation functions from Ascension data as the result of double mode-converted Scholte-Moho headwave propagation, and thereby demonstrate the feasibility of probing oceanic crustal and upper mantle structure using moored hydrophone data.

Lastly I apply a combination of ambient noise and teleseismic Rayleigh wave tomography to
image the shear structure of the mantle lithosphere beneath the continental collision zone of the South Island of New Zealand. The resulting models include high-wavespeed anomalies potentially associated with subducted Pacific lithosphere and a relict Eocene passive margin, and a low-speed zone that correlates with Cenozoic surface volcanism.
Dedication

For my parents, Gabriella and George Ball.

You have been there with me every step of the way, and I love you both so much.

This is not just my PhD. This is our PhD.
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Chapter 1

Introduction

1.1 Introduction

The chapters of this thesis consist of three stand-alone scientific journal articles. The primary dataset I analyze, and the subject of Chapters 2 and 4, was collected from 2009-2010 offshore the South Island, New Zealand by the Marine Observations of Anisotropy Near Aoteroa (MOANA) experiment. The experiment deployed 30 stations equipped with broadband Ocean-Bottom Seismometers (OBS) and Differential Pressure Gauges (DPGs). Most stations were sited in water depths of 1km, atop unconsolidated sediments hundreds of meters thick. These conditions lead to great difficulty resolving teleseismic body waves for receiver function analysis of MOANA data (Chapter 2) due to high-amplitude reverberations in the water and sediment [e.g. Bostock and Trehu, 2012] but the array proved effective for measuring both teleseismic and ambient noise-based Rayleigh wave dispersion (Chapter 4).

In Chapter 2 of this thesis, I attempt to model the reverberations of teleseismic body waves in the receiver-side sediment and ocean in order to image the crust beneath the seafloor using teleseismic P-S wave receiver functions. This effort requires additional information about the velocity structure of the sediments, which I obtain from seafloor compliance and Rayleigh wave dispersion data. Seafloor compliance measurements are obtained using the transfer function between seafloor pressure and acceleration under infragravity (50-250s period) wave loading [Crawford et al., 1991]. Compliance measurements are sensitive to shear rigidity in the uppermost several kilometers beneath the seafloor, at depths that increase with forcing period. The Rayleigh group and phase
velocity measurements, obtained from both teleseismic and ambient noise data at periods of 8-70s, complement seafloor compliance data by sensing shear velocity structure in the crust and upper mantle [e.g. Yao et al., 2011]. I jointly model these three datasets using a Markov Chain Monte Carlo algorithm [Haario et al., 2006] to estimate shear velocity profiles beneath several MOANA stations. Due to low signal to noise ratios (SNR) in the receiver function data, the method developed in Chapter 2 mainly proves useful for constraining sediment layer properties.

Although teleseismic body waves proved challenging to utilize in MOANA ocean bottom data, useful Rayleigh/Scholte wave signal was resolved from both teleseismic data and ambient noise interferometry. The remaining two chapters of this work leverage the spatially coherent noise wavefield observed in both microseism (2-10s period) and Rayleigh wave (8-24s) bands to probe Earth structure beneath the sediment layer. In Chapter 3 I investigate the modeling of doubly-converted seismo-acoustic wave propagation in a 2D adiabatic approximation (Brekhovskikh and Godin, 1999), using acoustic data recorded with submerged hydrophones off Ascension Island. The Ascension hydrophones were deployed as part of the International Monitoring System (IMS), which serves the Comprehensive Test Ban Treaty Organization (CTBTO) in global monitoring of underground nuclear testing. The Ascension IMS hydrophone stations are moored in the Sound Fixing and Ranging (SOFAR) waveguide at depths of about 1km in the water column (and are thus elastically decoupled from the seafloor). Ambient noise interferometry of Ascension hydrophone data has successfully been employed at high frequencies (above 1Hz) for acoustic thermometry [Woolfe and Sabra, 2015]. In the microseism band, interferometric Greens Functions from Ascension exhibit weak dispersion, which I show in Chapter 3 can be modeled as double mode-converted Scholte waves propagating as refracted shear waves along the Moho. In addition to observing a unique propagation phenomenon, the results of this chapter imply that seismologists are not limited to the use of streamer or ocean-bottom data for imaging oceanic crustal and mantle structure.

Perhaps the most common use of noise interferometry in seismology is the process of retrieving Rayleigh wave Green’s Functions from ambient noise cross-correlations, and tomographically inverting their dispersion for lithospheric shear speed structure [e.g. Barmin et al., 2001]. In Chap-
ter 4 I focus on jointly inverting the dispersion of interferometric Green’s Functions and teleseismic Rayleigh wave arrivals [e.g. Gaite et al., 2015] for shear velocity models of the continental collision zone beneath the South Island of New Zealand. The shear velocity models I obtain by augmenting land-based interstation paths with additional oceanic measurements from both sides of the South Island afford an enhanced view of the geometry of the Australian-Pacific plate boundary compared with strictly land-based studies [Fry et al., 2014]. Primary features of my model include a low-speed zone beneath the Campbell Plateau that is correlated with Cenozoic volcanic rocks on the surface, and a high-speed zone extending westward of the Alpine Fault beneath the Challenger Plateau, the extent of which was previously unimaged by prior tomographic studies [Fry et al., 2014; Kohler and Eberhart-Phillips, 2002]. In this work I have investigated means of both combating and utilizing the noise that pervades oceanic seismoacoustic observations, developing novel methods and applying established ones to further our ability to image beneath the oceans at scales from the sediment to the mantle lithosphere.

1.2 References


Chapter 2

A Joint Monte Carlo Analysis of Seafloor Compliance, Rayleigh Wave Dispersion and Receiver Functions at Ocean Bottom Seismic Stations Offshore New Zealand

as published in Geochemistry, Geophysics, Geosystems:


2.1 Abstract

Teleseismic body-wave imaging techniques such as receiver function analysis can be notoriously difficult to employ on ocean-bottom seismic data due largely to multiple reverberations within the water and low-velocity sediments. In lieu of suppressing this coherently scattered noise in ocean-bottom receiver functions, these site effects can be modeled in conjunction with shear velocity information from seafloor compliance and surface wave dispersion measurements to discern crustal structure. A novel technique to estimate 1D crustal shear-velocity profiles from these data using Monte Carlo sampling is presented here. We find that seafloor compliance inversions and P-S conversions observed in the receiver functions provide complimentary constraints on sediment velocity and thickness. Incoherent noise in receiver functions from the MOANA ocean bottom seismic experiment limit the accuracy of the practical analysis at crustal scales, but synthetic recovery tests
and comparison with independent unconstrained nonlinear optimization results affirm the utility of this technique in principle.

2.2 Introduction

The seafloor is a challenging environment for seismologists. Noise presents a significant obstacle to the ocean bottom seismometer (OBS) analyst. A diverse range of physical processes in the ocean and solid Earth generate seismo-acoustic energy across a wide spectrum ranging from tidal periods to the microseism band and above [Crawford et al., 1991; Godin and Chapman, 1999]. This “rich wavefield” [Ritzwoller and Levshin, 2002] recorded on seafloor seismometers can be either vexatious or auspicious to the investigator, depending on the task at hand. Body-wave methods utilizing horizontal-component information such as shear-wave splitting and receiver functions are particularly difficult with OBS data. High-amplitude current-induced tilt noise is often observed on OBS horizontal channels [Webb, 1998], and a ubiquitous diffuse infragravity wavefield is recorded on all components at periods greater than about 40s [Crawford et al., 1991; Webb, 1998; Willoughby and Edwards, 2000]. Between 2-10 s period, microseismic noise prevails [Webb, 1998], and at infrasonic frequencies above 5 Hz hydroacoustic phases can be prevalent. Simple filtering is not always effective in improving OBS data quality due to this diverse noise spectrum. At typical OBS sites, even high signal-to-noise ratio (SNR) events are distorted by reverberations in the water and shallow sediment columns that overprint deeper arrivals in the receiver function [Leahy et al., 2010]. Large volumes of low shear-velocity marine sediments can also impose delays in teleseismic traveltimes that necessitate static corrections [Harmon et al., 2007]. Mitigating these effects is a subject of active interest in the marine seismology community. The seafloor noise wavefield can be utilized to elucidate velocity structure. Ambient Noise Tomography (ANT) of Rayleigh waves effectively resolves group and phase velocity dispersion at periods from 6-27 s, which can be inverted for shear-velocity structure [Lin et al., 2007]. In similar fashion the seafloor’s response to loading by infragravity waves (the seafloor compliance) from 50-250 s also depends on the shear velocity structure beneath the OBS [Crawford et al., 1991; Willoughby and Edwards, 2000]. Observed shear
resonances in the sediment column excited by ambient noise can also constrain shallow sediment velocities [Godin and Chapman, 1999]. In this paper we combine the shear velocity information obtained from the ambient noise observations (dispersion and compliance) with receiver functions in hopes of teasing out crustal arrivals obscured by sediment reverberations in the receiver functions, with application to OBS data from offshore New Zealand.

2.2.1 MOANA Experiment and Geologic Setting

The Marine Observations of Anisotropy Near Aotearoa (MOANA) experiment included the deployment of 30 broadband OBS and Differential Pressure Gauges (DPG) off both east and west coasts of the South Island of New Zealand from 2009-2010 (Figure 2.1) [Yang et al., 2012]. The purpose of MOANA is to characterize the rheological controls on upper mantle deformation and its spatial extent in the region using observations of anisotropy as a strain gauge [Zietlow et al., 2014; Collins and Molnar, 2014].

The MOANA stations off the South Island’s East coast are situated in and around the inner Bounty Trough, a Cretaceous extensional basin containing thick deposits (up to 6 km) of terrigenous and marine sediments. Prior studies estimate crustal thicknesses of 18-25 km beneath the Eastern MOANA array and suggest the presence of a seismically fast, thin (~3 km) lower crustal layer [Scherwath et al., 2003; Van Avendonk et al., 2004] of possibly relict oceanic crust.

The Western MOANA array is mostly situated on the Challenger Plateau, an area of relatively undeformed submerged continental crust of the Australian plate estimated to range in thickness from 18-27 km beneath the MOANA OBS sites [Scherwath et al., 2003; Van Avendonk et al., 2004; Wood and Woodward, 2002]. A layer of high-grade metamorphic rock in the lower crust beneath the plateau is postulated to be responsible for high lower-crustal velocities observed in active-source studies off the West coast [Scherwath et al., 2003]. Sediment thicknesses beneath the Western array are estimated from sonobuoy and seismic reflection data to range from 460800 m [Divins, 2003]. The gravity modeling of Wood and Woodward [2002] estimates sediment thicknesses to be 1-3 km in the same locations assuming constant densities in the sediment and basement of 2.33 g/cm³ and
Figure 2.1: Map showing topography and bathymetry on and surrounding South Island of New Zealand. Yellow triangles represent ocean-bottom seismographs (OBS) deployed during the MOANA experiment. Purple triangles are land stations deployed during the same experiment. Red triangles are permanent GeoNet stations. OBS stations NZ16 and NZ06 are analyzed in this paper. Red circles show IODP boreholes used to ground-truth the analysis.
respectively.

In 1975, leg 29 of the Deep Sea Drilling Project (DSDP) took sediment core samples from site 284 on the Challenger Plateau at a distance of \( \sim 100 \) and \( \sim 160 \) km from stations NZ06 and NZ16 respectively. They recovered a 200 m section of homogeneous calcareous ooze lacking in detrital minerals and terrigenous sediments and dating to the late Miocene at depth. Sonic logs of the core yield a bulk average \( v_p \) of 1.570.02 km/s. The core did not penetrate to basement rock and based on shipboard sub-bottom profiler data sediment thickness was estimated to exceed 600 m [Kennett et al., 1975].

2.3 Data

The analyses presented below reveal information about absolute subsurface S-velocities (via Rayleigh-wave dispersion, seafloor compliance) and the tradeoff between bulk velocities and interface depths (via receiver functions). These data are forward-modeled over thousands of realizations of randomly-perturbed model states via a Markov Chain Monte Carlo (MCMC) algorithm [Haario et al., 2006] (Section 3) to produce a suite of model realizations fitting the data to within a user-specified tolerance. In this section we describe the data that go into our MCMC algorithm.

2.3.1 OBS Receiver Function Estimation

Teleseismic receiver functions are estimates of the near-receiver shear-wave impulse-response to an incident P-wave, and are obtained by deconvolving the radial component seismogram by the vertical component (a proxy for the source-time function) [e.g., Langston et al., 1979; Owens et al., 1984]. We calculate receiver functions from the MOANA experiment OBS data using earthquakes of \( M_b > 6.0 \) and the epicentral distance range of 20-100 degrees (Figure 2.2). A total of 93 events were found to fit these criteria, and low-SNR events were manually culled on a per-station basis. An extensive variety of frequency passbands were evaluated. In this study we utilize a passband of 0.5-5 Hz for sedimentary layer analysis, and 0.05-2.5 Hz for deeper crustal and Moho analysis.

We use the multitaper spectral correlation algorithm of Park and Levin [2000] to produce the
Analysis was restricted to earthquakes with magnitudes greater than 6.0 and epicentral distances of 20-100 degrees. A total of 93 events were found to fit these criteria.

Figure 2.2: Distribution of earthquakes used in receiver function analysis. Analysis was restricted to earthquakes with magnitudes greater than 6.0 and epicentral distances of 20-100 degrees. A total of 93 events were found to fit these criteria.
Figure 2.3: Radial and tangential receiver function stacks for station NZ16 (a) and NZ06 (b), binned by epicentral distance and stacked with weight determined from vertical-horizontal coherence (see text for details). Azimuthal smoothing is applied via 10 degree overlap in epicentral bins. Positive arrivals are shown in blue, negative in red. The RF stack from the 30 degree bin was employed in this analysis because it contains the greatest number of high-SNR teleseisms.
receiver function stacks employed in this analysis (Figure 2.3). This method produces coherence-weighted stacks in epicentral distance bins, suppressing random noise by stacking multiple receiver functions in the frequency domain. Since the majority of high-SNR teleseisms recorded by MOANA are clustered in epicentral distance ranges of ∼30 and ∼90 degrees, only bin-averaged receiver functions from these distances were employed in this study. Average estimated slownesses corresponding to these distance ranges were incorporated into the forward modeling of receiver functions by the Monte Carlo algorithm.

2.3.2 Sediment Effects on OBS Receiver Functions

General features of receiver functions in marine sediments are common to those observed in sedimentary basins on land [Hetnyi et al., 2006; Langston, 2011; Park and Levin, 2000]. The dominant early mode-converted arrivals observed in MOANA receiver functions result from shallow impedance contrasts in the sediments (Figure 2.3). Very large shear velocity contrasts are possible across the sediment-basement contact [Godin and Chapman, 1999; Harmon et al., 2007] giving rise to a strong pulse in the receiver functions at delay times of ∼1s (depending on sediment thickness). These strong early pulses are the most coherent feature in MOANA receiver function waveforms across all epicentral distance bins (Figure 2.3). The free surface of the ocean is a near-perfect reflector for upgoing P waves in the water column, causing water multiples to be observed on OBS receiver functions [Bostock and Trehu, 2012]. Crustal multiples and unambiguous Moho-converted arrivals are not clearly observable in MOANA receiver functions.

We investigated the relative contributions of water and sediment multiples to receiver function amplitudes using synthetic seismograms [Herrmann and Ammon, 2004; Herrmann, 2013] (Figure 2.4). As model sediment thickness is increased, the sediment-basement P-S converted arrival is observed at later times, as expected. The interference behavior between sediment, water, and crustal reverberations is complex and highly sensitive to sediment properties. Figure 2.4 illustrates the subtle modulation of the dominantly sediment-controlled receiver function signal by the Moho-converted shear-wave arrival that we seek to resolve in this analysis. The red traces in Fig. 4 are
Figure 2.4: Synthetic receiver functions illustrating the effects of sediment and water layers on the RF waveform. (a.) Blue traces are synthetic receiver functions generated from model with 1 km water, varying sediment thickness, and Moho at 25 km (b.). Red traces are synthetic receiver functions generated from model with 1 km water, varying sediment, and crustal half-space (no Moho). Sediment thickness varies from 0 km (top traces) to 1.5 km (bottom traces). Sediment $v_s = 400$ m/s. (bottom) Input velocity models for case with 1.5 km thick sediment. The difference between red and blue synthetic receiver functions illustrates the subtle modulating effect of the Moho P-S conversion on the sediment-dominated signal.
computed for a model with water and sediments over a crustal-velocity half-space, and the blue traces are for a model with water, sediments, and a crustal layer over a mantle half-space (i.e. blue trace is for a model with a Moho, red trace does not have a Moho). The difference between the two reflects the small magnitude of Moho P-S conversion and reverberations that we hope to resolve from beneath the sediment signal.

2.3.3 Absolute S-Velocity Estimation

Absolute shear velocity information is estimated from Rayleigh-wave dispersion and seafloor compliance and is used to better model the sediment reverberations in the receiver functions. Both surface wave dispersion and seafloor compliance are determined using recordings of ambient noise.

2.3.3.1 Rayleigh Dispersion

OBS sites are relatively quiet at periods in the noise notch from \( \sim 10-30 \)s [Crawford et al., 1991; Webb, 1998; Willoughby and Edwards, 2000], allowing robust seafloor Rayleigh dispersion curves to be estimated in this band [e.g., Forsyth et al., 1998; Ye et al., 2013]. Rayleigh waves at these periods are most sensitive to velocity structure from approximately 15-45 km depth. Group velocity dispersion from ambient noise tomography following the method of Stachnik et al. [2008] and phase velocity dispersion derived from earthquakes and ambient noise [Lin et al., 2008, 2009] were determined using frequency-time analysis. The ambient noise tomography employed stacked one-bit normalized cross-correlations of hourly data windows between all station pairs. The resulting phase and group velocity maps were sampled at MOANA station locations to produce the dispersion curves presented here.

To estimate starting models in agreement with our Rayleigh dispersion data, we first performed a damped, linearized inversion of dispersion measurements from MOANA OBS stations. Group and phase velocity dispersion at each station was inverted for shear velocity models using the surf96 software from the CPS330 package of Herrmann and Ammon [2004]. This software performs a damped, linearized least-squares inversion for shear velocities in a stack of discrete layers.
The inversion was performed using a fixed-velocity water layer and a constant-velocity starting model ($v_s = 4.7 \text{ km/s}$). The algorithm then solves for $v_s$ in a stack of layers of increasing thickness from 1-10 km with depth, and the results are used to formulate starting models for the Monte Carlo analysis.

### 2.3.3.2 Seafloor Compliance

The nonlinear interaction of ocean surface gravity waves generates propagating infragravity-band (50-250s) waves that attenuate very little on a basin-wide scale due to their long wavelengths (10-50 km) [Webb, 1998]. Infragravity waves excite evanescent displacements on the seafloor with depth sensitivity proportional to the forcing period and elastic parameters of the subsurface, for which we seek to solve [Crawford et al., 1991; Willoughby and Edwards, 2000].

MOANA OBS instruments included a differential pressure gauge (DPG) with a relatively flat frequency response extending to infragravity periods. Pressure (from the DPG) and acceleration (from the vertical component of the broadband OBS seismometer) fields can be measured and their transfer function in the frequency domain yields the seafloor compliance [Crawford et al., 1991; Willoughby and Edwards, 2000]:

$$\xi(\omega) = \frac{k(\omega)\gamma(\omega)}{\omega^2} \sqrt{\frac{S_a(\omega)}{S_p(\omega)}}$$

where $\xi(\omega)$ is seafloor compliance at angular frequency $\omega$, $k$ is wavenumber, $S_a$ and $S_p$ are acceleration and pressure power spectral densities (PSDs) and $\gamma$ is the pressure-displacement coherence.

We estimated seafloor compliance for the MOANA stations using stacked multitaper power spectral densities computed for hour-long sliding windows over 24-hour data segments encompassing 120 days of continuous data. Windows containing earthquakes or transients were rejected automatically based on their low pressure-vertical coherence.

Subsequently a gain correction obtained from the pressure-acceleration transfer functions of Rayleigh waves from 5 large earthquakes was applied to the compliance functions. At periods much longer than the quarter-wave resonance period of the water column a Rayleigh wave at the
seafloor exerts a force on a water column of depth $H$ equal to its mass $(\rho H)$ times the seabed acceleration. Thus at periods around 30s we expect the transfer function between seafloor pressure $P$ and acceleration $A$ to be described by $P/A = \rho H$ [Filloux, 1983; S. C. Webb, Pers. Comm.]. For 28 stations we investigated, the mean and standard deviation of the normalized transfer functions are $0.9 \pm 0.1 \rho H$. We found the transfer function for NZ16 to equal $\rho H$ within our measurement uncertainty and that for NZ06 to deviate from $\rho H$ by a factor of $0.7 \pm 0.1$. Thus to calibrate our compliance curve for NZ06 we divided it by 0.7, and we did not apply a correction to NZ16. This technique is particularly useful for seafloor compliance because since the pressure-acceleration transfer function is calibrated, errors in either the acceleration response or the DPG response are corrected simultaneously. Also, if the acceleration response is assumed to be correct, error in the nominal DPG gain factor can be estimated.

We estimate compliance uncertainty $\varepsilon$ based on the pressure-displacement coherence $\gamma$ following Crawford et al. [1991]:

$$\varepsilon[\xi(\omega)] = \frac{[1 - \gamma^2(\omega)]^{1/2}}{|\gamma(\omega)|\sqrt{2n_d}|\xi(\omega)|}$$

(2.2)

where $n_d$ is the number of data windows used to estimate the coherence and compliance functions.

We find compliance values ranging from approximately $10^{-9} - 10^{-10} Pa^{-1}$ at western MOANA sites (Figure 2.5), implying similar seafloor rigidities to those modeled at sites on calcareous sediments on the East Pacific Rise [Willoughby et al., 2000]. Crawford et al. [1991] report similar compliance values for sites on hemipelagic sediments off California, and compliances as low as $3 \times 10^{-11} Pa^{-1}$ are found for sediment-free oceanic crust off Cascadia.

### 2.4 Monte Carlo Inversion of Dispersion, Compliance, and Receiver Functions

We utilize a Markov Chain Monte Carlo (MCMC) algorithm to find shear velocity models that jointly fit the dispersion, compliance, and receiver functions. In contrast to many linearized
Figure 2.5: (a.) Seafloor compliance spectrum obtained from 120 days of infragravity noise at station NZ16. Compliance uncertainty was estimated using Equation 2.2. (b.) Coherence between the pressure and vertical acceleration for the same 120 days at NZ16. Coherence is very high at NZ16 in the infragravity (0.004-0.02 Hz) and microseism (0.2-0.3 Hz) bands.
inversion schemes, the MCMC method searches a broad model space and yields a suite of output models and their associated conditional probabilities rather than converging to a single best-fitting model [Mosegaard and Tarantola, 1995]. We adapted the Delayed-Rejection Adaptive Metropolis (DRAM) Monte Carlo algorithm from the MCMC Matlab Toolbox of Haario et al. [2006]. Our misfit functional and Gaussian model distributions were configured following Shen et al. [2013]. At each time step a traditional MCMC algorithm randomly perturbs the starting model parameters and computes the joint misfit:

$$S_{JOINT}(m) = S_{SW} + \frac{1}{\kappa} S_{RF} + \frac{1}{\alpha} S_{SFC}$$  (2.3)

Where $S_{SW}$, $S_{RF}$, $S_{SFC}$ are the least-squares misfits of surface wave, receiver function and seafloor compliance data for model realization $m$. The least-squares misfit of each dataset is scaled to the surface-wave data using ad-hoc cost-function weighting parameters $\kappa$ and $\alpha$ for the receiver functions and seafloor compliance respectively. The relative cost function weights $\kappa$ and $\alpha$ are chosen by inspection between inversion runs to weight the contribution of each dataset to the joint misfit approximately equally. The scaled misfit is then used to form a Gaussian likelihood functional:

$$L(m) = \exp \left( -\frac{1}{2} S(m) \right)$$  (2.4)

A trial model $m_i$ is accepted into the posterior distribution with a probability of 1 if its likelihood is greater than the current model $m_j$. If this is not the case, a conditional probability is assigned to model $m_j$ based on [Shen et al., 2013]:

$$P_{accept} = \begin{cases} 
1, & L(m_i) \geq L(m_j) \\
\frac{L(m_i)}{L(m_j)}, & L(m_i) < L(m_j) 
\end{cases}$$  (2.5)

The end result is a set of accepted model realizations with posterior model parameter variances that depend in this case on the choice of weighting parameters. If a full data covariance matrix were available, we could construct a joint likelihood function from the product of the likelihoods of the individual datasets and present meaningful model uncertainties within a Bayesian framework following Shen et al. [2013]. However, robust receiver function noise covariance estimates for
MOANA data are difficult to determine using bootstrap or harmonic stripping techniques [Shen et al., 2013] given the small number of usable receiver functions we employ compared to typical land-based studies.

The DRAM algorithm offers enhanced resistance to local minima over traditional Metropolis-Hastings implementations and reduces the necessity of accurate initial variance estimates [Haario et al., 2006] which are specifically difficult to assess in our receiver function data given the paucity of usable events for bootstrap uncertainty analysis. Delayed rejection is a modification to the Metropolis-Hastings algorithm that allows additional local perturbation steps within each global time step if the likelihood has decreased at the current model state. To complement the benefits of delayed rejection, Adaptive Metropolis forces a global perturbation if the joint likelihood has failed to increase after a preset number of iterations (50-500 in this study), thus continually adapting the covariance of the proposal distribution. Together these modifications to the Metropolis-Hastings Monte Carlo estimator used by Shen et al. [2013] allow the analyst to optimize the tradeoff between model-space size and chain mixing time while improving the chances of convergence to the global target distribution [Haario et al., 2006]. To allow the Markov chains to evolve to a stable state before models are kept in the posterior distributions a “burn-in” period of 500-10,000 iterations is first performed in which output model realizations are discarded.

The final posterior distributions are presented as histograms of model parameters, and the mean values of the Markov Chains represent the preferred velocity model parameters.

### 2.4.1 Model Parameterization

We evaluate two different parameterization schemes at sedimentary and crustal scales. In each case the Monte Carlo algorithm searches a model space of six parameters that define either three constant-velocity layers or two sediment layers with linear velocity gradients and a discontinuity possible between them. In the linear-gradient parameterization we solve for the shear velocity at the top and bottom of each sediment layer \(v_{s1}, v_{s2}, v_{s3}, v_{s4}\) and two layer thicknesses \(h_1, h_2\). The gradients are then discretized into constant-velocity layers of equal thickness.
In the constant-velocity parameterization we solve for three velocities \( (v_{s1}, v_{s2}, v_{s3}) \) and three thicknesses \( (h_1, h_2, h_3) \).

To model \( v_p \) in the uppermost sediment layer we use the average value measured at DSDP Site 284 of \( v_p = 1.57 \pm 0.02 \) km/s. The value in the Site 284 core log does not increase strongly with depth in the upper 200 m of sediments [Kennett et al., 1975].

We use the “mudrock” equation of Castegna et al. [1985] to estimate \( v_p \) given \( v_s \) in the lower sediment layers:

\[
V_p = 1.16 \times V_s + 1.36 \text{ (km/s)}
\]  

(2.6)

Densities in all sediment layers were modeled using the empirical relation from Willoughby et al. [2000]:

\[
\rho(z) = 1.7 + (0.2/300)z \text{ (g/cc)}
\]  

(2.7)

For the crustal scale inversion we solve for a single sediment layer over a simple crustal model of two uniform layers atop a mantle halfspace. The algorithm searches over thickness and \( v_s \) of each of these layers. This parameterization was adopted to avoid overfitting the data yet still potentially resolve such targets as hypothesized fast lower-crustal layers on both sides of the island [Stern et al., 2002; Van Avendonk et al., 2004], or consolidated basin sediments [Wood and Woodward, 2002]. A \( v_p/v_s \) ratio of 1.80 was assumed in the crust and mantle, and an uppermost mantle P-velocity of 8.1 km/s was assumed based on the results of the SIGHT experiment [Stern et al., 2002; Van Avendonk et al., 2004].

### 2.4.2 Solving the Forward Problem

Synthetic receiver functions were estimated using the program hspec96p from the CPS330 package of Herrmann [2013]. Rayleigh-wave phase- and group-velocity dispersion was forward modeled using the sdisp routine of Herrmann [2013], and we use the 1D forward modeling routine of Crawford et al. [1991] to calculate seafloor compliance. For each model realization synthetic receiver functions, Rayleigh-wave phase- and group-velocity dispersion, and seafloor compliance are
calculated and compared to the observed values.

2.5 Results

2.5.1 Synthetic Recovery Test

Synthetic recovery tests were performed using the same dispersion and compliance frequencies and receiver function estimation parameters employed on the real data. For a given starting Earth model (Table 2.1), synthetic surface wave dispersion, seafloor compliance, and receiver functions were generated. We added random Gaussian noise with standard deviations of 0.035 km/s and 7*10^{-12} Pa-1 to the synthetic phase-velocity dispersion and seafloor compliance data respectively, values which are comparable to our actual measurement uncertainties. Real pre-event noise waveforms were added to the synthetic vertical and radial components before deconvolution and scaled to produce synthetic receiver functions with SNR ranging from 1-10. The synthetic data were then used as input to the Monte Carlo analysis.

Results from 10,000 iterations using a naive starting model are shown in Figure 2.6. We find that the MC method accurately and reliably recovers the “true” crustal model for receiver function SNR of 10 and above, and correctly recovers sediment thickness to SNR as low as 5. In addition to this synthetic test, Monte Carlo results were compared to those of the Nelder-Mead Simplex nonlinear optimization method [Lagarias et al., 1998] using a common starting model and found to agree well.

Table 2.1: Model parameters used for synthetic recovery test

<table>
<thead>
<tr>
<th>Layer</th>
<th>Thickness (km)</th>
<th>$v_p$ (km/s)</th>
<th>$v_s$ (km/s)</th>
<th>Density (kg/m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>1</td>
<td>1.5</td>
<td>0</td>
<td>1</td>
</tr>
<tr>
<td>Calcareous Ooze</td>
<td>1</td>
<td>1.6</td>
<td>0.4</td>
<td>1.7</td>
</tr>
<tr>
<td>Limestone</td>
<td>1</td>
<td>3.6</td>
<td>1.9</td>
<td>2.5</td>
</tr>
<tr>
<td>Crust</td>
<td>24</td>
<td>6.5</td>
<td>3.6</td>
<td>2.7</td>
</tr>
<tr>
<td>Mantle</td>
<td>Half-space</td>
<td>8.1</td>
<td>4.7</td>
<td>3.3</td>
</tr>
</tbody>
</table>
Figure 2.6: Synthetic recovery test. (a.) Comparison of input and modeled compliance, surface wave, and receiver function data. Top: Observed vs. predicted seafloor compliance data. Middle: Observed vs. predicted surface-wave phase-velocity dispersion. Bottom: Observed vs. predicted receiver functions. (b.) (green) Model used to generate synthetic compliance, receiver function, and surface wave dispersion data for synthetic recovery testing. (black) Starting model for Monte Carlo inversion (note that black model intentionally chosen to differ from green model used to generate synthetics). 10,000 random model realizations using bounds given in text were generated and tested. Resulting mean shear-velocity model from Monte Carlo inversion with receiver function SNR of 10 (red) recovers both sediment and crustal thickness, while receiver function with SNR of 5 (light blue) recovers only sediment thickness and does not deviate from the starting crustal thickness.
2.5.2 Sediment Velocity Structure from P-S Mode Conversions and Seafloor Compliance at MOANA Stations

We jointly model MOANA seafloor compliance data and receiver functions using the MC algorithm to produce sediment velocity and thickness estimates for MOANA stations NZ06 and NZ16 (Figs 2.7, 2.8). Only the first 2 s of the receiver function waveforms are used in the sediment study in order to isolate the first sediment P-S conversions, which constrain only the total travel time through the sediment column. The seafloor compliance data provide a complimentary constraint on sediment shear velocity.

To estimate sediment velocity structure we fix crustal thickness in our model to the estimates of Grobys et al. [2008]. We then construct trial starting models with sediment thicknesses ranging between those presented in the sonobuoy/refraction results of Divins [2003] (assumed to be minimum thicknesses) and the gravity modeling results of Wood and Woodward [2002].

The Monte Carlo search was performed in successive stages of 5,000 iterations with a burn-in period of 1000 iterations to allow Markov chain stabilization. Between stages the misfit parameters were adjusted to weight receiver functions and seafloor compliance equally, and the starting model for the next stage was constructed using the posterior parameter Markov chain means from the previous stage. The resulting velocity models and data fits are presented here as the means of the parameter Markov chains plus or minus one standard deviation. The posterior parameter variances are not representative of true model uncertainties due to the absence of receiver function covariance estimates in this study.

2.5.2.1 Station NZ16

Station NZ16 is located on the Challenger Plateau, roughly 150 km off the west coast of the South Island and 200 km southeast of NZ06, at a water depth of 1063 m.

Receiver functions at NZ16 show a prominent double pulse in the first 2 s of the waveform that is coherent across the 30-90 degree epicentral distance bins (Figure 2.3a). This double pulse could
Figure 2.7: Joint compliance and receiver function inversion results for station NZ06 (surface waves not included see 2.9 for inversion including surface waves), parameterized as three constant-velocity sediment layers (a,b), and as two sediment layers with linear velocity gradients (c,d). (a,c): Observed vs. predicted seafloor compliance data (top) and receiver functions (bottom). (b,d): Starting (black) and mean posterior shear velocity model (red). Blue and gray shaded regions indicate models and resulting data within one standard deviation of the posterior parameter means.
Figure 2.8: Joint compliance and receiver function inversion results for station NZ06 (surface waves not included see 2.9 for inversion including surface waves), parameterized as three constant-velocity sediment layers (a,b), and as two sediment layers with linear velocity gradients (c,d). (a,c): Observed vs. predicted seafloor compliance data (top) and receiver functions (bottom). (b,d): Starting (black) and mean posterior shear velocity model (red). Blue and gray shaded regions indicate models and resulting data within one standard deviation of the posterior parameter means.
result from multiple reverberations in a single sediment layer or indicate the presence of multiple impedance contrasts in the sediment column. When searching over a model space of three constant-velocity layers the Monte Carlo analysis of seafloor compliance and receiver functions predicts a total sediment thickness of 1.4 km but underpredicts the amplitude of the first receiver function pulse (Figure 2.7a). When linear velocity gradients are employed in the model, the receiver function fit improves markedly and the estimated sediment thickness decreases to only 800m (Figure 2.7b-d).

2.5.2.2 Station NZ06

Station NZ06 is located ∼350 km off the west coast of the South Island of New Zealand on the Challenger Plateau in 859 m of water. Receiver functions at NZ06 show high-amplitude sediment reverberations and the waveforms are more weakly correlated across epicentral distance bins than those from NZ16 (Figure 2.3).

Monte Carlo results for the three-layered model at NZ06 estimate a total sediment thickness of 1.7 km, and indicate a shallow impedance contrast at 35m but fit the compliance curve poorly (Figure 2.8a). The linear-gradient model yields a total sediment thickness estimate of 2.2 km, exhibits no shallower discontinuities and does not fit the second pulse in the receiver function (Figure 2.8b-d).

2.5.3 Crustal Thickness and Shear Velocity Structure From Monte Carlo Modeling of Seafloor Compliance, Receiver Functions and Rayleigh Dispersion

To investigate the feasibility of this technique in resolving deeper features such as the Moho, we added Rayleigh dispersion to the input data and extended the modeled receiver function waveforms to include arrivals originating from deeper structure. We use the means of the posterior model parameter distributions from the sediment analysis in Section 4.2 combined with average crustal shear velocities from the surface-wave inversion in Section 2.2.1 to form starting models for the crustal-scale analysis (Figure 2.9). Starting Moho depths were estimated using Grobys et al. [2008]. Misfit in the predicted vs. observed receiver functions is calculated to 6 s in the waveform to
include Moho P-S conversions and reverberations. Predicted vs. observed phase velocity dispersion is added to the misfit functional and the algorithm is run in the same manner as discussed in Section 4.2 above. Due to low SNR in the MOANA receiver functions the full crustal scale combined receiver function, surface wave, and compliance analysis is limited to station NZ16.

A total crustal thickness of 22 km is obtained from joint inversion of RFs, compliance and surface wave data at NZ16 (Figure 2.9). However, in the starting model a Moho depth of 22km was used based on the models of Grobys et al. [2008], and the Monte Carlo accepted no model realizations that deviated substantially from this value, as shown by the posterior crustal thickness standard deviation plotted in blue shading in Figure 2.9b. Our synthetic test showed similar behavior at receiver function SNR below 10 (Figure 2.6) illustrating that receiver function data sensitivity is poor at longer times in the waveform due to noise, and that a local minimum in the cost function around the starting crustal thickness could result from using noisy receiver functions in this method. The shallow sediment parameters were left free in the deeper inversion and a sediment layer thickness of 800m was estimated, falling near the values of 711m and 800m produced by the constant-velocity and linear gradient sediment-scale analysis (Figure 2.7).

2.6 Discussion

At station NZ16 the estimated sediment thickness of 800 m from the gradient model agrees generally with the 711 m acoustic basement depth reported by Divins [2003]. Sediment thickness estimates from both the 3-layer and gradient models fall within the 2.0±1.5km thickness estimate from gravity modeling by Wood and Woodward [2002], but the gradient model yields lower joint misfit at this station. We also observe shear resonances in horizontal-component ambient noise data at NZ16 that are potentially related to the presence of shear velocity gradients in the shallow sediments there [Godin and Chapman, 1999].

When the surface wave data are modeled in conjunction with receiver functions at NZ16 we estimate a crustal thickness of 22km, which does not deviate from the 22km starting Moho depth we assume from Grobys et al. [2008]. The implausibly low standard deviation of the posterior
Figure 2.9: (a.) Joint inversion of Rayleigh phase velocities, seafloor compliance and receiver functions at station NZ16. (b.) Inversion results at NZ16. Our mean model (red) crustal thickness of 22km does not deviate from our starting estimate (black) because noise in the receiver function at later times prevents it from constraining crustal thickness in the inversion.
crustal thickness illustrates that the joint misfit is not being substantially reduced by varying this parameter in the Monte Carlo search. This behavior was also seen in our synthetic testing (Figure 2.6). According to our synthetic tests a receiver function SNR on the order of 10 is likely necessary to constrain crustal thickness with this method. This is atypically high for events used in this study. For instance, at station NZ16 only six radial receiver functions had SNR > 3. Thus, based on these results we cannot robustly constrain interface depths below the sediment-basement contact.

At station NZ06 our three-layered model predicts a total sediment thickness of 1.7 km, within the bounds of 2.5±1.5km estimated by Wood and Woodward [2002]. A large shallower discontinuity is predicted at 35m, and the model produces no interface corresponding with the 466m depth to acoustic basement reported by Divins [2003]. The 2.2 km sediment thickness resulting from the gradient model at NZ06 also falls within the range reported by Wood and Woodward [2002], but produces poorer fits to the receiver function data than the three-layered model. The horizontal-component resonances that we observe at NZ16 and associate with the presence of shallow sediment shear velocity gradients are not prominently seen on NZ06 noise data.

It is evident from these results that the choice of model parameterization has a crucial influence on sediment models resulting from the inversion of seafloor compliance and receiver function data. Based on joint data misfit the linear-gradient model is more appropriate for NZ16, while the layer-cake model is preferred at NZ06. Our observations of sediment shear modes also support the presence of large shallow velocity gradients at NZ16 but not at NZ06.

Seafloor compliance probes bulk shear velocities while receiver functions are sensitive to impedance contrasts across interfaces. We find that the simple parameterizations investigated here are not always sufficient to jointly model features demanded by these separate datasets. For example, the compliance data for NZ06 are best fit by a linear velocity gradient, but this parameterization does not result in a model with the dual impedance contrasts required to fit both receiver function pulses. A nonlinear velocity gradient parameterized using a power law [Godin and Chapman, 1999] or spline [Shen et al., 2013] formulation may provide an improved fit to the slope of the compliance curve while allowing for the large impedance contrasts necessary to fit the receiver function data.
The evident tradeoff in data sensitivity between compliance and receiver functions also motivates the use of a trans-dimensional inversion scheme for these data, in which the model parameterization itself is treated as an unknown [Bodin et al., 2012]. In addition, data noise can be treated as a free parameter in a trans-dimensional Monte Carlo search. In such case the posterior distributions would reflect meaningful model uncertainties, rather than simply depending on ad-hoc cost-function weighting parameters as they do in this study.

Our receiver function data did not conclusively resolve crustal structure beneath the sediment layer with this method. However, an alternate technique to mitigate the overprinting effects of sediments in OBS receiver functions is re-datuming to virtual receivers below the sediments, as successfully applied on land data from the Mississippi Embayment by Langston [2011]. The success of re-datuming relies critically upon accurate a priori knowledge of sediment thickness and velocity structure. Promisingly, the method we present here can be used to estimate these parameters using P-S converted arrival times and seafloor compliance data to respectively constrain sediment thickness and shear velocity, opening up a potential research avenue for improving the utility of OBS receiver functions in the future.

2.7 Conclusions

We introduce a new method to combine seafloor compliance, receiver functions and Rayleigh wave dispersion in order to determine suboceanic crustal properties. We utilize a Markov Chain Monte Carlo approach to sample the model space and produce probabilistic estimates of shear velocity structure. We use the method to first determine shallow sediment properties, which are then used as an a priori estimate in our modeling of the deeper crust. Well-determined sediment properties are essential for modeling the sediment layer reverberations that can dominate OBS receiver functions. Our tests with synthetic data are promising. However, tests with real data are mixed. We have good success in determining shallow sediment properties from the combination of seafloor compliance and receiver functions, and good success in determining crustal shear wave velocity from the combination of seafloor compliance and Rayleigh wave dispersion. Our results
with synthetic and real data indicate that resolution of the Moho interface from OBS receiver functions remains elusive, even in combination with a well-determined sedimentary layer. Redatuming methods may hold more promise, and will rely on a well-determined sedimentary layer, as provided here. Our success in determining shallow interface properties suggests that our method may be of use for application to other shallow targets, such as magma chambers or hydrocarbon reservoirs.

2.8 Acknowledgements

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2.9 References


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Chapter 3

Long-range correlations of microseism-band pressure fluctuations in the ocean

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3.1 Abstract

We investigate the spatial coherence of underwater ambient noise using a yearlong time series measured off Ascension Island. Qualitative agreement with observed cross-correlations is achieved using a simple range-dependent model, constrained by earlier, active tomographic studies in the area. In particular, the model correctly predicts the existence of two weakly dispersive normal modes in the microseism frequency range, with the group speed of one of the normal modes being smaller than the sound speed in water. The agreement justifies our interpretation of the peaks of the measured cross-correlation function of ambient noise as modal arrivals, with dispersion that is sensitive to crustal velocity structure. Our observations are consistent with Scholte to Moho head wave coupled propagation, with double mode conversion occurring due to the bathymetric variations between receivers. We thus demonstrate the feasibility of interrogating crustal properties using noise interferometry of moored hydrophone data at ranges in excess of 120 km.
3.2 Introduction

Wave fields generated by spatially distributed random sources are known to remain partially coherent at points separated by distances that are large compared to the wavelength, with the two-point cross-correlation function of the random wave fields approximating the Green’s function, which describes deterministic wave propagation between the two observation points [Lobkis and Weaver 2001; Snieder 2004; Roux et al. 2004; Wapenaar 2004; Godin 2006; Goudard et al. 2008]. Cross-correlation functions of pressure fluctuations in the ocean have been investigated in the 0.530 mHz band, where the correlations characterize deep-water infragravity waves and their sources [Webb 1986; Webb et al. 1991; Harmon et al. 2012; Godin et al. 2014b], and at acoustic frequencies above 1 Hz, where geoacoustic parameters of the seafloor [Brown et al. 2014], spatial [Godin et al. 2010] and temporal [Woolfe et al. 2015] variations of the sound speed in water, and ocean current velocity [Godin et al. 2014a] have been retrieved from noise cross-correlations.

Here we study one of the most energetic parts of the ambient noise spectrum, the microseism band between 0.1 Hz and 1 Hz, and investigate what information about the ocean and the seafloor can be retrieved from the cross-correlations of pressure fluctuations. This effort is motivated, in part, by successful applications of wave interferometry to the microseism-band seismic noise recorded by seismometers located on land [Sabra et al. 2005; Shapiro et al. 2005; Gerstoft et al. 2006; Bensen et al. 2007; Brooks et al. 2009] and on the seafloor [Harmon et al. 2007; Yao et al. 2011; Takeo et al. 2014; Tian et al. 2013, 2015; Zha et al. 2015]. Using methodology developed by the seismic community for extracting fundamental and higher-mode Rayleigh wave arrivals [Yao et al. 2011] from ambient noise, we obtain robust dispersion measurements from long-range correlations in the microseism band for the first time using moored hydrophone data.

This study is based on a yearlong time series of underwater ambient noise measured during 2011 off Ascension Island in the Central Atlantic. The Comprehensive Nuclear-Test-Ban Treaty Organization (CTBTO) maintains a global network of hydrophones for the passive monitoring of global nuclear testing as part of the International Monitoring System (IMS). In addition to their
Figure 3.1: (a) Ascension Island CTBTO arrays with red line connecting elements N1 and S1. S1 element coordinates are 8.9412S, 14.480W, and N1 is located at 7.8457S, 14.480W. (b) Noise power spectra from January 2011, and daily cross-correlation functions between stations N1 and S1 over year 2011 (c), showing seasonal variation in amplitudes.
primary mission, IMS hydrophone data have been previously used to characterize the ocean and ambient noise coherence at acoustic frequencies [Sabra et al. 2013; Evers et al. 2014; Evers and Snellen 2015; Woolfe et al. 2015]. The hydrophone station at Ascension consists of two triangular arrays spaced 123km apart along a line oriented roughly NE-SW (Figure 3.1a). The array element spacing is 2km, and the hydrophones are moored within the Sound Fixing And Ranging (SOFAR) channel at about 850m depth, where data are continuously recorded at a sample frequency of 250sps. Noise spectra from all Ascension elements show a prominent secondary microseism peak centered at 5s period (Figure 3.1b), and southern elements (S1-S3) exhibit higher amplitudes than their northern counterparts (N1-N3) in shorter periods of the microseism band, from 2-3.3s.

3.3 Noise Correlation in the Microseism Band

To estimate Green’s functions from the noise data we follow closely the method of Yao et al. [2011]. Data are windowed into tapered 60min segments, demeaned and detrended. Spectral whitening is applied before one-bit normalization, to suppress the influence of energetic transient signals. The preprocessed windows are then bandpass filtered in four period ranges (1-4s, 3-6s, 5-8s, and 7-11s) and cross-correlated in the time domain. The resulting cross correlation functions are filtered again in the same bands to mitigate the nonlinear effects of one-bit normalization before they are averaged to yield the estimated daily and yearly inter-array Green’s functions.

Our correlation functions exhibit prominent temporal variation in amplitudes over year 2011 (Figure 3.1d), with the greatest signal energy generally occurring during the northern hemisphere winter from January-March. These variations are consistent with the expected mechanisms of secondary microseism noise generation by nonlinear wave interaction in the north Atlantic, where a strong source region has been identified to the south of Greenland [Stehly et al. 2006; Kedar et al. 2008; Tian and Ritzwoller 2015]. Despite these amplitude variations the arrival times of the dominant peaks remain stationary in time. Once the approximate Green’s functions have been determined, we use Frequency-Time Analysis (FTAN) [Dziewonski and Hales 1972; Bensen et al. 2007] to estimate group velocity dispersion from the spectrogram of the yearly average cross
Figure 3.2: (a) Yearly average cross-correlation function for station path N1-S1, filtered in the microseism band (2-10s). (b) Spectrogram showing period-dependent group velocity (U) and Power Spectral Density (PSD) of symmeterized cross correlation function. FTAN measurements presented in Figure 3.4 were obtained from the separate positive and negative lag spectrograms (not shown).
correlation function (Figure 3.2).

We observe two primary modes that are weakly dispersive but have very distinct group speeds. Of these two dominant arrivals, one propagates at a group speed of ~1 km/s, slower than the ~1.5 km/s average speed of sound in water, and the other mode propagates faster than the speed of sound in water at ~3 km/s. Additionally, we observe what appears to be a lower-energy signal with a group speed of ~1.5 km/s that is close to the average sound speed in the water column and does not exhibit strong dispersion (Figure 3.2b).

The difference between the dispersion curves retrieved from positive and negative time delays (Figure 3.2b) provides an estimate of accuracy of our dispersion curve retrieval from the noise cross-correlations. Our forward and reverse measurements differ by ~100 m/s for the fundamental mode, and ~80 m/s for the first overtone.

3.4 Range-Dependent Dispersion Modeling

Bathymetry, and therefore the propagation conditions for seismo-acoustic waves, vary strongly along the propagation path (Figure 3.3a). This is made evident by the strong variation in higher mode cutoff periods along the N1-S1 path (Figure 3.3b) and by comparison of dispersion curves that are modeled for various ocean depths between the stations (Figure 3.3c).

To model seismo-acoustic wave propagation, we employ the adiabatic approximation [Brekhovskikh and Godin 1999] and disregard horizontal refraction. In this approximation, the modal phase and travel time in the horizontally inhomogeneous waveguide are obtained, respectively, by integration of the accumulated phase and travel time from a series of 1-D dispersion curves modeled using the bathymetric variations along the N1-S1 station path (see Appendix A).

Our bathymetric model is a hybrid of satellite altimetry [Smith and Sandwell 1997] and multibeam sonar data where available. The latter was provided by CTBTO. Water depths increase over the first 50 km to the southwest of N1 from 1 km to ~3 km, then generally remain constant for much of the central portion over the abyssal plain, until decreasing again to ~2 km on the flank of the seamount to which the southern hydrophones are moored.
Figure 3.3: (a) Bathymetric profile between stations N1 and S1 (triangles), with crosses showing locations of mode 1 cutoffs at 4s, 5.5s, 7s and 8.5s. (b) Group dispersion curves for first two modes computed using a subset of locations along the N1-S1 path. Curves are labeled with the water depth of each model and the mode number (0 or 1). Note the variation in mode 1 cutoff period with water depth. (c) Phase dispersion curves corresponding to the models shown in (b).
We use a compressional velocity ($v_p$) model for the flank of Ascension Island based on the P-wave tomography of Evangelidis et al. [2004]. In this model, $v_p$ increases from 3km/s at the seafloor to greater than 8km/s at 10km depth (typical of oceanic upper mantle). Shear velocities ($v_s$) in the crust were estimated from the $v_p$ model using a $v_p/v_s$ ratio of 1.78, which we chose based on trial values within the range $v_p/v_s = 1.740.09$ reported by Mocquet et al. [1989] for Atlantic oceanic crust younger than 50Ma. We then estimate density from $v_p$ using the empirical relation of Carlson and Raskin [1984]. Beneath the ocean and crustal layers is a half-space with shear velocity of 4.2km/s and compressional velocity of 8.0km/s, based on the Moho speed from Evangelidis et al. [2004]. The solid Earth properties are assumed to have the same dependence on the depth below seafloor at every point along the propagation path. Figure 3.4 shows model $v_p$, $v_s$, and density profiles for one water depth (3km) along the N1-S1 path. We show group velocity sensitivity kernels computed for this model in Figure 3.5. Our sensitivity kernels illustrate that mode 1 is generally sensitive to deeper structure than mode 0, and both modes' sensitivity to $v_s$ extends to greater depths than their sensitivity to $v_p$.

We calculate dispersion at 176 nodes along the interarray path using the CPS330 software package of Herrmann and Ammon [2004]. The software yields fundamental and higher mode group and phase velocity dispersion curves at each node, in a period range spanning 210s. The resulting suite of dispersion curves was then path-integrated at each period to produce a dispersion curve for the entire propagation path between stations N1 and S1. In calculating dispersion curves in the range-dependent waveguide, the phase and group speeds of the first mode are formally set to equal the shear velocity in the halfspace when the cutoff frequency of the mode is higher than the wave frequency. The physical meaning of this assumption is discussed below. Our modeling results are shown in Figure 3.6. While we did not perform an inversion in this study, the path-integrated dispersion we obtain generally agrees with our observations. Various terms have historically been used to describe guided waves that propagate within a model of a fluid layer over an elastic half-space, including Stoneley waves [Ewing et al. 1957], pseudo-Rayleigh waves [Scholte 1949; Roever et al. 1959], Rayleigh waves [Harmon et al. 2007], Rayleigh-Scholte waves [Yao et al. 2011] and
Figure 3.4: Representative P-velocity ($v_p$), S-velocity ($v_s$) and density models for a single node on the N1-S1 path. Our P-velocity model is adapted from the tomography results of Evangelidis et al. [2004].
Figure 3.5: Group velocity sensitivity kernels computed for the models shown in Figure 4, at periods of 4s and 7s. Partial derivatives of group velocity $c_g$ with respect to shear velocity $v_s$ and compressional velocity $v_p$ are shown as a function of depth below seafloor (BSF). The Moho is represented by the blue reference line at 6km BSF.
Scholte waves [Cagniard 1962; Bromirski et al. 2013]. In the strict sense, Scholte waves are surface waves that exist at the plane interface between homogeneous fluid and solid half-spaces [Roever et al. 1959; Cagniard 1962; Brekhovskikh and Godin 1998]. Here, we follow [Essen et al. 1998; Park et al. 2005; Kugler et al. 2007; Vanneste et al. 2011; and Soloway et al. 2015] and use the term Scholte waves more broadly to designate those normal modes in the fluid-solid waveguide which are strongly affected by fluid parameters and the shear rigidity of the ocean bottom.

We interpret the slower arrival (to which we refer as mode 0) as the fundamental mode of Scholte waves [Cagniard, 1962]. Unlike Scholte waves in an ocean with a constant depth, our modeling predicts weak dispersion on the N1-S1 path, in agreement with the observations (Figure 3.2b, Figure 3.6). To understand this result, note that, in the 2-D adiabatic approximation, the effective group slowness of a normal mode is a path average of the local (i.e., calculated for a given, constant ocean depth) group slownesses. In our environmental model, only ocean depth changes along the propagation path, and the path average is a (weighted) average over ocean depths. Averaging over depth is similar to averaging over frequency (as illustrated by a dispersion equation for a simplified problem in Appendix B), which suppresses the strong dispersion of local normal modes.

At shortest periods, the faster arrival can be interpreted as first Scholte wave overtone (mode 1). However, in most of the frequency band we analyze, mode 1 encounters a cutoff at one or more (typically two) points along the N1-S1 path (Figure 3.3a). We therefore interpret this fast arrival as a converted, or hybrid, wave, which propagates as a refracted, vertically-polarized shear wave along the Moho on those parts of the N1-S1 path where the first mode is cut off. In our environmental model we have a fluid layer, which overlays a stratified solid layer on a homogeneous solid half-space. At cutoff, the phase and group speeds of a normal mode equal the shear wave speed in the half-space, and theory [Brekhovskikh and Godin, 1999] predicts that the bulk of the normal energy continues propagating past the cutoff in the same direction (as opposed to being reflected) as the head wave with the shear wave speed.
Figure 3.6: Measured dispersion of the cross-correlation function of stations N1-S1 (symbols), shown with prediction from path-integrating 1-D dispersion curves computed for a range-dependent model based on the results of Evangelidis et al. [2004] (solid lines).
3.5 Discussion and Conclusions

Results of modeling the frequency-dependent travel times for the hybrid wave are in agreement with observations (Figure 3.6). Typically, we have doubly converted waves. The theory of coupling between the discrete (normal modes) and continuous (body waves) spectra of the wave field is reviewed in Brekhovskikh and Godin [1999], including double conversion in the case of an acoustic (fluid) waveguide. To our knowledge, this is the first observation of doubly converted seismo-acoustic waves in the ocean. As seen in Figure 3.6, the first overtone data are fit well with our simple range-dependent propagation model, while the fundamental mode is fit more poorly, with our model exhibiting a systematic offset of +200 m/s from the observations. This may indicate either a higher sensitivity of the fundamental mode to cross-range environmental gradients, the effects of which we have neglected, or a need to refine the $v_p$ and $v_s$ models used in our calculations.

The sensitivity of the converted wave’s travel time to shear speed in the upper mantle is of practical significance. Because of the rich frequency content of ambient noise, noise interferometry can serve as a powerful tool to reveal new propagation regimes in two- and three-dimensionally inhomogeneous ocean.

We have shown that noise interferometry in the microseism band is practical using data from hydrophones tethered ~1 km above the seafloor, at interarray distances exceeding 120 km. We interpret the two major peaks in the Green’s functions to be Scholte modes, exhibiting dispersion that is sensitive to the 2-D acoustic-elastic properties of the ocean and underlying crust along the propagation path. We find that path-integrated dispersion from a model incorporating prior tomographic results and the along-track bathymetry agrees generally with our observations. Our modeling indicates that the fundamental mode exhibits strong dependence on the $v_p$ and thickness of the water layer, while the higher mode is sensitive to velocity structure at greater relative depths in the crust and upper mantle (Figure 3.5).

Observed power spectra of pressure fluctuations in the water column have a pronounced peak in the microseism frequency band. Long (~1 km or longer) wavelengths of seismo-acoustic waves
in this band make evaluation of the noise cross-correlations insensitive to tidally induced motions of IMS moorings, which are expected to result in horizontal displacements of hydrophones in the tens of meters. The results presented here, which we obtained using hydrophones on long moorings, indicate that hydrophone-equipped autonomous underwater vehicles and floats can provide useful data for interferometric studies of the seafloor in the microseism frequency band. We have demonstrated the feasibility of inverting the dispersion of Ascension hydrophone correlations for uppermost-mantle shear velocities. Our results imply that crustal structure can be investigated using passive data from hydrophones instead of ocean bottom seismometers, assuming the hydrophones are deployed where similar coupling conditions to those at Ascension exist. Ascension Island’s tectonic origin remains in debate. The volcanic edifice is presumed to be either the surface expression of a hotspot, or the result of melt produced by the interaction of the Ascension fracture zone to the north with the nearby Mid-Atlantic Ridge to the east [Gaherty and Dunn, 2007]. It is conceivable that future work on the data investigated here can contribute to discerning between competing models of Ascension’s genesis, and a greater overall understanding of ridge-hotspot interaction processes.

3.6 Acknowledgements

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3.7 References

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3.8 Appendix 1: Guided Propagation of Seismo-Acoustic Waves in a Horizontally Inhomogeneous Ocean

Consider linear seismo-acoustic waves in an ocean where physical parameters, including the ocean depth, vary gradually in the horizontal plane. The spatial scale of the horizontal variations is assumed to be large compared to the wavelength. The ocean is stationary in the absence of waves. Let seismo-acoustic waves be generated by a monochromatic point source of mass with the amplitude \( a_0 \) of the volume injection rate. Within the water column, acoustic pressure \( p \) (i.e., wave-induced perturbation in the pressure at a given point) satisfies the reduced wave equation [Brekhovskikh and Godin, 1999]

\[
\nabla \cdot \left( \frac{\nabla p}{\rho} \right) + \frac{\omega^2 p}{\rho c^2} = i\omega a_0 \delta(R - R_1)
\]

where \( \omega \) is the wave frequency, \( \rho \) and \( c \) are the water density and sound speed, \( \delta(\cdot) \) is the Dirac delta function, \( R = (x, y, z) \), and \( R_1 = (x_1, y_1, z_1) \) is the location of the wave source.

In a horizontally invariant waveguide, the seismo-acoustic wave field can be represented as a superposition of normal modes, which propagate horizontally without coupling. Generally, horizontal inhomogeneities in a waveguide lead to energy exchange (coupling) between the modes. In the case of gradual, slow variation of the waveguide parameters with horizontal coordinates, wave propagation can be asymptotically described in the adiabatic approximation [Weinberg and Burridge, 1974; Brekhovskikh and Godin, 1999], where each mode adjusts to the varying propagation conditions without coupling to the other normal modes. Conditions of validity of the adiabatic approximation are discussed, e.g., in [Brekhovskikh and Godin, 1999]. These conditions are typically met in the ocean in the frequency range considered in the main text. In the adiabatic approxima-
tion, the acoustic pressure in a normal mode excited in a horizontally inhomogeneous waveguide by the point source in Equation 4.1 is [Weinberg and Burridge, 1974; Brekhovskikh and Godin, 1999]

\[ p(R, \omega) = i\omega a_0 P(z; r)P(z_1; r_1)G(r, r_1) \]  

where \( r = (x, y) \) and \( r_1 = (x_1, y_1) \) are 2-D horizontal vectors and the function \( G \) satisfies the 2-D Helmholtz equation

\[ \frac{\partial^2 G(r, r_1)}{\partial r^2} + k^2(r)G(r, r_1) = \delta(r - r_1) \]  

with radiation conditions at \( |r - r_1| \to \infty \). Here, \( P(z; r) \) and \( k(r) \) are the mode shape function (i.e., the vertical profile of pressure) and the mode wave number in an auxiliary, horizontally homogeneous waveguide having the same depth, seabed properties, and sound speed and density profiles that the original, horizontally inhomogeneous waveguide has at the given \( x \) and \( y \).

The function \( G \) can be viewed as the Green’s function of the 2-D Helmholtz equation (Equation 3.3), i.e., the field due to a unit point source in the 2-D problem in the horizontal plane. It satisfies the reciprocity relation \( G(R, R_1) = G(R_1, R) \). An asymptotic solution for \( G(R, R_1) \) can be found in the ray, or geometric optics, approximation [Weinberg and Burridge, 1974; Brekhovskikh and Godin 1999]. In this approximation, normal modes propagate from the wave source along horizontal trajectories (rays). The position of a point \( r(l, \phi) \) on a horizontal ray and the mode wave vector \( k(l, \phi) = (k_x, k_y) \) at this point are found from the differential ray equations

\[ \frac{dr}{dl} = k/k, \quad \frac{dk}{dl} = \partial k/\partial r, \]  

where \( l \) and \( \phi \) are the arc length along the ray and the azimuthal angle giving the direction of the ray at the source. For a generic dependence of the ocean depth on horizontal coordinates, Equations 3.4 have to be integrated numerically. In the ray approximation,

\[ G(r, r_1) = \left[ 8\pi \left( k_x \left( \frac{\partial y}{\partial \phi} \right)_l - k_y \left( \frac{\partial x}{\partial \phi} \right)_l \right) \right]^{-1/2} \exp \left( i\Phi(r, r_1) - \frac{3i\pi}{4} \right), \]  

\[ \Phi(r, r_1) = \int_{r_1}^{r} kdl. \]  

Integration in Equation 3.5 is along an eigenray, i.e., the horizontal ray that connects points \( r_1 \) and \( r \). In a horizontally homogeneous ocean, \( \partial k/\partial r = 0 \), horizontal rays are straight lines, \( x = x_1 + l \cos \phi \),
y = y_1 + l \sin \varphi$, and Equation 3.5 simplifies to $G(r, r_1) = (8 \pi k|r - r_1|)^{-1/2} \exp (i k|r - r_1| - 3i \pi/4)$, which coincides with the dominant term of the asymptotic expansion at $k|r - r_1| \to \infty$ of the exact solution $G(r, r_1) = -0.25i H_0^{(1)}(k|r - r_1|)$ of Equation 3.3.

When $\nabla k$ has a constant direction and a normal mode propagates in this direction, i.e., $(r - r_1) \cdot \nabla k \equiv 0$, horizontal rays (Eq. 3.4) are again straight lines that connect points $r_1$ and $r$. In underwater acoustics, waveguides with this type of horizontal inhomogeneity are referred to as range-dependent ones. Choosing the $Ox$ coordinate axis in the direction from $r_1$ to $r$, from Equation 3.5 one finds

$$\Phi(r, r_1) = \int_{x_1}^{x} k \, dx$$

Note that geometry of the horizontal rays is independent of wave frequency in this case. Equation 3.6 for the mode phase is often applied to generic horizontally inhomogeneous waveguides and provides then an approximation to the true mode phase. In agreement with Fermat’s principle, the difference between the true and approximate mode phase is of second order in the cross-range gradients of $k$, i.e., in $\partial k/\partial y$. For a discussion of conditions of validity of the approximation (Equation 3.6) in generic horizontally inhomogeneous waveguides see Godin [2002].

For the mode travel time $t(r, r_1)$ in a range-dependent waveguide, from Equation 3.6 we find

$$t(r, r_1) = \frac{\partial \Phi(r, r_1)}{\partial \omega} = \int_{x_1}^{x} \frac{dx}{c_g}$$

where $c_g = (\partial \omega/\partial k)_x$ is the group velocity of the mode in a corresponding horizontally homogeneous waveguide. Equation 3.7 has been used in the main text to calculate modal travel times and the modeled effective group speeds $U$ shown in Figure 3.4. Equation 3.7 shows that the effective group slowness, $U^{-1} = t(r, r_1)/|r - r_1|$, in a range-dependent waveguide is an average of the modal slowness $c_g^{-1}$ over range. When changes in $k$ are due to changes in the ocean depth $H$, and the depth varies steadily with range, the group slowness becomes a weighted average of $c_g^{-1}$ over depth:

$$\frac{1}{U} = \frac{1}{H(x) - H(x_1)} \int_{H(x_1)}^{H(x)} \frac{dH}{c_g}, \quad \alpha = \frac{H(x) - H(x_1)}{(x - x_1)dH/dx}$$

Equation 3.8
As discussed in Appendix 2, Equation [3.8] helps to understand the striking difference in the frequency dependencies of $U$ (see Figure [3.4] in the main text) and $c_g$ (see Figure [3.3b] in the main text).

3.9 Appendix 2: Dispersion equation of Scholte waves in a benchmark problem

Consider a homogeneous water layer of depth $H$ overlying a homogeneous solid half-space. The ratio of densities of the solid and the fluid is $M$. Wave frequency is $\omega$; sound speed in water, and speeds of compressional and shear waves in the bottom are $c$, $v_p$, and $v_s$ respectively.

The dispersion equation for such a waveguide is obtained by requiring that a monochromatic wave with a horizontal wavenumber $k$ satisfies corresponding reduced wave equations in the fluid and the solid, conditions at infinity (i.e., only evanescent waves are allowed in the solid half-space), the boundary condition of zero acoustic pressure at the free upper surface of the water layer and the boundary condition at the fluid-solid interface. Three boundary conditions should be met at the fluid-solid interface: the normal (vertical) displacement is continuous; the normal (vertical) component of the traction vector is continuous; tangential (horizontal) components of the traction vector equal zero [Brekhovskikh and Godin, 1998].

The resulting dispersion equation [Ewing et al., 1957] can be written as follows:

$$\tan\left(\omega H \sqrt{c^{-2} - v^{-2}}\right) = \frac{M (v_s/v)^4}{\sqrt{v^{-2} - v_p^{-2}}} \left[4\sqrt{\left(1 - \frac{v^2}{v_p^2}\right)\left(1 - \frac{v^2}{v_s^2}\right) - \left(2 - \frac{v^2}{v_s^2}\right)^2}\right]$$

(3.9)

Here $v \equiv \omega/k$ is the phase speed of the normal mode. In all normal modes, $c \leq v \leq v_s$. Therefore, the right-hand side of Equation [3.9] is always real. When $v < c$ it is convenient to write the left-hand side of Equation [3.9] as $(v^{-2} - c^{-2})^{-1/2} \tanh(\omega H \sqrt{v^{-2} - c^{-2}})$. Note that the expression in the square brackets in the right-hand side of Equation [3.9] is the same as appears in the dispersion equation for the Rayleigh surface wave in a solid half-space with a free boundary [Ewing et al., 1957; Brekhovskikh and Godin, 1998]. It is positive when $v < c_R$ and negative when $v > c_R$, where $c_R$ is the Rayleigh surface wave speed in a solid half-space with a free boundary.
Every continuous, real-valued solution \( v(\omega) \) of the dispersion equation (3.9) determines a dispersion curve of a particular normal mode. Surface waves supported by a fluid-solid interface and, more generally, normal modes in a fluid layer over a solid half-space are often referred to as Scholte-Stoneley or Scholte waves [Ewing et al., 1957; Brekhovskikh and Godin 1998].

In four special cases: (i) \( M \to 0 \), (ii) \( M \to +\infty \), (iii) \( H \to 0 \), and (iv) \( c_s \to 0 \), the dispersion equation (B.1) reduces to well-known elementary dispersion equations [Ewing et al., 1957; Brekhovskikh and Godin, 1998] for (i) a fluid layer with two free boundaries, (ii) a fluid layer with one free and one rigid boundary, (iii) Rayleigh surface wave in a solid half-space with a free boundary, and (iv) the Pekeris waveguide, i.e., a waveguide with a homogeneous fluid bottom. (In the case of a fluid bottom, there are no shear waves to carry energy to infinity, and the condition \( 0 \leq v \leq v_s \) no longer applies. It is replaced by \( 0 \leq v \leq v_p \).)

One can easily solve Equation (3.9) explicitly for the product \( \omega H \) as a function of \( v \) and parameters \( M, v_p, v_s, \) and \( c \):

\[
\omega_0 H = \frac{1}{\sqrt{v^2 - c^2}} \arctanh \left( \frac{M(v_s/v)^4 \sqrt{v^2 - c^2}}{v^2 - v_p^2} \right) \left( \frac{1}{4} \left( 1 - \frac{v_p^2}{v_s^2} \right) \left( 1 - \frac{v^2}{v_s^2} \right) - \left( 2 - \frac{v^2}{v_s^2} \right)^2 \right), \quad 0 < v \leq \min(c, c_R);
\]

\[
\omega_0 H = \frac{1}{\sqrt{c^2 - v^2}} \arctan \left( \frac{M(v_s/v)^4 \sqrt{c^2 - v^2}}{v^2 - v_p^2} \right) \left( \frac{1}{4} \left( 1 - \frac{v_p^2}{v_s^2} \right) \left( 1 - \frac{v^2}{v_s^2} \right) - \left( 2 - \frac{v^2}{v_s^2} \right)^2 \right), \quad c \leq v \leq c_R;
\]

\[
\omega_n H = \frac{1}{\sqrt{c^2 - v^2}} \left( \pi n - \arctan \left( \frac{M(v_s/v)^4 \sqrt{c^2 - v^2}}{v^2 - v_p^2} \right) \right) \left( \frac{1}{4} \left( 1 - \frac{v_p^2}{v_s^2} \right) \left( 1 - \frac{v^2}{v_s^2} \right) - \left( 2 - \frac{v^2}{v_s^2} \right)^2 \right), \quad c \leq v \leq v_s,
\]

where \( n = 1, 2, \ldots \). Equations (3.10) and (3.11) give the frequency of the fundamental mode with the phase speed \( v \). The branch of its dispersion curve, which is described by Equation (3.10).
always exist; Equation [3.11] describes an additional branch that exists provided \( c < c_R \). The phase speed of the fundamental mode satisfies the inequality \( v < c_R \).

Equation [3.12] gives the frequencies of the higher-order modes with the phase speed \( v \). Only the fundamental mode (Equation [3.10]) exists when \( c \geq v_s \). When \( c < v_s \), the \( n \)-th mode exists at frequencies \( \omega \geq \Omega_n \) where the cutoff frequency

\[
\Omega_n = \frac{1}{H \sqrt{c^{-2} - v_s^{-2}}} \left( \pi n - \frac{1}{v} \arctan \left( \frac{M \sqrt{c^{-2} - v_s^{-2}}}{\sqrt{v_s^{-2} - v_p^{-2}}} \right) \right).
\]

(3.13)

Note that the cutoff frequency increases with decreasing ocean depth.

An explicit equation for the modal group speed

\[
c_g = \frac{\partial \omega}{\partial k} = \left( 1 - \frac{\omega}{v} \frac{\partial v}{\partial \omega} \right)^{-1} v
\]

(3.14)

is readily obtained by differentiating both sides of Equation [3.9] with respect to \( \omega \). The equation for \( c_g \) is cumbersome and will not be reproduced here. Figure [3.7] shows group and phase dispersion curves modeled for the benchmark problem using the software of Herrmann and Ammon [2004].

We calculate dispersion for a model with density ratio \( M = 3.0 \), \( c = 1.5\text{km/s} \), \( v_p = 7.79\text{km/s} \) and \( v_s = 4.5\text{km/s} \). Dependencies of the modal phase and group velocities on wave period \( 2\pi/\omega \) in the benchmark problem, which follow from Equation [3.9] are qualitatively similar to those shown in Figures [3.3] in the main text.

It follows from Equations [3.9] and [3.14] that the modal phase and group velocities are functions of the product \( \omega H \) when the parameters \( M \), \( c \), \( v_p \), and \( v_s \) are kept constant. (\( v \) and \( c_g \) retain this property also in other waveguides as long as the acoustic impedance [Brekhovskikh and Godin, 1998] of the boundary \( z = H \) is a function of \( v \) and is independent of frequency.) For any function of \( \omega H \), averaging over frequency is equivalent to averaging over ocean depth. Indeed, the average over frequency

\[
\frac{1}{\omega_2 - \omega_1} \int_{\omega_1}^{\omega_2} f(\omega H) d\omega = \frac{1}{\omega_2 H - \omega_1 H} \int_{\omega_1 H}^{\omega_2 H} f(a) da
\]

(3.15)
equals the average over depth

\[
\frac{1}{H_2 - H_1} \int_{H_1}^{H_2} f(H) dH = \frac{1}{\omega H_2 - \omega H_1} \int_{\omega H_1}^{\omega H_2} f(a) da
\]

(3.16)
Figure 3.7: (a) Group dispersion curves for first two modes computed for the benchmark problem with water layer thicknesses of 1-3km. Curves are labeled with the water depth of each model and the mode number (0 or 1). (b) Phase dispersion curves corresponding to the models shown in (a). Halfspace compressional and shear velocities are 7.79km/s and 4.5km/s respectively, and the density ratio of the fluid layer to the halfspace layer is $M=3.0$. 
as long as $\omega_1 H = \omega H_1$ and $\omega_2 H = \omega H_2$.

As discussed in Appendix 1, mode travel time in a waveguide with range-dependent bathymetry is proportional to an average of $c_g^{-1}$ over ocean depth $H$. In the benchmark problem we consider, averaging over $H$ at fixed $\omega$ is equivalent to averaging over $\omega$ at fixed $H$ and results in a suppression of the frequency dependence of the modal travel time.

### 3.9.1 References


Chapter 4

Lithospheric shear velocity structure of South Island New Zealand from amphibious Rayleigh wave tomography

as submitted to Journal of Geophysical Research:


4.1 Abstract

We present the first 3D shear velocity model extending well offshore of New Zealand’s South Island, imaging the lithosphere beneath the South Island as well as the Campbell and Challenger plateaus. Our model is constructed via linearized inversion of both teleseismic (18-70 s period) and ambient noise-based (8-25 s period) Rayleigh wave dispersion measurements. We augment an array of 4 land-based and 29 ocean-bottom instruments deployed off the South Islands east and west coasts in 2009-2010 by the Marine Observations of Anisotropy Near Aotearoa (MOANA) experiment with 28 land-based seismometers from New Zealand’s permanent GeoNet array. Major features of our shear wave velocity (Vs) model include a low-velocity (Vs<4.4km/s) body extending from near surface to greater than 75km depth beneath the Banks and Otago peninsulas, and high-velocity (Vs~4.7km/s) mantle anomalies underlying the Southern Alps and off the northwest coast of the South Island. Using the 4.5km/s contour as a proxy for the lithosphere-asthenosphere boundary, our model suggests that the lithospheric thickness of Challenger Plateau and central South Island
is substantially greater than that of the inner Campbell Plateau. The high-velocity anomaly we resolve at sub-crustal depths (>50km) beneath the central South Island exhibits strong spatial correlation with upper-mantle earthquake hypocenters beneath the Alpine Fault [Boese et al., 2013]. The ~400km-long low velocity zone we image beneath eastern South Island and the inner Bounty Trough underlies Cenozoic volcanics and the locations of mantle-derived helium measurements [Hoke et al., 2000], consistent with asthenospheric upwelling in the region.

4.2 Introduction

The South Island of New Zealand sits at the juxtaposition of three different tectonic environments. To the north, the Pacific plate subducts beneath the Australian plate, with a well-defined Benioff zone and abundant deep and intermediate depth seismicity [Kohler et al., 2003]. In Fiordland to the south, the sense of subduction is reversed, with the Australian plate subducting beneath the Pacific plate. Connecting these regions along the west coast of the South Island, the right-lateral Alpine fault is the surface expression of a continental transform plate boundary. The islands of present-day New Zealand are the subaerial portion of a much larger submerged continental fragment, Zealandia, which was part of the Gondwana supercontinent until the opening of the Tasman Sea in the late Cretaceous [Bache et al., 2014; Cande and Stock, 2014]. The oldest rocks found in New Zealand are from the early-middle Paleozoic Buller (quartz-rich clastics of continental affinity) and Takaka (oceanic arc assemblage) terranes of the Western Province [Bradshaw et al., 2009; Cooper and Tulloch, 1992]. These units are inferred to record complex accretionary processes that occurred along the southeast Gondwana Margin during the middle-late Cambrian Ross Orogeny [Federico et al., 2009], and have been correlated with rocks of the Lachlan Fold Belt in Eastern Australia [Bradshaw et al., 2009; Cooper and Tulloch, 1992]. Rocks of the Western Province are separated from those of the younger Eastern Province by a tectonic boundary, delineated on the surface by the Junction Magnetic Anomaly (JMA) [Sutherland et al., 1999]. Prior to the breakup of Gondwana, Zealandia was likely located seaward of Antarctica and Australia and inboard of the Pacific-Phoenix plate boundary, with the Campbell plateau contiguous with West Antarctica,
and the Challenger Plateau separated from the Lachlan Fold Belt by the Lord Howe Rise [Bache et al., 2014]. Convergent conditions mostly dominated the SE Gondwana margin from the middle Cambrian through the early Cretaceous, and seismically-imaged northeast dipping crustal/mantle reflectors suggest that relict structures from this long-lived convergent episode are preserved within the Challenger Plateau lithosphere today [Davey, 2005; Melhuish et al., 2005]. The Hikurangi oceanic plateau subducted beneath the Gondwana margin at the present-day Chatham rise circa 105Ma, and jammed the subduction margin by 100Ma [Davy et al., 2008]. The breakup of Gondwana initiated with widespread continental rifting and extension beginning 100Ma, whereafter Zealandia transitioned to a passive margin setting when Tasman Sea spreading began circa 85Ma [Bache et al., 2014]. The opening of the Tasman Sea ceased around 52Ma when the Lord Howe Rise and Australian plates joined. Plate reconstructions indicate that the Zealandian Australia/Pacific plate boundary likely initiated at this time, cutting across the Challenger Plateau [Cande and Stock, 2004], possibly in the location of the present-day Alpine fault [Sutherland et al., 2000]. Eocene rifting along the Resolution Ridge beginning circa 45Ma likely created oceanic crust eastward of the present-day Alpine Fault, which is no longer present in surface rocks, and is inferred to have been overridden by the Campbell Plateau during Cenozoic convergence [Spasojevic and Clayton, 2008; Sutherland et al., 2000]. The Eocene rifting episode requires a pair of passive margins, one of which is evidently preserved at the western edge of the Campbell Plateau. Plate reconstructions suggest that the conjugate margin, and possibly the missing Eocene oceanic lithosphere, have been incorporated into the continental collision zone beneath the Alpine Fault since the Miocene [Sutherland et al., 2000]. The present-day Alpine Fault initiated with slow, diffuse transtension around 45Ma, which evolved into almost purely strike-slip motion by 25Ma [Sutherland et al., 2000].

Sutherland et al. [2000] propose that the evolution and morphology of the present-day plate boundary could be controlled by lithospheric discontinuities extant since the Paleozoic and manifested in the location of the Eocene passive margin, while Reyners [2013] proposes that the continental collision process since the Cenozoic has been influenced by the ongoing subduction of the Hikurangi oceanic plateau.
Much of the continental lithosphere of Zealandia is below sea level and not easily accessible for seismic studies to probe the nature of the plate boundary and adjacent mantle lithosphere. Previous work using New Zealand land-based seismic stations has produced images of crustal structure from ambient noise surface tomography [Lin et al., 2007], Moho depth from receiver functions [Spasojevic and Clayton, 2008], and body wave tomography for crust and mantle structure [Kohler and Eberhart-Phillips, 2002]. Teleseismic shear wave splitting analysis [Zietlow et al., 2014] and Pn traveltime residuals [Collins and Molnar, 2014] reveal a pattern of lithospheric anisotropy consistent with strain distributed across a 200km region parallel to the Alpine fault, but uncertainty persists as to the depth extent of localized vs. distributed shear as revealed by seismic anisotropy. The latter two studies utilized an array of 29 broadband ocean bottom seismometers (OBS) deployed from 2009-2010 in the Marine Observations of ANisotropy off Aoteroa MOANA experiment [Yang et al., 2012], with the primary objective of using these anisotropy measurements to characterize the distribution of strain about the plate boundary.

In this study we utilize ocean bottom seismometers offshore both east and west coasts of the South Island of New Zealand, and along with existing land-based seismometers, perform surface wave tomography for Rayleigh wave group and phase velocities using ambient noise and earthquake data recorded at periods from 8-70 s. The incorporation of ocean bottom seismometers greatly increases the array aperture available to interrogate both the onshore and offshore portions of the New Zealand continental lithosphere, resulting in improved resolution for the land portion of the study area, and the first lithospheric shear velocity models of offshore Campbell and Challenger plateaus.

4.3 Data and Methods

Ambient noise and teleseismic surface wave analysis is conducted using waveforms collected by the 29 MOANA OBS stations and 28 land based stations, 24 of which are from the permanent New Zealand GeoNet network [Petersen et al., 2011] shown in Figure 4.1. The OBS stations were deployed between February 2009 and February 2010 over a 900x1300km area centered on South
Figure 4.1: (a) Station map showing location of OBS and land seismic stations. Color scale represents topography and bathymetry relative to sea level. (b) Regional map showing main tectonic features. Feature labeled J.M.A. is the trace of the Junction Magnetic Anomaly after Sutherland, [1999].
Island. The average station spacing was 100km. The OBS instrument packages included three-component seismometers with 240s lower corner frequencies (excepting one 40s sensor), and Cox-Webb Differential Pressure Gauges (DPG) with long-period sensitivity extending to over 1000s [Webb and Crawford, 1999]. Data were digitized continuously at a sample frequency of 50Hz. Seismic data quality on the OBS vertical component can approach or exceed that of quiet land stations in the noise notch [Webb and Crawford, 1999] at periods between 10-40s. MOANA OBS station performance is described in more detail in Yang et al. [2012].

4.3.1 Ambient noise cross-correlation

Interstation group and phase velocity dispersion measurements are obtained from ambient noise cross-correlations calculated between all station pairs for every hour of data between February 2009 and February 2010 using the method of Stachnik et al. [2008]. We downsample the continuous vertical component seismic waveforms from 50sps to 1sps, and preprocess by removing the linear trend and instrument responses before bandpass filtering from 3-100 seconds to mitigate artifacts during cross-correlation. The waveforms are then Fourier transformed, the amplitude spectrum is pre-whitened and the cross-correlation is performed in the frequency domain. This method is similar to time domain one-bit normalization techniques in its effectiveness at suppressing transient signal contamination [Bensen et al., 2007]. Information about absolute amplitudes is lost with both of these methods, however it is not needed to calculate interstation group and phase velocity, and the amplitude normalizing techniques allow data to be utilized and stacked in the absence of source information. The hourly cross-correlations for a given station pair are stacked to give a day record and archived in a database. All available day records for each station pair are then stacked to yield an experiment average cross correlation function for each station pair. Examples of stacked interstation cross-correlations are shown in Figure 4.2 for different combinations of OBS stations and land-based stations. Clear Rayleigh wave packets are visible with propagation speeds expected of surface waves. Cross-correlations for ocean-ocean and ocean-land paths all show clear and symmetric Rayleigh wave packets for both the causal and acausal signal implying good
Figure 4.2: Yearly stacks (waveforms at top) of 24-hour ambient noise Greens Functions used in this analysis. Cross-correlations between selected OBS stations in the Tasman Sea (a), land stations (b), and OBS in the Bounty Trough (c) are shown.
azimuthal distribution of ambient noise sources (Figure 4.2). Rayleigh waves are consistent over the duration of the experiment, which encompassed both austral summer and winter cycles, indicating minimal temporal seasonal bias. Cross-correlations between OBS stations on opposite sides of the island and those between land and OBS stations are less symmetric, yet still reveal good signal to noise (SNR) and dispersive characteristics. Fundamental mode Rayleigh wave dispersion is measured for each station pair from the stacked cross correlation using frequency-time analysis [Dziewonski and Hales, 1972]. Similar to Lin et al. [2008], the amplitudes of the positive and negative lag times of the cross correlations are averaged to extract a Rayleigh wave packet unbiased by signal strength related to direction of propagation and source distribution. The period band of reliable dispersion measurements is determined by calculating signal to noise ratio of the Rayleigh wave envelope relative to the root mean square of cross correlation amplitudes more than 50 seconds past the timing of an arrival with velocity 2.0 km/s. We measure group and phase velocity dispersion at periods from 8-25s from the interstation Greens functions using the Frequency-Time Analysis (FTAN) method of Levshin et al. [1972]. We employ the automated FTAN software aftan 1.1 (http://ciei.colorado.edu/Products/, last accessed May 19, 2014). The group velocity is determined by applying a sequence of narrow band Gaussian filters to the cross correlation and selecting the peak of the envelope within a reasonable group velocity window (1.4-4.5 km/s). The dispersion curves are the mean group velocity calculated at periods from 8 to 25 s after culling based on group velocity standard deviation less than 0.10 km/s and SNR greater than 10. Subsequently we remove all measurements exceeding 3 standard deviations from the mean, and those where inter-station distance is less than 2 wavelengths, calculated given the velocity and period. We use the 2-wavelength criterion following [Yao et al. 2011; Lin et al. 2013] to preserve more measurements at long periods, given the relatively short inter-station distances in the MOANA array. Cumulatively, these culling steps remove about 20 percent of our measurements. We use these interstation dispersion measurements as input to a tomographic inversion for noise-based group and phase velocity maps (Figure 4.3) to complement our longer-period teleseismic phase velocity maps produced using Eikonal tomography (Figure 4.4).
Figure 4.3: (a.) Isotropic group velocity maps derived from ambient noise tomography at periods of 8-25s. (b.) Phase velocity maps derived from ambient noise tomography at periods of 8-24s. (c.) Phase velocity maps derived from teleseismic data (left) and ambient noise tomography (right) at 24s period.
Figure 4.4: Phase velocity maps from teleseismic tomography at periods of 40-70s.
4.3.2 Ambient noise group velocity tomography

To transform the culled path-dependent inter-station dispersion measurements into a set of period-dependent 2D group velocity maps we employ the straight-ray tomographic inversion method of Barmin et al., [2001]. We use the software package tomo_sp_cu_8 1.1 (http://iei.colorado.edu/Products, last accessed May 19, 2014). This software performs a damped least-squares inversion, producing isotropic and anisotropic 2D group and phase velocity maps at periods from 6-30s.

Figure 4.5: Resolution maps from straight-ray ambient noise group velocity tomography procedure at periods from 8-25s.

While we analyze only the resulting isotropic group velocity maps here (Figure 4.3), the anisotropic maps are interpreted in Yeck [2015]. We also generate maps of azimuthal coverage, ray density and resolution (Figure 4.5). The spatial resolution maps (Figure 4.5) are a measure of the minimum distance from each node at which a delta-shaped anomaly can be resolved from another at the target node by the tomographic inversion. In accordance with prior applications of this tomographic method [Lin et al., 2008], resolution is finest in the center of the array (where the largest number of interstation paths cross), and decreases in quality toward the edges of the array. The tomography produces maps with a node spacing of 0.5 degrees, which we then interpolate to
Figure 4.6: (a.) RMS uncertainty maps for ambient noise-derived phase velocity measurements at periods from 8-24s. (b.) Maps of RMS uncertainty of teleseismic phase velocity measurements at periods from 24-70s.
resample the maps at 0.2 degrees node spacing. The resulting group speed maps are shown in Figure 4.3.

### 4.3.3 Ambient noise and teleseismic phase velocity tomography

At a given period, Rayleigh wave phase velocity dispersion generally depends on deeper velocity structure than group dispersion [Yao et al., 2011]. The addition of teleseismic (earthquake) dispersion measurements enables longer periods, and thus deeper lithospheric shear velocities, to be modeled [Gaite et al., 2015] compared with noise-based measurements alone. Therefore, to complement our group velocity measurements from straight-ray tomography and resolve deeper structure, we also measure phase velocities from both noise cross-correlation functions (8-24s period) and teleseismic Rayleigh wave signals (18-70s period). All available vertical component waveforms for earthquakes with $M_s > 5.0$ are analyzed. High noise levels are commonly observed at shallow OBS stations at long periods (> 30 sec), which necessitates special treatment before these data can be further analyzed. This long period noise is clearly correlated with pressure perturbations observed by the co-located differential pressure gauges (DPG) and is likely caused by seafloor compliance, or the deformation of the seafloor under loading by ocean surface infragravity waves. Seafloor compliance is defined by the transfer function between displacement and pressure at the seafloor [Crawford and Webb, 2000], and is sensitive to sediment rigidity and thickness [Ball et al., 2014]. To reduce the compliance noise level, we implement the method described by Crawford and Webb [2000]. For a quiet day (June 9, 2009) with no detectable earthquakes in the observed waveforms, we calculate the transfer function between the vertical component OBS and DPG measurements for each station. The transfer function is then used to predict and remove the compliance signal in the vertical component due to pressure perturbation for any other days. We follow closely the eikonal tomography method described by Lin et al. [2009] to determine Rayleigh wave phase velocity maps across the array for the noise correlation functions and teleseismic data. For each station and earthquake, and each interstation noise correlation function, we first perform frequency time analysis (FTAN; Levshin et al. [1972]) to obtain dispersive phase travel time
measurements. For each earthquake and period, all measurements with SNR higher than 8 are then used to determine the phase travel time map. Note that we follow the phase front tracking method described by Lin & Ritzwoller [2011] to resolve the 2 ambiguity in phase velocity measurements and we determine the travel time maps on a 0.2 by 0.2 grid through the minimum curvature surface fitting [Smith and Wessel, 1990]. Based on the eikonal equation, we estimate the phase velocity at each location using the gradient of each phase travel time map. For each period, all phase velocity measurements at the same location are averaged to obtain the final isotropic phase speed maps for both noise and teleseismic data. The resulting noise-based and teleseismic phase speed maps are shown in Figures 4.3 and 4.4. A comparison of noise and teleseismic phase speed measurements at a common period (24s) is shown in Figure 4.3. While the noise and teleseismic datasets data generally agree well, subtle differences are apparent, likely resulting from an inhomogeneous noise source distribution and/or differing finite-frequency sensitivities between datasets [Yao et al., 2009]. Period-dependent measurement uncertainties are estimated following Lin et al. [2009] and uncertainty maps are shown in Figure 4.6. For both noise-based and teleseismic datasets, measurement uncertainty is lowest in the center of the analyzed period ranges (Figure 4.6).

4.3.4 Inversion for 1D Shear Velocity Profiles

We sample the Rayleigh wave group and phase velocity tomograms at 0.2 degree increments in latitude and longitude to determine local dispersion curves at each node. At each node, the local dispersion curves are then simultaneously inverted for a one dimensional isotropic shear velocity (Vs) profile using depth sensitivity kernels and the linearized least-squares method of Herrmann and Ammon [2002]. The software performs an iterative, damped inversion, computing updated sensitivity kernels at each iteration to minimize the dependence on the starting model. We fix the Vp/Vs ratio in the medium to that of the initial model and iteratively invert for the Vs of each layer. Density is calculated from the new Vp at each iteration using the Nafe-Drake equation. The inversion minimizes the following objective function [Aster 2013]:

$$ F = \| J(m)(m + \Delta m) - (d - G(m) + J(m)m) \|^2_2 + \lambda^2 \| L(m + \Delta m) \|^2_2 $$

(4.1)
where $\mathbf{m}$ is a vector of unknown model parameters perturbed by $\Delta \mathbf{m}$, $\mathbf{d}$ is the observed data vector, and $\mathbf{J}(\mathbf{m})$ is the Jacobian matrix containing partial derivatives of the forward equation $\mathbf{G}(\mathbf{m})$ with respect to the model parameters (in this case $\mathbf{J}(\mathbf{m})$ contains partial derivatives of group and phase velocity with respect to shear velocity in each layer). The finite difference operator $\mathbf{L}(\mathbf{m}+\Delta \mathbf{m})$ includes a matrix of weights controlling model smoothness by limiting the change in velocity across each layer [Herrmann and Ammon, 2002], and $\lambda$ is the damping (regularization) parameter. We use a high damping parameter (10) in the first iteration to mitigate model overshoot effects, and lower the damping to 0.3 for subsequent iterations. We chose the global damping value and number of iterations (10) via trial and error to optimize the tradeoff between model roughness and data misfit. We apply moderate weighting (0.8) to the layers in the uppermost 20km to smooth the upper-crustal model. Below 20km individual layer weighting is decreased with depth, so that the deepest layers (exceeding the peak depth sensitivity of our dispersion data) remain unperturbed from the reference model. The starting model for the Vs inversion is based on the ak135 reference Earth model of Kennett et al., [1995]. The velocities of the uppermost 50 km of the starting model are constant, set to the 40km-depth ak135 values of $V_p=8.04$ km/s and $V_s=4.48$ km/s, in order to eliminate bias from the ak135 models Moho depth. Below 50km the starting model follows ak135 velocities with depth. We parameterize the model using 2 km thick layers in the upper 50 km, increasing to 5 km thick layers down to 100 km. Below 100 km, layers are 10 km thick. The increasing layer thicknesses with depth were chosen by trial and error to optimally match the depth-sensitivity of our dispersion data. In offshore regions the local starting model is capped by a layer of zero shear velocity with thickness equal to the water depth taken from the SRTM30_PLUS global elevation and bathymetry model [Becker et al., 2009]. The velocity and thickness of this water layer is fixed during the iterated linearized inversion.

We perform the inversion twice, once with no a-priori crustal thickness estimates, and once incorporating estimates of Moho depth from Salmon et al. [2013]. In the constrained-Moho inversion we identify the model layer corresponding to the estimated Moho depth, and force a velocity contrast to exist across that layer by increasing the layers weighting by a factor of 5. The crustal
Figure 4.7: Data fits (top) and 1D models (bottom) resulting from inversions with (left column) and without (right column) Moho constraint from Salmon et al. [2013]. Blue and orange lines show PEM-O and PEM-C reference Earth models respectively, from Dziewonski et al., [1975].
thickness model of Salmon et al., [2013] incorporates results from onshore/offshore refraction and reflection studies dating to the 1980s, and includes Moho depth estimates from recent onshore receiver function analyses [Spasojevic and Clayton, 2008]. Where data are sparse (i.e. offshore), they are interpolated using the Crust2.0 model [Bassin et al., 2000]. This model is very well-constrained by dense datasets in some areas (onshore, and offshore areas with active source lines) and poorly-constrained in others (offshore away from active source lines). Representative 1D velocity models with and without Moho constraint are shown in Figure 4.7. In the models shown, imposing the Moho constraint tends to reduce both data misfit at longer periods and the relative variance of deeper velocities between nodes.

4.4 Shear Velocity Inversion Results

Major features of our 3D shear velocity model that we interpret here (Figures 4.8-4.9) include a southward-dipping high-speed ($V_s > 4.5$ km/s) body underlying the South Island south of Mt. Cook, a lower-velocity ($V_s < 4.4$ km/s) body underlying the Canterbury/Otago region and Bounty Trough, and deep (>100 km) high-velocity ($V_s > 4.5$ km/s) zones extending offshore the northwestern coast of the South Island beneath Challenger Plateau and southward beneath the west coast of the South Island. Depth slices at 15-140 km are shown in Figure 4.8. Figure 4.9 shows $V_s$ model cross sections perpendicular to the strike of the Alpine fault and extending from the Tasman Sea across the South Island to the Bounty Trough, and fault-parallel transects spanning the central South Island and the Challenger and Campbell Plateaus. At shallow depths (15 km) in our model, the most prominent feature we resolve is an arcuate lateral velocity discontinuity cutting across the southern South Island (Figure 4.8a-b), which generally follows the trace of the Junction Magnetic Anomaly (Figure 4.1). To the north of this zone, crustal shear velocities are 3.4 km/s at 15 km depth, while to the south they exceed 3.6 km/s. At a depth of 40 km (Figure 4.8b-c) we resolve an apparent crustal root with shear velocities less than 3.8 km/s underlying the Southern Alps, in contrast with surrounding higher shear velocities (> 4 km/s) characteristic of upper mantle material. The imposition of a Moho constraint focuses this crustal root beneath
Figure 4.8: Depth slices through final shear velocity model, from $z=15\text{km}$ to $z=140\text{km}$. Maps labeled "Moho" result from inversion using Moho depth constraint from Salmon et al., [2013]. Maps labeled "No Moho" result from unconstrained inversion.
Figure 4.9: Cross-sections through final Vs models. Top cross-section in each subplot shows model resulting from inversion with Moho depth constraint of Salmon et al. [2013], bottom section shows unconstrained inversion result. Transect locations are shown on maps in right column, overlain on the 80km unconstrained depth slice shown in Figure 8e. Red stars on sections a, b, c, and e are subcrustal earthquake hypocenters presented in Boese et al., [2013]. Blue stars are Low Frequency Earthquake (LFE) hypocenters reported by Chamberlain et al., [2014].
the highest topography of the Southern Alps in the 40km depth slice (Figure 4.8c). The central South Island high-velocity anomaly in our model at depths below 50km dips southward from Mt. Cook, underlies the Southern Alps and extends westward of the Alpine fault beneath Challenger Plateau (Figure 4.9c). The depth of highest shear velocities beneath central South Island ranges from ~80km beneath Mt. Cook to over 100km beneath Fiordland, and west of the Alpine fault the anomaly extends to over 150km depth (Figure 4.9b). The northern, offshore Challenger Plateau high-velocity anomaly we image extends several hundred kilometers northwestward from the west coast, with the highest shear speeds concentrated at depths below 80km and latitudes below 43S (Figure 4.8a-h). We also resolve a low-velocity zone at depths from 60-150km beneath the East Coast of South Island which underlies the Otago Peninsula at the shallowest depths (Figure 4.8at) and, with increasing depth, increases in lateral extent beneath the Canterbury Basin (Figure 4.8at-h), underlying the entire inner Bounty Trough at 150km depth.

4.5 Discussion

4.5.1 Challenger Plateau

We image a zone of high shear velocities ($V_s > 4.6\text{km/s}$) offshore the northwestern coast of South Island and extending southward along the west coast (Figure 4.8a-h), with the highest shear speeds extending in depth from 50-150km beneath the Challenger Plateau. This feature in our model is also resolved to greater depths ($>300\text{km}$) with teleseismic body wave tomography by Zietlow et al., [2015], who interpret it as subducted Pacific plate lithosphere. While observed Benioff zone seismicity associated with modern Hikurangi subduction [Reyners, 2013] does not extend nearly as far westward of the Alpine fault as the high-velocity feature we model at depth, we too consider westward-subducted lithosphere to be the simplest explanation for this high-wavespeed anomaly. The continental lithosphere of the Challenger Plateau is at least Paleozoic in age [Sutherland et al. 2000], and anomalous mantle reflectors that have been interpreted as relict structures from Gondwana margin subduction [Davey et al., 2005; Fry et al., 2014; Melhuish et al., 2005]
coincide with the southwest boundary of the Challenger Plateau high velocity anomaly we image. A prominent magnetic discontinuity, separating high-amplitude anomalies on western Challenger Plateau from lower amplitudes to the east, may delineate a tectonic boundary within the plateau [Sutherland, 1999], and coincides with the westward boundaries of the slab-like high-speed anomaly we image offshore northwestern South Island (Figure 4.8-h and Figure 4.9-b-c). Because the high-amplitude magnetic signature of western Challenger Plateau is presumed to originate in near-surface rocks [Sutherland, 1999], the coincidence of this magnetic boundary with the lateral extent of the deep high-velocity structure we image may suggest that the geometry of the current plate boundary is indeed controlled by older rheological discontinuities spanning the entire lithosphere. Both group velocity resolution scale (Figure 4.5) and phase velocity RMS uncertainty (Figure 4.6) generally increase toward the outer stations of the array. Thus, uncertainty in the velocity structure we image beneath Challenger Plateau increases with increasing distance from South Island, tempering our interpretation.

4.5.2 Onshore South Island

Our results onshore South Island generally agree with the onshore velocity models presented by Fry et al. [2014]. In the crust we observe a change in shallow velocity structure that correlates spatially with the Junction Magnetic Anomaly (JMA) [Sutherland, 1999] across southern South Island (Figure 4.8-b) at depths up to 25km. Crustal shear velocities north of the JMA are systematically slower than those south of the JMA by roughly 10%. Sub-crustal earthquake hypocenters with reverse and strike-slip focal mechanisms, presented by Boese et al. [2013], align with both the inferred leading edge of a relict Eocene passive margin [Sutherland et al., 2000], and the mantle high-velocity zone we image extending southward of Mt. Cook and westward of the Alpine Fault (Figure 4.9). The upper-mantle earthquakes reported by Boese et al., [2013] directly underlie Low Frequency Earthquake (LFE) hypocenters [Chamberlain et al., 2014] (Figure 4.9-d,e) that spatially correlate with a zone of high attenuation of P-waves (low Qp) [Eberhart-Phillips et al., 2008] and a region of high crustal electrical conductivity [Wannamaker et al., 2002]. The conductive and
low Qp features have been interpreted to result from fluid migration upward from a zone of high-pressure metamorphism of the relict Eocene oceanic lithosphere [Boese et al., 2013; Chamberlain et al., 2014]. The dehydration of this oceanic lithosphere would presumably lead to a localized zone of increased density and more brittle rheology relative to the surrounding mantle [Karato and Jung, 1998], localizing subcrustal seismicity within the dehydrated zone. Our results show a high-Vs zone (Figure 4.9a-c, e) corresponding to the hypocenters of the inferred zone of brittle deformation defined by the subcrustal earthquakes reported by Boese et al. [2013]. Our modeled feature thickens to the west of the Alpine fault and South of Mt. Cook, consistent with the location of the passive margin interpreted by Boese et al., [2013]. Our model is therefore consistent with the hypothesis that the thickened edge of the remnant passive margin could indeed be localizing brittle shear strain [Boese et al., 2013] at depth in the mantle. A thickened high-speed zone, offset to the west of the crustal root (Figures 4.8 - 4.9), could result from convergent thickening below the brittly-deforming zone as proposed by Pysklywec et al., [2002]. Because the MOANA array completely surrounds central South Island, both group velocity resolution (Figure 4.5) and phase velocity RMS uncertainty (Figure 4.6) are improved in this region relative to the edges of the array. Therefore we consider our Vs model of the Central South Island region to be better-constrained than those of the Challenger and Campbell Plateaus.

4.5.3 Canterbury Basin and Bounty Trough

The spatial extent of the low-velocity zone we resolve at shallow depths (Figure 4.8) is consistent with the regions of Cenozoic intraplate volcanism between the Otago and Banks peninsulas, which correspond to high mantle helium isotope concentrations reported by Hoke et al., [2000]. Mantle helium anomalies measured at the Banks and Otago peninsulas are interpreted by Hoke et al., [2000] to derive from relatively recent asthenospheric melts, perhaps originating from the lithosphere-asthenosphere boundary (LAB) at about 80km depth, and which coincide with measurements of high heat flow. Hoernle et al., [2006] propose that the Cenozoic shield volcanism beneath Otago can be explained by asthenospheric upwelling into cavities produced by lithospheric
delamination, which thus do not require a plume origin. The low velocity zone from 60-140km depth we image in this region is consistent with the latter interpretation, but the limited vertical extent of our model (<150km) precludes ruling out a plume-like feature (Figure 4.9). The stalling of Hikurangi subduction along the Chatham Rise has been proposed to result in slab detachment and subsequent asthenospheric upwelling in the last 20Ma [Davy et al., 2008]. The lowest velocities at depths greater than 100km in our model (Figure 4.8f-h) form a region subparallel to the Chatham rise fossil margin, as would be expected for a convective upwelling caused by slab detachment inboard of the Chatham Rise margin. While the geometry of the MOANA array limits raypath coverage on Campbell plateau and confines our modeled area to a small region (e.g. Figure 4.8), the RMS uncertainties of teleseismic phase velocity measurements (the main control on deeper structure in our models) are low relative to those on Challenger plateau (Figure 4.6b). Thus we consider the mantle low velocity structure we image here to be robustly constrained by our data.

4.5.4 Inferred Lithospheric Thickness Variations

Defining the base of the lithosphere using the 4.5km/s shear wave velocity contour as a proxy allows us to gauge first-order changes in the thickness of the high-velocity lid across the MOANA array. Beneath the inner Campbell Plateau we image the thinnest lid in our model, where the 4.5km/s contour ranges from <75km depth to totally absent beneath a portion of the east coast, south of Banks Peninsula (Figure 4.9). The maximum lithospheric thickness of over 150km (exceeding the depth extent of our model) coincides with the high-velocity anomaly we image beneath the western South Island and the Eastern Challenger Plateau, and lithospheric thicknesses average ~80km beneath the Western Challenger Plateau and the Easternmost Campbell Plateau (Figure 4.9). Zietlow et al., [2014] used measurements of shear-wave splitting to distinguish between distributed and localized deformation at depth in the plate boundary. In that work the authors hypothesized that given a lithospheric thickness of ~150km, all distributed shear resulting in the measured anisotropy could be accommodated purely in the lithosphere. The lithospheric
thickness we estimate of ∼140 km beneath western South Island supports this model of distributed deformation in the lithosphere. Beneath the East Coast of South Island and Inner Bounty Trough, we image substantially thinner lithosphere (∼75 km) than that of Challenger Plateau beneath the west coast of the South Island (> 140 km). Based on petrological analysis of Cenozoic Dunedin and Otago volcanics, Hoke et al., [2000] predict a lithospheric thickness of ∼80 km in the region, an asthenospheric source for the magma, and the absence of a plume. Our models indicate that Campbell Plateau lithosphere is indeed on the order of 80 km thick (Figure 4.9), consistent with a model of asthenospheric upwelling resulting from slab detachment beneath the stalled Hikurangi subduction margin at Chatham Rise [Hoernle et al., 2006; Davy et al., 2008].

4.6 Conclusions

Ambient seismic noise and teleseismic surface wave data recorded on continuous vertical components from 29 broadband OBS and 28 broadband land seismometers deployed in the South Island region of New Zealand have been analyzed for Rayleigh wave phase and group velocity variations at periods between 8-70 seconds. By inverting these data we have constructed the first onshore/offshore lithospheric shear velocity model extending beneath the Challenger and Campbell plateaus adjacent to the South Island. We observe that the lithospheric thickness of Challenger Plateau and western South Island is substantially greater than that of inner Campbell Plateau. Our estimate of the inner Campbell Plateau thickness is comparable to the 80 km predicted by [Hoke et al., 2000] given an asthenospheric origin of South Island volcanics. The thickest lithosphere we image occurs beneath the west coast of the South Island and south of Mt. Cook, and could result from underthrusting of Australian plate lithosphere [Allis, 1981] or a combination of convergent overthickening and subduction [Fry et al., 2014; Sutherland et al., 2000]. The spatial extent of this thick, high-speed anomaly in our models corresponds with the presumed location of a remnant Eocene passive margin [Boese et al., 2013; Sutherland et al., 2000] that has been proposed to control present-day deformation along the Alpine fault. We also image a low-velocity feature beneath the Banks and Otago peninsulas, which correlates with surface observations of mantle-sourced helium
isotopes to support the existence of upwelling asthenosphere beneath this area [Hoke et al., 2000; Hoernle et al., 2006]. To the northwest of the South Island we model a high-velocity body consistent with the subduction of Pacific plate oceanic lithosphere. The boundaries of this anomaly coincide with those of inferred pre-existing subduction structures [Davey, 2005, Melhuish et al., 2005] and a magnetic discontinuity across the Challenger Plateau [Sutherland, 1999]. These features all strike sub-parallel to both the paleo-Pacific Gondwana margin and the present-day Alpine Fault, implying that the location and geometry of the Cenozoic Australian-Pacific plate boundary in New Zealand could potentially be related to discontinuities inherited from the Paleozoic.

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