

ABSTRACT

Title of Thesis: **CONSTRAINING LITHOSPHERIC STRUCTURE
ACROSS THE CONTINENTAL BORDERLAND
USING RECEIVER FUNCTIONS**

Zachary Reeves, Master of Science, 2014

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Due to its complex history of deformation, which includes subduction, rifting, and transtensional motions, the California Continental Borderland provides an interesting geological setting for studying how the structure of oceanic and continental plates responds to deformation. We calculate Ps and Sp receiver functions at permanent stations of the Southern California Seismic Network as well as ocean bottom seismometer data gathered as part of the ALBACORE seismic experiment in 2010-2011. Our results indicate that the Outer Borderland has been translated with little to no internal deformation, while the Inner Borderland underwent significant crustal thinning to compensate for the 90° clockwise rotation of the western Transverse Range block. We detect an oceanic seismic lithosphere-asthenosphere boundary at 58 km depth west of the Patton Escarpment. Sp common conversion point stacks confirm wholesale lithospheric thinning of the Inner Borderland and suggest the presence of a slab fragment beneath the Outer Borderland.

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FUNCTIONS

by

Zachary Reeves

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Table of Contents

Acknowledgements	ii
Table of Contents	iii
List of Figures	iv
List of Tables	vi
List of Abbreviations	vii
1 Introduction	1
1.1 Geological Background	3
1.2 Receiver Functions	11
2 Data & Noise	14
2.1 Ocean Bottom Data Quality	20
2.2 Estimating Depth and $\frac{V_p}{V_s}$	23
3 Results	29
3.1 Geographical Patterns in Receiver Functions	29
3.2 Average Crustal Properties	34
3.3 Sp Common Conversion Point Stacks	47
3.4 Discussion	50
3.4.1 Ocean Bottom Data Quality	50
3.4.2 Comparison to Previous Work	52
3.4.3 Mid-crustal Discontinuity	60
3.4.4 Lithosphere-Asthenosphere Boundary	62
3.5 Conclusion	65
A Appendix-H- κ Stacks	68
B Appendix-Receiver Functions	84
Bibliography	124

List of Figures

1.1	Regional Map	4
1.2	Models of Space Accomodation for Rotation of WTR	8
1.3	Receiver Function Schematic	12
2.1	Earthquake Locations	15
2.2	Map of Stations	18
2.3	Handpicking the Data	21
2.4	Effects of Sediment on Synthetics	24
2.5	An Exemple of a High Quality H- κ Stack	26
3.1	Example of Qualitatively Grouping Stations	30
3.2	Quantitative Clustering Map	33
3.3	Depth to Moho Map	41
3.4	Depth to Mid-Crustal Discontinuity Map	44
3.5	CCP Stacks	49
3.6	SNCC Validation	57
3.7	Cross Section of Study Region	60
A.1	Inner Borderland H- κ Stacks	69
A.2	Outer Borderland H- κ Stacks	71
A.3	Abyssal Plain H- κ Stacks	72
A.4	western Transverse Range H- κ Stacks	73
A.5	Peninsular Ranges H- κ Stacks	76
A.6	L.A. Basin H- κ Stacks	78
A.7	Control Station H- κ Stacks	82
A.8	North of western Transverse Range H- κ Stacks	82
A.9	Stations inbetween abyssal plain and Outer Borderland H- κ Stacks	83
B.1	Inner Borderland Ps Receiver Functions	85
B.2	Outer Borderland Ps Receiver Functions	87
B.3	Abyssal Plain Ps Receiver Functions	88
B.4	western Transverse Range Ps Receiver Functions	89
B.5	Peninsular Ranges Ps Receiver Functions	92

B.6	L.A. Basin Ps Receiver Functions	94
B.7	Control Station Ps Receiver Function	98
B.8	North of western Transverse Range Ps Receiver Functions	99
B.9	Stations inbetween abyssal plain and Outer Borderland Ps Receiver Functions	99
B.10	Inner Borderland Sp Receiver Functions	100
B.11	Outer Borderland Sp Receiver Functions	102
B.12	Abyssal Plain Sp Receiver Functions	103
B.13	western Transverse Range Sp Receiver Functions	104
B.14	Peninsular Ranges Sp Receiver Functions	107
B.15	L.A. Basin Sp Receiver Functions	109
B.16	Control Station Sp Receiver Function	113
B.17	North of western Transverse Range Sp Receiver Functions	113
B.18	Stations inbetween abyssal plain and Outer Borderland Sp Receiver Functions	114
B.19	Alternate Water Level for Select Stations for Sp Receiver Functions .	115

List of Tables

3.1	Table of Moho Depths	36
3.2	Table of Average Moho Depth and Standard Deviation by Region . .	42
3.3	Table of Mid-Crustal Discontinuity Depths	45
3.4	Table of Average Mid-Crust and Standard Deviation by Region . . .	48
3.5	Table of Moho Comparisons	53
3.6	Table of Seismic LAB Depth vs. Age	64

List of Abbreviations

ALBACORE	Asthenospheric and Lithospheric Broadband Architecture from the California Offshore Region Experiment
L.A.	Los Angeles
LAB	Lithosphere-Asthenosphere Boundary
OBS	Ocean Bottom Seismometer (Seismic)
Moho	Mohorovičić Discontinuity
$\tilde{P}$	Parent Wave
$\tilde{D}$	Daughter Wave
$\tilde{G}$	Receiver Function of Structure
*	Convolution in time domain or Multiplication in frequency domain
H	Depth
κ	V_p/V_s Ratio
V_p	P wave velocity
V_s	S wave velocity
σ	1 Standard Deviation
WTR	Western Transverse Ranges
OB	Outer Borderland
IB	Inner Borderland
CCP	Common Conversion Point
LARSE	Los Angeles Region Seismic Experiment
IRIS	Incorporated Research Institutions for Seismology
EARS	EarthScope Automated Receiver Survey
SCSN	Southern California Seismic Network

Chapter 1: Introduction

Though the theory of tectonic plates can be traced to Alfred Wegener's Theory of Continental Drift in 1912, the theory only became widely accepted in the 1960s (Frankel, 1987). While phenomenally successful in explaining the kinematics of surface deformation, earthquakes, and volcanoes, plate tectonics does not address what makes the continental and oceanic plates rigid enough to experience little internal deformation, but weak enough to be subducted, rifted, and deformed. Seismic imaging can provide new constraints on how both oceanic and continental lithosphere respond to deformation at depth. However, the challenge with using seismological data to observe difference in structure between continental and oceanic plates is that the vast majority of data that has been gathered is from land based stations. This does not allow oceanic lithosphere and crust to be observed with the same resolution as the continental lithosphere and crust.

The primary goal of the Asthenospheric and Lithospheric Broadband Architecture from the California Offshore Region Experiment (ALBACORE) Project was to observe the Continental Borderland, off of the coast of southern California, because this region has experienced a varied history of deformation that involves both oceanic and continental plates. Therefore, it was located in an ideal setting to study the

differences between the structure of tectonic plates and their response to deformation. In this study of the California Continental Borderland, we used conversions of P waves to S waves (and visa versa) - analyzed using the receiver function method - observed at land station data as well as data collected by ocean bottom seismometers.

This region has had few seismological investigations (Lekic et al., 2011; Zhu and Kanamori, 2000), despite the proximity to Los Angeles (L.A.), mainly due to a lack of data available on the offshore islands and previous ocean bottom deployments, which makes this experiment the first step in understanding this unique region with very specific goals in mind. One of these goals was to develop and verify methods for making the use of ocean bottom data better suited for receiver function analysis. These methods should be applicable to other ocean bottom experiments that seek to perform analysis with receiver functions, such as the Cascadia Initiative (<http://obsip.iris.edu/experiments/cascadia/>). We also want to better understand the large scale relationship between oceanic and continental crust and/or lithosphere. The abyssal plain stations should be able to distinguish between two possible models regarding oceanic lithosphere thickness. Either thickness of the oceanic lithosphere varies with age — and gets thicker as it gets older, due to cooling from above (e.g. Kumar and Kawakatsu, 2011) — or the thickness is dictated by dehydration during partial melting at the ridge axis (Evans et al., 2005; Gaherty et al., 1996). The final, and most specific objectives, were to observe how the depth of the seismic lithosphere-asthenosphere boundary (LAB) varies within oceanic/continental plates and where this change occurs within the region, as well

as to map variations in crustal thickness (Moho depths) to draw interpretations regarding the crustal deformation history in this region.

We present a simple method (attempted other techniques have not been successful) that allows us to estimate Moho depths using ocean bottom seismic (OBS) data for receiver function studies. We also present crustal thickness and LAB depth estimates for the coast of Southern California, the Inner and Outer Borderland, and the abyssal plain. We use H- κ stacking to determine the Moho depths and $\frac{V_p}{V_s}$ ratios and compare these to standard values (Brocher, 2005). We compare our estimate of crustal thickness in the Continental Borderland to previous receiver function estimates made in this region (Zhu and Kanamori, 2000) and discuss implications for interpretations of the history of deformation in this region.

1.1 Geological Background

The Continental (California) Borderland is arguably one of the most geologically unusual regions in the United States because of its plate-boundary location and somewhat enigmatic history of deformation. The term Continental Borderland is applied to the region of the southwestern coast of California that spans from $\sim 32^\circ$ to $\sim 34^\circ$ N latitude and from the coast of California (117°) to 121° W longitude (Figure 1.1). This area coincides with the tectonic plate boundary between the North American and Pacific plates. The Patton Escarpment marks the region's western boundary, beyond which lies the abyssal plain of the Pacific plate. The Malibu Coast Fault separates the Borderland from the WTR block to the north (Crouch

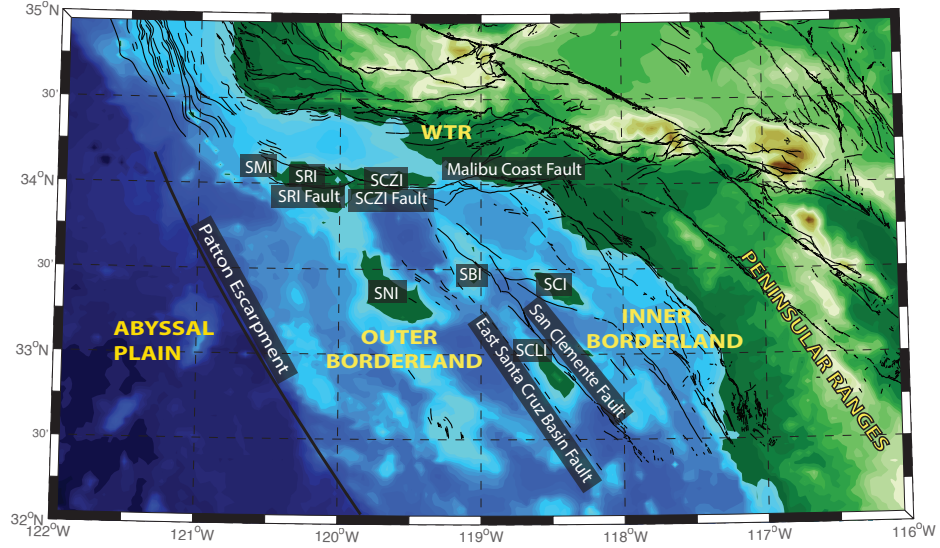


Figure 1.1: Map of study region showing the main faults and structural blocks. Faults are from the Southern California Earthquake Center (www.scec.org) database, except for the East Santa Cruz Basin Fault, which is from Howell et al. (1976). (WTR - western Transverse Range block, SMI - San Miguel Island, SRI - Santa Rosa Island, SCZI - Santa Cruz Island, SNI - San Nicolas Island, SBI - Santa Barbara Island, SCI - Santa Catalina Island, SCLI - San Clemente Island, SRI Fault - Santa Rosa Island Fault, SCZI Fault - Santa Cruz Island Fault)

and Suppe, 1993) The offshore continuation of the Malibu Coast Fault is also referred to as the Santa Cruz Island Fault near and across Santa Cruz Island and the Santa Rosa Island Fault near and across Santa Rosa Island (Hornafius et al., 1986). The Continental Borderland is host to four islands, which are, from west to east: San Nicolas, Santa Barbara, San Clemente, and Santa Catalina. The islands in the northwestern portion (34° N and ~ 120.5 to $\sim 119.5^{\circ}$ W) of the California Borderland (from west to east) are San Miguel Island, Santa Rosa Island and Santa Cruz Island (Junger, 1976), lie within the boundaries of the western Transverse Range (WTR) block.

The Continental Borderland can be subdivided into two blocks with different

morphology, lithology, and history of deformation. The Outer Borderland stretches from the Patton Escarpment to the East Santa Cruz Basin fault, which strikes northwest and lies just to the west of the Santa Barbara and San Clemente Islands, bisecting Santa Cruz Island, until it is truncated by the west-striking Santa Monica Fault (Malibu Coast Fault west of the coast) (Howell et al., 1976) in the north. The Inner Borderland stretches from the coast to the East Santa Cruz Basin Fault, which strikes northwest and runs to the west of the Santa Barbara and San Clemente Islands (Crouch and Suppe, 1993).

Structurally, the Continental Borderland has a basin and range structure that is similar to the basin and range structures observed in the western part of the U.S. and runs generally parallel to the coastline (strikes northwest), with the islands denoting local peaks of the ranges and the bathymetric lows corresponding to the basins (Junger, 1976). Lithologically, the region bears widespread evidence of the subduction zone that existed between the Farallon plate and North American plate. The Outer Borderland consists primarily of the Franciscan Belt, which is a subduction complex and Great Valley Belt, which is a forearc basin deposit (Crouch, 1979). Beneath, seismic studies and offshore drilling detect mafic basement (Bohannon and Geist, 1998; ten Brink et al., 2000). The Great Valley Belts observed in the WTR, Inner Borderland and Outer Borderland have undergone very slight metamorphism and deformation and seem to be derived from a continental source terrane (Crouch, 1979). Within the Inner Borderland, the Santa Catalina Island represents an unroofed metamorphic core complex, where the Catalina Schist is exposed at the surface. The Catalina Schist primarily contains blueschist fa-

cies, which requires relatively high pressure and low temperature conditions that are present within subduction zones, though it also has some higher-temperature lithologies (Grove et al., 2008). The metamorphic grades are consistent with regional subduction-related metamorphism followed by exhumation. The Cataline Schist is exposed on Santa Catalina Island, some on-land (i.e. Palos Verdes Peninsula) and submarine exposures across the Inner Borderland exist (Crouch and Suppe, 1993).

Structural, petrological, and paleomagnetic data help constrain reconstructions of the history of deformation of the Continental Borderland, which has been dominated by the transition from compression to transform motion following cessation of subduction ~ 30 Ma. The fact that Catalina Schist is exposed throughout the Inner Borderland necessitates removal of a substantial amount (~ 40 km) of overburden. Crouch and Suppe (1993) propose that the overburden was removed through rifting that unroofed the metamorphic core complex. The rifting occurred in the wake of the rotation of the WTR (Figure 1.2). There are also observations by ten Brink et al. (2000) of the metamorphic core complex on Catalina Island that support the theory of unroofing proposed by Crouch and Suppe (1993), but propose that the extension in this region was caused by temporary mantle upwelling that accompanied the rotation of the WTR. ten Brink et al. (2000) claim that the heat source for the Inner Borderland was a gap when the Monterey microplate was partially subducted and the mantle upwelling was a product of the extension, as opposed to a driving force for the formation of the metamorphic core complex. ten Brink et al. (2000) point out that they do not contradict the claim that the Inner Borderland is a metamorphic core complex, but propose an alternative method of extension where

they rely on a thermomechanical model where the tear in the Monterey microplate would provide the thermomechanical conditions that would allow for a metamorphic core complex mode of extension, volcanism and relatively fast uplift. An alternate theory, proposed by Luyendyk (1991) and Luyendyk et al. (1980) is that rotation of the WTR was accompanied by rotation of other blocks, which resulted in highly localized extension producing the Catalina and L.A. basins, but did not result in wholesale lithospheric extension (Figure 1.2).

Nicholson et al. (1994) gives a compelling argument for the importance of microplate formation and capture in the history of deformation in the region. The subduction zone between the overriding North American plate and subducting Farallon plate existed until the Farallon plate fragmented, forming the Monterey and Arguello microplates ~ 30 Mya, at which time the spreading rate between the Pacific plate and Monterey microplate slowed down and the Monterey microplate appeared to be captured by the North American plate. The Pacific-North American plate boundary formed ~ 28 Mya and subduction of the Monterey and Arguello microplates continued until subduction of the Monterey microplate completely stopped ~ 20 Mya, at which time it attached itself to the Pacific plate and subduction of the Arguello microplate ceased ~ 17.5 Mya (Nicholson et al., 1994). The difference in subduction rates between the Monterey and Arguello microplates created a right lateral shear zone and this time frame (~ 22 Mya to ~ 18 Mya) is also when the relative plate motion between the North American and Monterey plates shifted from convergent to transtensional. The ridge was eventually subducted, which explains why the Pacific plate gets older from California to Japan.

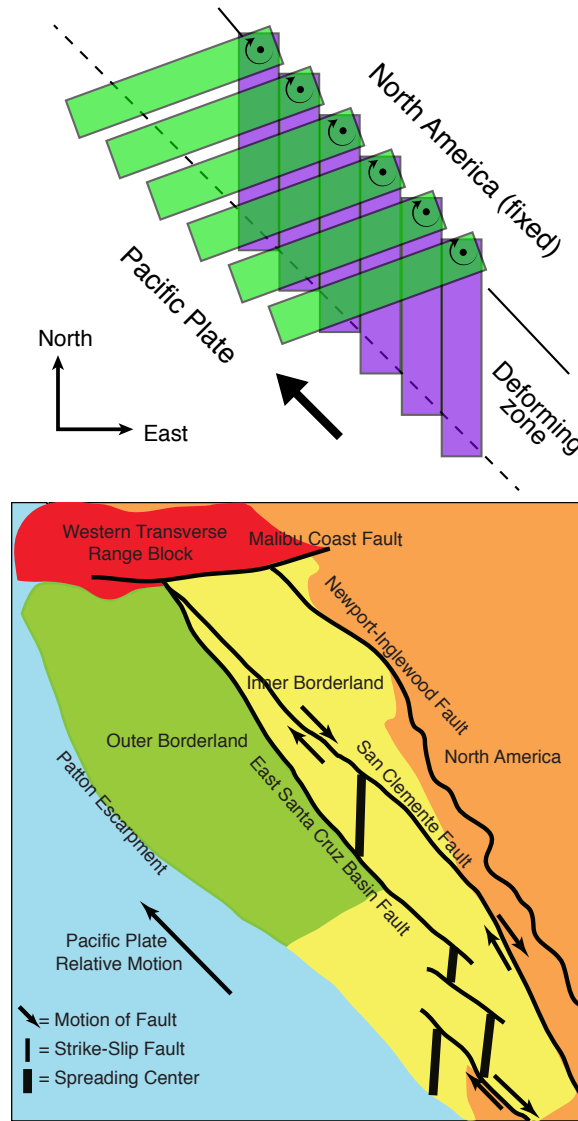


Figure 1.2: (top) The Luyendyk et al. (1980) and Luyendyk (1991) model for the WTR block rotation where blocks rotated together, with relatively little extension occurring in the present-day Inner Borderland. The green rectangles are the blocks post rotation with the spaces representing detachment faulting between blocks. (bottom) The Crouch and Suppe (1993) model where substantial rifting of the Inner Borderland accompanied the rotation of the WTR block. The spreading centers accommodate the rotation of the western Transverse Range block in this model.

Paleomagnetic data from volcanic rocks in the Santa Monica Mountains, Conejo Hills, and east Anacapa Island was collected by Kamerling and Luyendyk (1979) in the WTR showing that multiple instances of rotation have occurred at different times that indicates $\sim 90^\circ$ clockwise rotation. This implies that the WTR block rotated at least 64° (up to 81°) in a clockwise direction. Kamerling and Luyendyk (1979) combined this range with the claim made by Jones et al. (1976) that the Santa Monica Slate and Santa Cruz Island Schist were rotated in a clockwise direction (no amount specified) to get to the generally accepted value of $\sim 90^\circ$ clockwise rotation (Carter et al., 1987; Crouch and Suppe, 1993; Luyendyk, 1991; Luyendyk et al., 1980) in the WTR block. The combination of the transtensional motion and the Monterey microplate taking on the motion of the Pacific plate converted what was once the interface of the subduction zone into a transtensional interface and subjected the North American plate to crustal extension (Nicholson et al., 1994). After the Monterey microplate was captured, the WTR block began to rotate because it was subjected to dextral shear on the eastern boundary and sinistral shear on the southern boundary, with the northern region being coupled with the North American plate at a left step in the boundary, which acted like a hinge allowing the block to rotate (Nicholson et al., 1994). In this model, the WTR block rode atop the part of the Monterey microplate that has been captured by the Pacific plate.

The Outer Borderland was not captured by the Pacific plate until ~ 18 Mya, which gave it time to distance itself from the North American plate (Nicholson et al., 1994). The fact that everything north of the Outer Borderland already had the same motion as the Pacific plate meant that there was no "hinge" like there

was for the WTR, so it was rifted and translated away from the North American plate without rotation occurring (Nicholson et al., 1994). As the subducting oceanic crust became younger, the spreading along the northern portion of the East Pacific rise slowed down ~ 14 Mya, and the motion of the Guadalupe and Magdalena microplates changed from subduction to dextral transform motion (Nicholson et al., 1994). The spreading completely stopped ~ 12 Mya and the Guadalupe and Magdalena microplates were captured by the Pacific plate and adopted the transtensional motion of the Pacific plate (Nicholson et al., 1994). Due to the initial extension in the Gulf of California, beginning between ~ 12 and ~ 14 Mya, Baja California was slightly rotated ($\sim 10^\circ$) and moved ~ 400 km northwest relative to North America, with ~ 150 km of movement to the west occurring prior to ~ 6 Mya when all of the Pacific-North American plate motion was shifted into the Gulf of California (Nicholson et al., 1994). After the Monterey Microplate was captured by the Pacific plate, partially subducted oceanic crust under the North American plate was moving with the same motion of the Pacific plate (Nicholson et al., 1994). This and other (e.g. McQuarrie and Wernicke, 2005) reconstructions of the structural evolution in the region are based on incomplete constraints from geology, paleomagnetism, mineralogy, petrology, geochronology, and fault slip interfaces. This paper will help place more constraints on the crustal structure and depth to seismic LAB Continental Borderland to narrow down the possible models in regard to the deformation history of the region.

1.2 Receiver Functions

We apply the receiver function method to constrain variations in thicknesses of the crust and lithosphere because it has the potential to achieve both high vertical precision and horizontal resolution estimates of depth of lithospheric velocity interfaces (Figure 1.3). This high resolution is necessary given the relatively small scale of the study area. The receiver function method relies on the conversion of P waves to S waves (and vice versa), and is therefore useful in characterizing impedance (velocity, density) contrasts due to structure (i.e. Moho or LAB) in the subsurface (Langston, 1977). The geometry of converted Ps and Sp waves is schematically shown in Figure 1.3. Gradual variation of seismic wavespeeds with depth would not produce detectable conversions. The difference in arrival times between the parent and converted wave allows an estimate to be made about the thickness of the structure.

We are using Ps (P is parent, S is daughter) to observe the Moho for this study because Ps waveforms tend to have much higher signal-to-noise levels compared to Sp, which are contaminated by microseismic noise (e.g. Kolb and Lekic, 2014). Furthermore, typically, the Ps waves converted across shallow impedance contrasts usually though not always arrive before signals bouncing within shallow layers, such as the sedimentary pile. We are using Sp for deeper structures (LAB) because Ps waves converted across deeper structures are typically overprinted by crustal multiples (see Figure 1.3); therefore, Sp receiver functions are particularly well suited for studying deep lithospheric structure. This is because at teleseis-

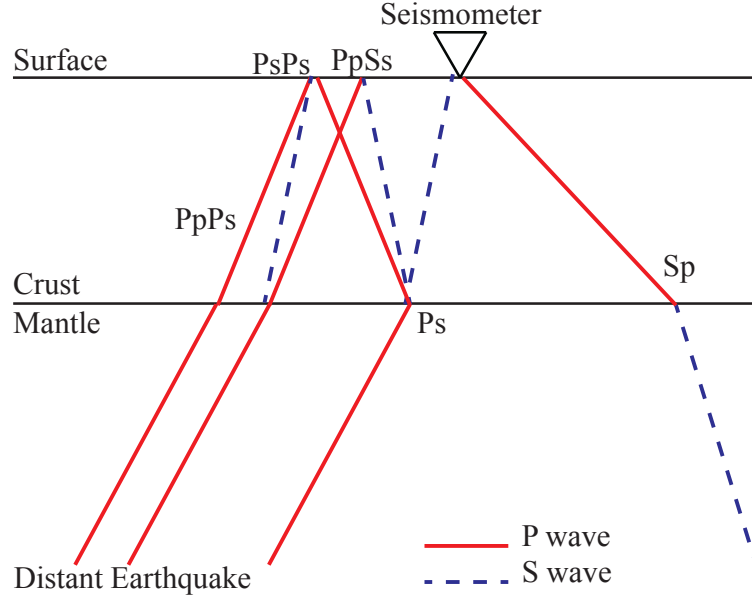


Figure 1.3: Schematic illustration of raypaths of P-to-s and S-to-p conversions across the Moho. Note that while the Ps conversion point is almost directly beneath the seismometer, the Sp conversion point is displaced in the direction of the earthquake.

mic distances, the large time lag between the P and S arrivals means that the Ps multiples have attenuated completely by the time that the signals from the S-to-P conversions arrive. The Sp signals from deep structures, like the LAB, always reach the seismometer before conversions across shallower structures and before the direct S-wave.

To isolate the receiver function, the parent wave is deconvolved from the daughter wave in the time and/or frequency domains. Mathematically, this is written as $P * G = D$ (* denotes convolution in the time domain and multiplication in the frequency/Fourier domain), where P represents the parent wave, G represents the receiver function and D represents the daughter wave. We found higher quality results when we used spectral division in the frequency domain for the Sp receiver

functions; Ps was deconvolved in the time domain. Spectral division is the simplest deconvolution method in the frequency domain, in which the Fourier transform of the daughter wave ($\tilde{D}$) is divided by the Fourier transform of the parent wave ($\tilde{P}$) yielding the Fourier transform of the receiver function of the structure ($\tilde{G}$), ($\frac{\tilde{D}}{\tilde{P}} = \tilde{G}$). Though simple, this method is often unstable because of spectral holes (zeroes) in the spectrum of the parent wave (denominator). To stabilize this method, we apply a water level, where zeroes are set equal to a positive constant to avoid spectral holes or frequencies at which there would be a division by 0. The water level used in this study was allowed to vary by several orders of magnitude with respect to optimal level obtained using the method of Abt et al. (2010). However, even with the use of the water level, it is difficult to obtain reliable receiver functions from single parent-daughter waveform pairs in the presence of high noise levels, which are characteristic of Sp waveforms. Therefore, while we use damped spectral division and simultaneously deconvolve all waveform pairs at a station (e.g. Bostock, 1998), we calculate Sp receiver functions for individual parent-daughter pairs using a version of the extended time multi taper method (Helffrich, 2006; Lekic and Fischer, 2013). With Ps waveforms, we are able to obtain reliable results in the time domain, and therefore use iterative time domain deconvolution (which picks the strongest signals of a receiver function first, then proceeds iteratively to fit smaller details) (Ligorria and Ammon, 1999). In order to estimate uncertainty, we used a bootstrapping procedure (Abt et al., 2010), in which the entire dataset of sources at each station was randomly resampled 20 times allowing replacement, and receiver functions calculated for each resampled dataset realization.

Chapter 2: Data & Noise

We analyzed P-to-S waveforms from events with epicentral distances ranging from 30° to 90° (blue circles in Figure 2.1). To compensate for the sparsity of usable events and high noise levels associated with the OBS data, we extended the standard 35° to 85° range by 5° on both ends to increase the number of waveforms that could be analyzed. The ideal epicentral distance for earthquakes is between 60° and 75° and when the depth is less than 300 km for the S-to-P method. This is because S-to-P waveforms from deeper or more distant earthquakes are contaminated by ScS and SKS (core phases) and P waves that have reflected multiple times off of the surface (Wilson et al., 2006). The S-to-P method is limited to a further epicentral distance range than P-to-S because when the S-waves are converted to P-waves, the angle relative to the vertical increases substantially, whereas when the P-wave is converted to an S-wave, it becomes more vertical. This results in a more narrow range of epicentral distances to generate Sp receiver functions from, which means there are fewer waveforms (data) to use (red circles in Figure 2.1). As with P-to-S, we expand the allowed epicentral distance range to 55° to 85° to try to include more data because there were so few events (< 10) available for analysis with the OBS data. We limited our analysis to earthquakes with a minimum moment magnitude

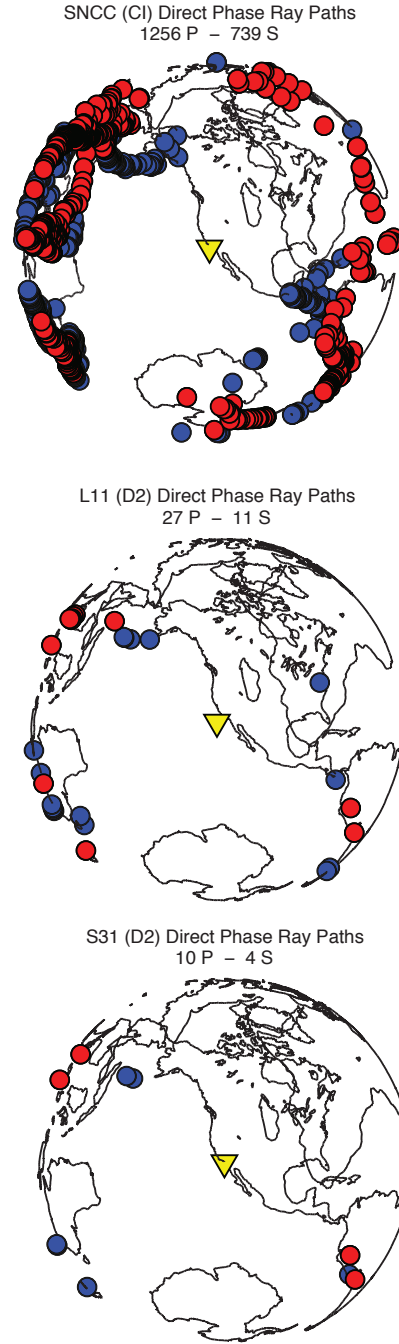


Figure 2.1: Sources of earthquakes for stations SNCC (top), L11 (middle), and S31 (bottom). Blue circles are the earthquakes used in Ps. Red circles are earthquakes used in Sp and Ps. The yellow triangles are the stations.

of 5.8 because, for smaller events, it is unlikely that the signal to noise ratio would be high enough to draw interpretations.

For ALBACORE stations, there were 5549 waveform pairs (1 waveform pair = 1 earthquake detected at 1 station) with a moment magnitude exceeding 5.8 between the broadband stations (squares in Figure 2.2) and of these 5549, data for 3577 ($\sim 64\%$) were obtained (from the Southern California Seismic Network-SCSN) because of earthquakes generating multiple signals that overprint each other, and after I handpicked the data by looking at every individual waveform and keeping only waveforms that had a clear P and S wave arrival, 327 were kept ($\sim 6\%$). The fraction of usable data was much lower at the short period seismic stations (diamonds in Figure 2.2). The main reasons for data rejection were seismometer malfunction, noise, or overprinting of signal by sedimentary multiples. This distribution of the handpicked events with high signal to noise ratios (>2) is not identical from station to station and the actual events selected to be used at each station are not necessarily the same. In general, the stations west of the Patton Escarpment tended to have more waveforms with high signal to noise ratios than the events in, or on the fringe of the California Borderland. The broadband seismic stations have more useful events than the short period stations because the long period stations are more effective at low frequencies, suggesting that the signal to noise ratios decrease at shorter periods. The effects of hand-picking the data are discussed in more detail in the Ocean Bottom Data Quality section.

Data was also requested for the stations on the seven islands in the region, as well as 54 stations along the coastline, islands, L.A. Basin, Peninsular Ranges,

western Transverse Range block, and one station (VTV) that is known to be high quality further inland to serve as a reference station (triangles in Figure 2.2). These stations are part of the permanent Southern California Seismic Network. The data quality for these stations is higher and more abundant. For example, there were 294 waveforms handpicked from the seven island stations (average 42/station) during the same timeframe as the ALBACORE Project, where there were 327 waveforms handpicked from 22 long period ocean bottom stations (average of 14.9/station). The data was also requested for each land/island station from whenever the station was deployed until the present time and there were 85,145 station-event pairs requested for the 61 stations and the number of waveforms obtained was 68,357 ($\sim 80\%$).

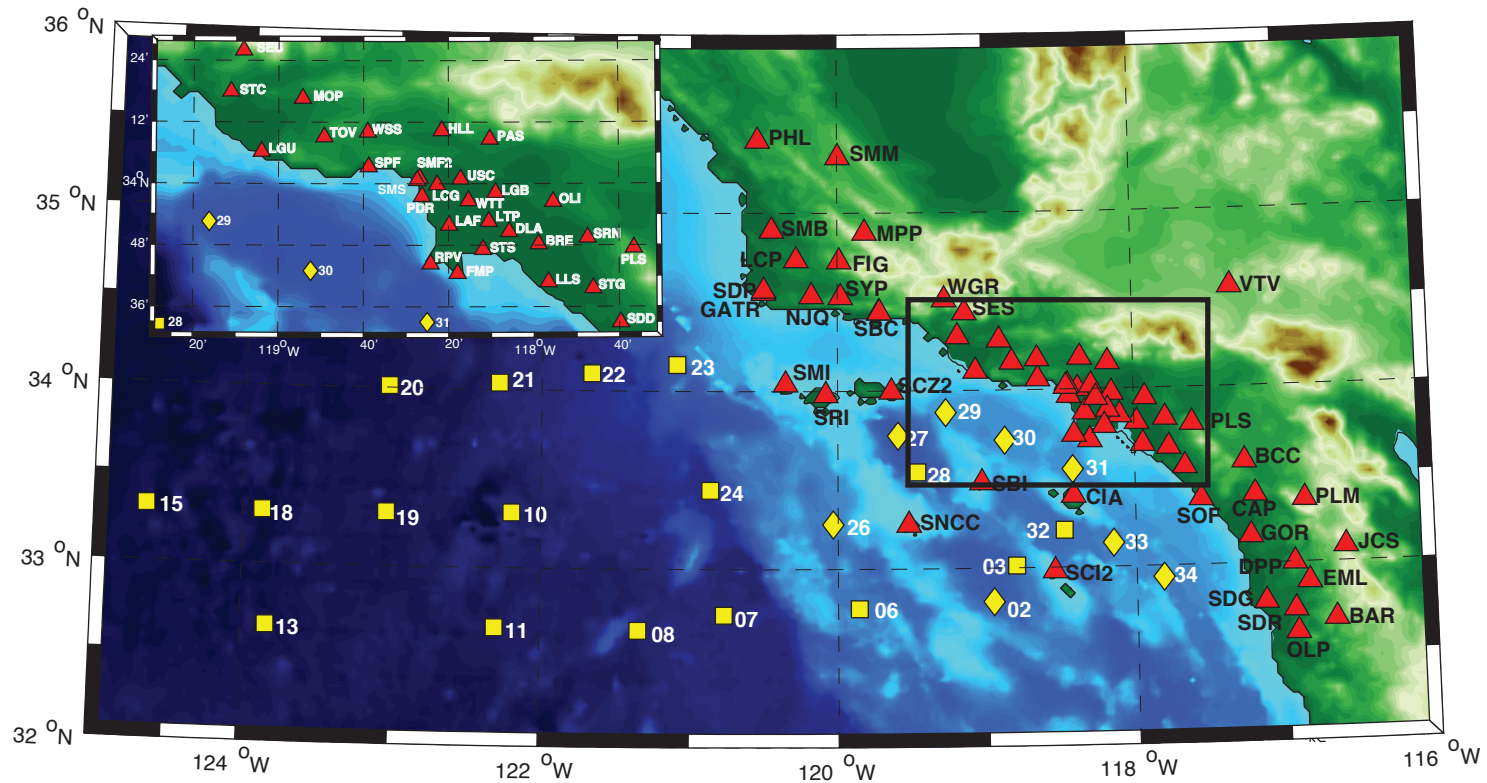


Figure 2.2: Map of locations of the ALBACORE (yellow) long period (squares) and short period (diamonds) deployment as well as the Southern California Seismic Network (CI, red triangles) used in this study. Stations at which no data was recovered are not shown. The inset corresponds to the Los Angeles area (black square). The darker blues are more negative elevations and light greens and browns are higher elevations relative to sea level.

The biggest event during the time of the ALBACORE deployment was the Tohoku earthquake on March 11th, 2011, that recorded a moment magnitude of 9.0 (Ammon et al., 2011). This event also had numerous aftershocks providing a large amount of data during the time of the deployment that was unfortunately unusable. The Tohoku Earthquake itself, lasted so long (rupture duration of 150-160 seconds (Ammon et al., 2011)) that this creates problems analyzing the data because there are P-waves and S-waves still being generated at the same time as the converted waves from the first-arriving P and S are themselves arriving. This overprinting can be corrected for with the deconvolution, but the long duration of the event is incompatible with the 75 and 25 second long P and S analysis windows our standard analysis uses. Furthermore, the data from the Tohoku aftershocks was also unusable since the rate of aftershock occurrence was so great that data from a given aftershock was often overprinted by arrivals from subsequent aftershocks. Despite the inability to benefit from Tohoku (losing ~ 200 waveforms), we are able to extract enough high quality waveforms in the ALBACORE dataset to perform receiver function analysis.

The receiver function method relies on the high quality recordings of P and S waves and their codas. Therefore, we hand-pick all the data available from the ALBACORE deployment and retain only those seismograms on which the P and S arrivals are both clear, and background noise level observed before the P arrival is small ($> \sim 2$). The benefit of handpicking the data can be seen in Figure 2.3, where the top receiver function is a Ps receiver function from station L11 in the abyssal plain after handpicking only high quality events and the bottom receiver function

is a receiver function from the same station with all of the data L11 recorded. The quality difference is substantial, and clear water multiples can be seen in the hand-picked data, whereas coherency of the result is much lower in the data that was not handpicked. The flip side of handpicking high signal-to-noise data from a short deployment is that at many of the ocean bottom seismic stations, only a handful of waveforms remain for which to calculate receiver functions (as low as 1). Therefore, although handpicking allows us to obtain estimates of crustal and lithospheric structure in the first place, these estimates have relatively high uncertainty and low reliability compared to constraints from permanent land stations with more data. Nevertheless, the crustal and lithospheric constraints presented in this study are the only ones available at the time.

2.1 Ocean Bottom Data Quality

There are multiple reasons that ocean bottom data has higher noise levels relative to land station data. One reason is the water layer above the seismometer allows for P waves to travel through the ocean and reflect off of its surface, traveling back down onto the seismometer (i.e. water multiples), which cannot happen with land stations. These water multiples typically arrive after the signal from the Moho, but before the LAB (although in some very shallow water stations it can arrive before the Moho signal as well). Another source of ocean-bottom noise is the fact that the seismometers are often resting on soft, water-saturated sediments, which yield under pressure variations in the ocean environment, tilting the seismometers. Furthermore,

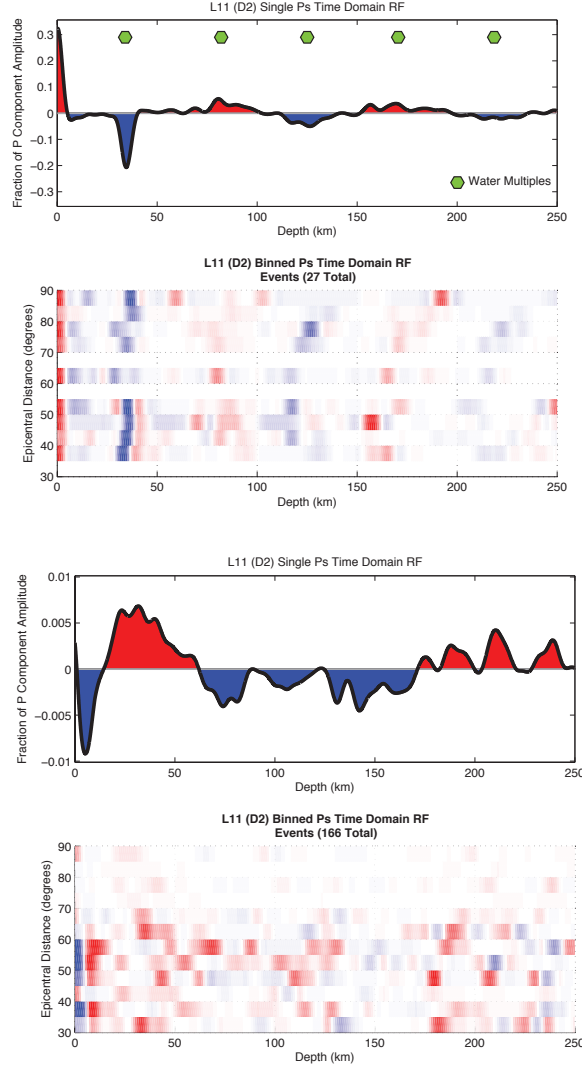


Figure 2.3: (top) A Ps receiver function calculated from handpicked ALBACORE ocean bottom seismometer data for station L11, located in the abyssal plain, with water multiples evident (green hexagons). (bottom) A Ps receiver function for the same station using data that has not been handpicked. Note that hand-picking results in substantially higher signal-to-noise ratios and clear Ps signals. Both receiver functions are constructed from data that has been bandpass filtered from .03 to .12 Hz.

ocean currents, which are driven by tides and are strong even in the abyssal plain, also buffet and tilt the ocean bottom seismometer. This tilting creates long period noise of greater amplitude than typically encountered within land stations, especially on the horizontal components, which start to feel the acceleration due to gravity as they are tilted away from horizontal. Another deleterious effect of the saturated sedimentary layer is that it traps seismic energy bouncing within it, overprinting signals from deeper structures such as the Moho and the LAB. Correcting for this effect is difficult since the P wave velocity, the ratio of the P and S wave velocities, and the thickness of the sediment are all unknown. Finally, the weight of the ocean above the seismometer affects the estimation of the upgoing P-SV wavefields (all P energy on one component, all S energy on a perpendicular component), which the receiver function method heavily relies on, because the common rotation methods (Kennett, 1991) assume that the surface has zero tractions, and the weight of the ocean exerts a vertical traction.

In order to better understand the effects of thin sedimentary layers on the P and S waveforms, we simulated the effects of this sedimentary layer by calculating synthetic waveforms through a range of structures with different sediment thicknesses (Figure 2.4) using the SPECFEM2D code, which is capable of accurately modeling wave propagation through complex structures (Komatitsch and Tromp, 1999). We generated synthetics with a variety of thicknesses for the sedimentary layer, ranging from .05 to 3 km, with an assumed V_p of 1.8 km/s and V_s of .3 m/s. The very high $\frac{V_p}{V_s}$ ratio is appropriate for water saturated sediment. As anticipated, the synthetics showed that that when the sedimentary layer was thin, higher am-

plitude multiples from energy trapped in this layer arrived soon after the parent (P) wave. When the sedimentary layer was thicker, the amplitude of multiples decreased and their arrival was delayed, but the P-to-S conversions across the Moho and other features also arrived relatively later because of the increased $\frac{V_p}{V_s}$ ratio in the sedimentary layer. Therefore, both thick and thin sedimentary piles complicate interpretation of Ps waveforms, but do so in different ways: on one hand, thin sediment does not cause a significant delay in the travel times of Moho P-to-S converted waves, but it contaminates the data with high amplitude multiples; on the other hand, the thick sedimentary layer creates very few and very small amplitude multiples, but also a significant delay in the travel times of converted phases. Signal of water multiples is clearest when the sedimentary layer is not present, but, since the sedimentary layer creates large amplitude signals following the P arrival, it makes the water multiples a secondary issue when the sedimentary layer is present. Based on these synthetic analyses, we determined that the sedimentary layer particularly degrades signals at periods shorter than ~ 8 seconds ($> .125$ Hz); therefore, we filter our data to longer periods using a fourth order Butterworth filter with corner periods of .03 Hz and .12 Hz.

2.2 Estimating Depth and $\frac{V_p}{V_s}$

In order to translate the receiver functions, which are a function of time, to constraints on the depths of structural interfaces such as the Moho, we have to use a velocity model for the subsurface. We can obtain simultaneous constraints on both

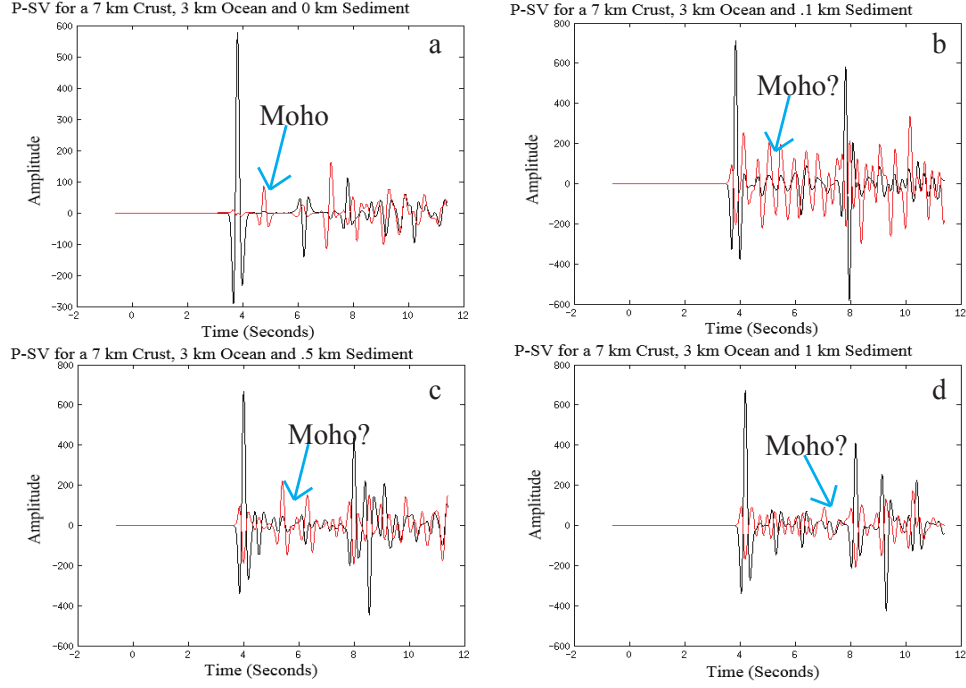


Figure 2.4: Synthetic seismograms (black - vertical component, red - radial component) calculated using SPECFEM 2D showing how the thickness of the sedimentary layer influences Ps receiver functions. As thickness of the layer increases (a. 0 km, b. 0.1 km, c. 0.5 km, d. 1.0 km), the Moho arrival time is later and amplitude smaller. Note the high amplitude (with very little lag time from the expected arrival) ringing set up by constructive interference of the multiples in panel b. Also note that even though sedimentary multiples are smaller when the sediment is thicker, the Moho also has a smaller amplitude. A V_p of 1800 m/s with a $\frac{V_p}{V_s}$ ratio of 6 was assumed for water-saturated sediment. The large amplitude arrivals that start ~ 4 s after the first arrival, and are particularly high amplitude on the vertical component, are the water multiples.

interface depth and the velocity of the material overlying it using the H- κ stacking method. H- κ stacking relies on the fact that the lag time between the arrival of the parent (P) and daughter (s) waves is a function of three structural parameters: the compressional wave speed (V_p), the $\frac{V_p}{V_s}$ ratio (κ), and the depth (H) to the discontinuity across which the conversion occurred. Therefore, the tradeoffs between particularly κ and H mean that Ps by itself can not simultaneously constrain all three of these parameters because there are an infinite number of combinations of the three parameters that could produce the same difference in arrival time between the P and S-waves. When multiple phases are analyzed simultaneously, these tradeoffs can be reduced. Fortunately, converted waves that bounce within crustal layers PpPs, PsPs, and PpSs - depend on the V_p , κ , and H in different ways. We combine these constraints through a stacking procedure in which the lag times of the Ps, PpPs, PsPs and PpSs are predicted for a range of H and κ values. Then, the values of the receiver functions at the predicted times are assigned to each H, κ combination. At the correct choice of H, κ , the three values will all have large amplitudes; therefore, stacking the values of all the phases for all the possible H, κ combinations will produce a maximum at the preferred crustal thickness and $\frac{V_p}{V_s}$ ratio. We perform both linear (summation) and non-linear (multiplication) stacks (Figure 2.5). In the multiplication stacks, we set to zero the values if their amplitude is opposite from that predicted for the converted phase. We elected not to differentially weigh the Ps, PpPs, and PsPs + PpSs contributions as is typically done (e.g. Zhu and Kanamori, 2000) because we found that doing so did not seem to have a very large influence on our parameter selection. A search for optimal weights could be carried out, but

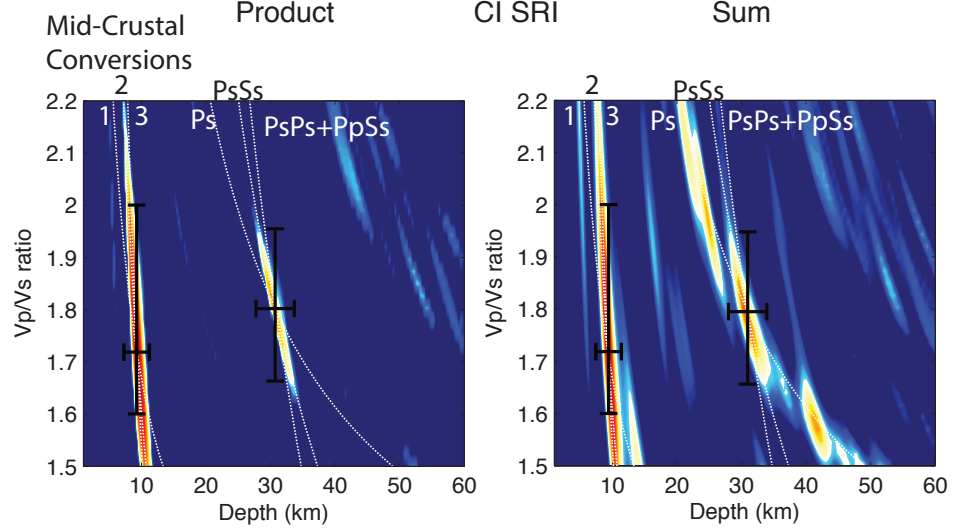


Figure 2.5: Hk stacks of Ps receiver functions (left) product of the Ps, PsSs, and PsPs+PpSs phases to select parameters to see where the signs (+/-) align correctly for the H- κ stack. (right) summation of the phases to observe where the amplitudes are relatively the strongest. The black bars are uncertainty estimates. The dark blue is zero and the brighter the red is, the more positive the signal is. The numbers show the first mid-crustal discontinuity (1) and its first two multiples (2,3).

they would vary from station to station and this type of analysis is beyond the scope of this study. Incidentally, by using unweighted parameters, we ensure that every phase arrives at the proper time and not just the initial Ps phase, to which the most weight is typically assigned.

The thickness of and $\frac{V_p}{V_s}$ ratio estimates that we obtain depend, albeit weakly, on the absolute compressional wavespeed. For a reasonable range of average crustal V_p , this dependence introduces up to a ± 3 km uncertainty on Moho depth estimates. We assumed an average crustal velocity of 6.4 km/s (Zhu and Kanamori, 2000), but we also carried out the H- κ analysis using a crustal velocity of 5.9 km/s to obtain a range of depths as a way to quantify uncertainty. The other way we tried to

quantify uncertainty on H was to specify the values corresponding to the high and low ends of the acceptable $\frac{V_p}{V_s}$ ratios, instead of simply tabulating the thickness associated with the highest amplitude part, which lies in the middle of the high amplitude portion of the stack. A more quantitative treatment of uncertainty on H and $\frac{V_p}{V_s}$ from H - κ stacks, such as that used by Zhu and Kanamori (2000), requires that uncertainty of the stack itself be estimated. Though this can be done with a bootstrapping analysis, it is computationally expensive and we do not attempt it here. Furthermore, estimated uncertainties in this way tend to be far smaller than the ones we obtain (see Table 3.5). The uncertainty due to $\frac{V_p}{V_s}$ ratio on H was typically around 2 km and for the ALBACORE data and very poor land stations, could be as high as 8.

We also constructed common conversion point (CCP) stacks from S_p receiver functions, shown in Figure 3.5, using a procedure identical to that of Lekic et al. (2011), except for the deconvolution, which is done using the modified methodology described in Lekic et al. (2011). To investigate deeper structure in the lithosphere, we analyze S_p receiver functions. Because the S_p conversion points can be displaced far from the receiver (see Figure 1.3), and to take advantage of redundant sampling of different regions in the subsurface by teleseismic S_p for different source-receiver pairs, we construct common conversion point stacks. We begin by mapping S_p receiver functions calculated using the extended time multi taper implementation of Lekic and Fischer (2013) into the volume beneath our region of interest. We calculate the latitude, longitude and depth corresponding to the lag-times of the receiver function, by using ray tracing along the corresponding ray parameter and

back azimuth through a depth profile of velocity constructed using the average crustal thickness and $\frac{V_p}{V_s}$ ratio (Table 3.2) and mantle velocities from ak135 (Kennett et al., 1995). To obtain the value of the CCP stack at each point, we calculate a weighted average of nearby Sp receiver functions in a way that mimics the decreasing resolution with depth (see Lekic et al. (2011) for more detail).

Chapter 3: Results

3.1 Geographical Patterns in Receiver Functions

Qualitatively, we find that Ps receiver functions — and the associated H- κ stacks and Moho / mid-crustal phase maps — among stations located on certain structural blocks are compellingly similar, while those on other blocks are strikingly dissimilar. In Figure 3.1, we plot Ps receiver functions from six geographically distant stations, but which all show a similar pattern of three positive peaks coming from crustal depths. For example, all stations located within the Los Angeles Basin have very similar Ps receiver functions, except for the two located on the western side of the Newport-Inglewood Fault. We discuss this laterally abrupt transition in more detail in the Discussion section. Similarly, the Ps receiver function of station SNCC in the Outer Borderland has much more in common with those at stations of the Peninsular Ranges than the much closer station CIA, located in the Inner Borderland.

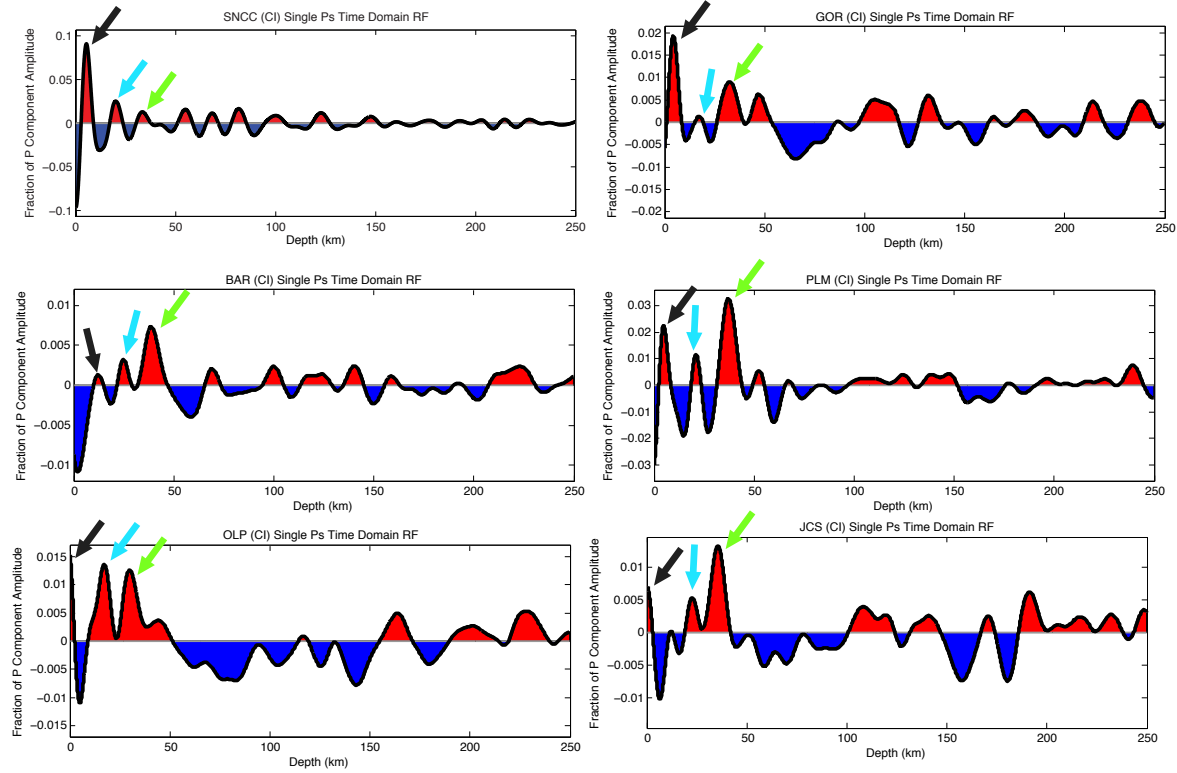


Figure 3.1: The six Ps receiver functions (SNCC in the Outer Borderland and GOR, BAR, PLM, OLP, and JCS in the Peninsular Ranges) are very similar. The first velocity increase is a mid-crustal discontinuity (black arrow). The second velocity increase is (typically) a multiple of the mid-crustal discontinuity (blue arrow). The third velocity increase is the Moho selection for all of these stations (green arrow).

Since receiver functions are proxies for subsurface structure, qualitative differences and similarities among them hold direct implications for the structural evolution and potential deformation history of the region; therefore it is important to undertake a more quantitative analysis. In order to objectively quantify the similarity between various pairs of receiver functions, we use a cluster analysis. Cluster analysis quantitatively evaluates the significance of and would verify the patterns in our qualitative analysis. The usefulness of cluster analysis in identifying regions of similar structure from seismological constraints has recently been demonstrated using both upper mantle (Lekic and Romanowicz, 2011) and lower mantle (Lekic et al., 2012) shear wavespeed profiles. Owing to the receiver functions being representations of sub-receiver structure, we believe that cluster analysis is an appropriate means for quantifying the similarities and differences of structures across the region.

Cluster analysis can be considered an objective means of grouping objects — in our case, Ps receiver function timeseries — in the sense that it uses a distance metric to define the distance/similarity between objects. On the other hand, the choice of the distance metric is itself subjective, especially if it is chosen with a specific result in mind. Cognizant of this danger, we selected a distance metric that ignores geographic distance between stations, to avoid grouping of stations just because they were relatively close to one another. Instead, our chosen distance metric equals one minus the correlation coefficient between two receiver functions, in order to identify which locations are characterized by most similar sub-receiver structures. The distance between groups (clusters) of receiver functions is taken as the arithmetic mean of the distances between each pair of receiver functions across

the two clusters.

In Figure 3.2, we show the results of hierarchical, agglomerative cluster analysis we obtained by setting a minimum threshold of 0.84 for our distance metric (equivalent to a minimum correlation of 0.16 between the clusters). This threshold value results in seven clusters, with two outliers (possibly malfunctioning seismometers), and two groups of stations that only have two receiver functions. The rest of the stations all fit within three geographically widespread groups. The reason that we choose to analyze results obtained with 5-7 clusters is that there are a similar number of major physiographic units (WTR, OB, IB, Coastline, Peninsular Range, L.A. Basin) within our region, making it possible that each area might correspond to one cluster. Instead, we find clusters that are prominent in mostly one region (e.g. the magenta cluster corresponding to thick sedimentary basins), and other times divide regions (e.g. the Western Transverse Range block hosts four different clusters). Reassuringly, the pattern of stations with similar structure confirms the qualitative inferences we made previously. This is particularly clear when looking across the Newport-Inglewood Fault (near Long Beach, California), with stations to the west and east belonging to different clusters, and when seeing that stations SNCC (Outer Borderland), SBI (Island Block), are grouped together with stations in the Peninsular Ranges and station SRI (Western Transverse Range Block). We consider it unlikely that the observed similarities are due to event region (source) character as each receiver function is calculated from waveforms of several hundred earthquakes spanning different source regions (see Figure 2.1).

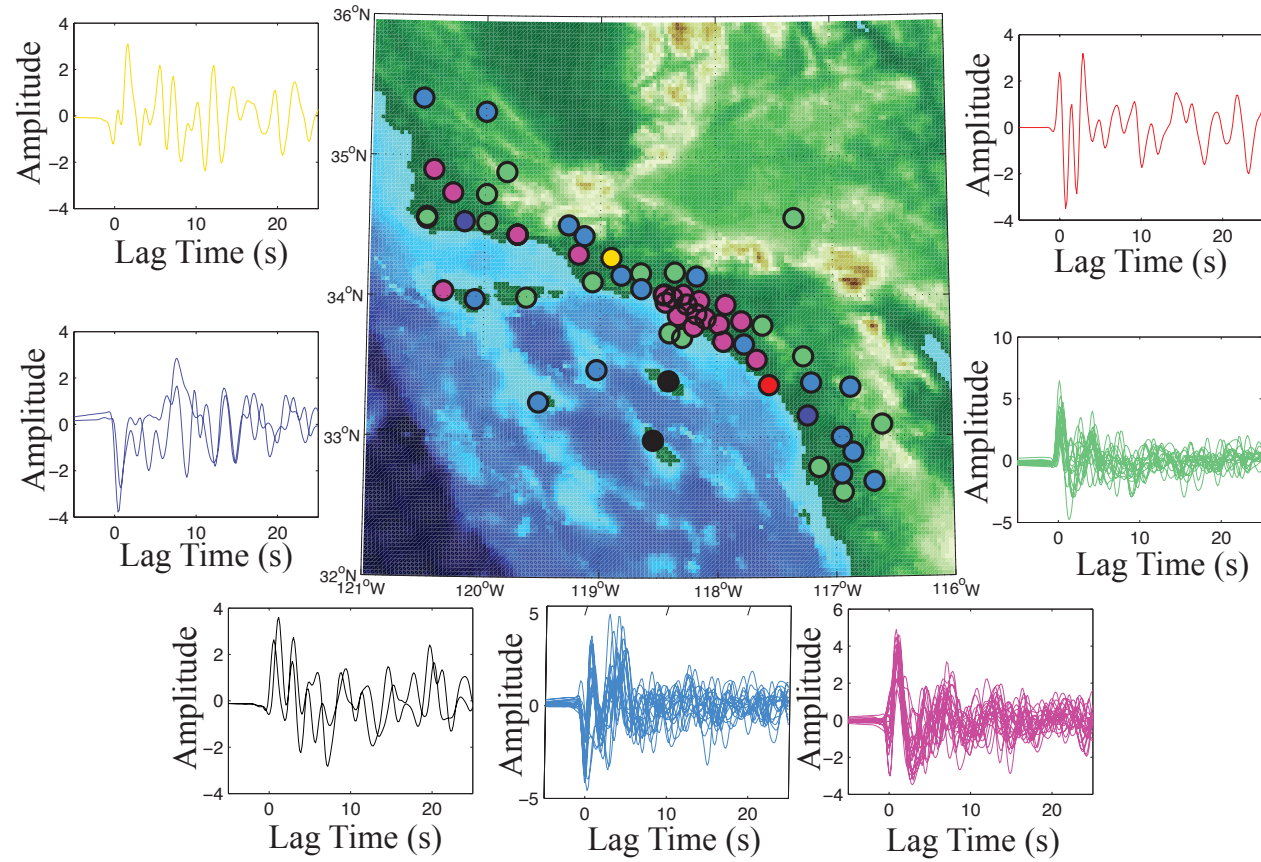


Figure 3.2: Map of all land stations quantitatively clustered (matching colors are one cluster) based on Ps receiver functions. The receiver functions around the map are stacked to show that the receiver functions are similar (correlation $\geq .16$).

At stations that had very poor quality data, the H- κ stacks could not be unambiguously interpreted. In these cases, we use stations with qualitatively similar receiver functions identified through cluster analysis to help guide our interpretation. This allowed us to have more confidence in our inferences of crustal structure parameters from the H- κ stacks, and helped ensure that our Moho (and mid-crustal discontinuity) depths were consistently interpreted across stations. This is particularly important for our interpretations at station SNCC, which is detailed in the Discussion section. Figure 3.2 also shows how similar the receiver functions actually are when all receiver functions in each cluster are plotted on top of each other. This visual similarity increases our confidence that the cluster analysis yields meaningful results.

3.2 Average Crustal Properties

H- κ stacking of Ps receiver functions yielded estimates of Moho depth and average crustal $\frac{V_p}{V_s}$ ratio at most stations in our dataset. Some stations, such as those in the Los Angeles Basin, exhibited no clear Moho-related arrival that could be identified; we strongly suspect this is due to interference with multiples of the sedimentary piles. At thirteen stations, more than one choice of crustal parameters was compatible with the H- κ stacks. Since the duration of deployment, noise levels, and geologic setting varied widely between different stations, so did the reliability of the H- κ stacks, and, therefore, the inferences on crustal structure. Therefore, we assign a quality grade to each station, with a grade of A signifying unambiguous,

low-noise, and high reliability, a grade of B signifying unambiguous but more noise and higher uncertainties on depth, C denoting somewhat ambiguous and high noise results, and a grade of D indicating ambiguous constraints of low reliability. Any station for which we could not estimate the Moho received an E. The Moho depth (measured relative to the elevation of the seismometer), average crustal $\frac{V_p}{V_s}$ ratios, and associated uncertainties appear in Table 3.1. Figure 3.3 shows the maps of Moho depth (top) and average crustal $\frac{V_p}{V_s}$ ratio (bottom) we obtained from H- κ analysis of Ps receiver functions.

Table 3.1: Moho depth (H) and $\frac{V_p}{V_s}$ selections. The * is the control station and ' is the alternate choice for the Moho and V_p/V_s , if it exists. The first H, H', σ_H , V_p/V_s and V_p/V_s' are parameters for an assumed average crustal V_p of 6.4 km/s and the second appearance for these parameters shows values for an assumed average crustal V_p of 5.9 km/s.

Station	H (km)	H' (km)	σ_H	$\frac{V_p}{V_s}$	$\frac{V_p'}{V_s}$	Range	H (km)	H' (km)	$\frac{V_p}{V_s}$	$\frac{V_p'}{V_s}$	Grade
Island											
SCZ2	27.4	20.4	2	1.94	2.02	1.8-2.1	31.2	18.8	1.78	2.03	B
SMI	24.7		1.5	1.92		1.75-1.95	22.6		1.92		B
SRI	30.8		3	1.8		1.65-1.95	27.7		1.82		A
SNCC	31.5		3	1.8		1.65-1.95	28.3		1.84		C
SBI	31		3	1.8		1.7-1.9	28.3		1.8		A
CIA	21.9		2	1.82		1.65-2.0	20.3		1.8		A
SCI2	23.4	28.9	1.5	1.75	1.62	1.6-1.9	21.4	27.8	1.75	1.62	B
Coast											
VTV*	32.4		1.5	1.7		1.6-1.8	28.8		1.74		A
GATR	27		2	1.8		1.7-1.9	24.9		1.79		A
SBC	29.8		2.5	1.89		1.75-2.05	27.1		1.92		B
LGU	30.4	25.4	1.5	1.71	1.86	1.70-1.95	28	23.4	1.7	1.86	B
FMP	23.6		2	1.7		1.65-1.85	12.6		1.7		A
SOF	30.4	23.2	3	1.78	1.98	1.75-2.1	28.6	21.2	1.78	1.98	C
GOR	34.8	27.9	3	1.73	1.9	1.7-2.0	31.2	25	1.72	1.94	B
SDG	25		2	1.64		1.6-1.75	23.2		1.66		C
NJQ	26.2		2	1.84		1.8-1.95	23.7		1.88		C
SYP	28.6		2.5	1.88		1.75-2.0	26.3		1.88		B

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Table 3.1 – *continued from previous page*

Station	H (km)	H' (km)	σ_H	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Range	H (km)	H' (km)	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Grade
STC	21		1.5	1.7		1.6-1.9	19		1.7		B
PHL	25.6		2	1.74		1.65-1.85	23.6		1.74		A
SMM	37.2	23.5	1.5	1.8	1.84	1.75-1.95	34	21.5	1.8	1.84	C
MPP	34.8		1.5	1.72		1.65-1.80	31.8		1.72		B
SMB	N/A										E
LCP	27.1		2	1.72		1.65-1.80	24.8		1.73		C
SMF2	N/A										E
PDR	N/A										E
FIG	30.1		2.5	1.7		1.65-1.90	26.8		1.74		B
RPV	22.3	29.2	1.5	1.82	1.64	1.6-1.9	20.6	26.7	1.82	1.64	B
SDP	30.1		2	1.88		1.8-1.95	27.6		1.88		D
STS	29.2		2	1.86		1.75-1.95	27		1.86		B
SBC	29.8		2.5	1.89		1.75-2.05	27.1		1.92		B
WGR	22.9		4	1.84		1.6-1.95	20.6		1.88		C
LLS	N/A										E
SES	N/A										D
MOP	N/A										D
OLP	28.4		2.5	1.82		1.8-2.1	26.2		1.83		B
BAR	33.7		5	1.86		1.7-1.95	31		1.86		A
SDR	29		3	1.8		1.75-2.0	25.2		1.9		B
EML	35.7		3	1.82		1.7-1.9	32.7		1.82		A
DPP	36		2	1.82		1.75-1.95	33		1.82		B
JCS	36.3	29.9	3	1.74	1.9	1.65-2.0	33.3	26.9	1.74	1.9	A
PLM	38.3	27.4	5	1.73	2.02	1.7-2.05	35.1	25	1.73	2.02	A
CAP	29.2		4	1.8		1.7-2.0	25.6		1.83		A
BCC	33.7		2	1.72		1.65-1.80	30.1		1.74		B

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Table 3.1 – *continued from previous page*

Station	H (km)	H' (km)	σ_H	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Range	H (km)	H' (km)	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Grade
SDD	N/A										E
TOV	25.6		3	1.8		1.7-2.1	23.6		1.8		B
SPF	31	29.5	4	1.85	1.6	1.6-1.95	28	27.4	1.85	1.58	C
WSS	28.3		3	1.9		1.75-2.0	26.1		1.9		B
HLL	N/A										E
USC	N/A										E
PAS	24.4		6	1.88		1.6-2.1	22.3		1.88		A
LGB	28.8		4	1.9		1.75-2.0	26.2		1.9		C
LAF	N/A										E
SMS	N/A										E
LCG	N/A										E
DLA	30		3.5	1.78		1.7-1.85	27.4		1.78		C
LTP	N/A										E
BRE	N/A										E
SRN	28.5		2.5	1.92		1.8-2.1	26.2		1.92		C
STG	31		2	1.77		1.65-1.85	28.3		1.77		B
PLS	28.9		1.5	1.76		1.65-1.80	26.5		1.76		B
OLI	30		1.5	1.82		1.75-1.90	27.5		1.82		C
WTT	27		3	1.9		1.75-2.0	24.7		1.9		C
OBS											
S02	22.6		4	1.71		1.6-1.9	20.6		1.71		B
L03	23		5	1.8		1.65-1.95	21.1	14	1.8		B
L06	23.5	15.5	7	1.71	1.76	1.6-1.9	21.3	14	1.7	1.76	D
L07	23	7.5	5	1.7	1.71	1.6-1.8	21.1	6.9	1.7	1.71	C
L08	8		2	1.8		1.6-1.95	7.5		1.8		A
L10	6.5		2	1.76		1.6-1.95	5.9		1.76		A

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Table 3.1 – *continued from previous page*

Station	H (km)	H' (km)	σ_H	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Range	H (km)	H' (km)	$\frac{V_p}{V_s}$	$\frac{V_{p'}}{V_s}$	Grade
L11	8.5		2	1.9		1.6-1.95	7.8		1.9		A
L13	8.5		2	1.9		1.6-1.95	7.6		1.9		A
L15	9		2	1.8		1.6-1.85	8.4		1.8		A
L18	8.5		2	1.9		1.6-1.95	7.8		1.9		A
L19	9.5		2	1.9		1.6-1.95	8.7		1.9		A
L20	7.8		2	1.9		1.6-1.95	7.2		1.9		A
L21	9		2	1.92		1.6-2.0	8.4		1.92		A
L22	22	7	4	1.7	1.7	1.6-1.8	20.6	6.5	1.68	1.7	B
L23	17		4	1.78		1.6-1.95	15.5		1.78		B
L24	21.4		5	1.71		1.6-1.9	19.7		1.71		C
S26	29.5		6	1.73		1.7-2.0	26.5		1.76		C
S27	19.4		5	1.66		1.6-1.9	18		1.66		D
L28	21.4		5	1.73		1.7-1.85	20		1.71		C
S29	21.8		8	1.9		1.6-2.0	19.4		1.92		E
S30	20.6		6	1.71		1.6-1.95	18.5		1.71		C
S31	18.5		5	1.78		1.6-1.9	16.7		1.8		B
L32	17		4	1.76		1.6-2.0	15.5		1.76		C
S33	20.2		4	1.76		1.6-1.9	19.1		1.75		C
S34	21.2		4	1.8		1.65-1.95	19.4		1.8		B

We find that the crustal thicknesses vary substantially across the region of study, from 6-8 km in the abyssal plain to >35 km beneath the Peninsular Ranges. We also find significant variations between average crustal thicknesses across structural blocks (see Table 3.2). Relatively thick crust is observed for the WTR block, as constrained by stations SCZ2, SMI, SRI, GATR, SBC, LGU, SDP, NJQ, SYP, WGR, STC, LCP, FIG, SMB, and MPP, which have an average Moho depth of 28.4 ± 3.9 km (1σ). Similar crustal thicknesses are obtained for the L.A. Basin, which has an average crustal thickness of 28.5 ± 2.1 km, and the Outer Borderland, whose three stations SNCC, OBS06, and OBS26 having an average Moho depth of 28.2 ± 4.2 km. On the other hand, the Inner Borderland exhibits a substantially thinner crust, as constrained by stations CIA, FMP, RPV, SCI2, SBI, S02, OBS03, OBS27-OBS34, with an average Moho depth of 21.9 ± 3.2 km. The crust of the abyssal plain (L08-L21) appears to be unremarkable, with an average thickness of $8.4 \pm .9$ km. It should be noted that much of the variation could be due to the lower quantity and quality of data. Thickest crust is found beneath the Peninsular Range, with stations (BCC, CAP, PLM, JCS, DPP, EML, SDR, BAR, OLP, SDG, GOR, SOF) having an average Moho depth of 32.5 ± 4.0 km, in agreement with previous studies (Ichinose et al., 1996). It should be noted that stations that are not included in the average and standard deviation calculations (Table 3.2) are stations with no Moho, L07, L22, L23, L24, PHL, and SMM. Higher $\frac{V_p}{V_s}$ (> 1.8) ratios suggest some sort of sedimentary or unconsolidated layer is on top of the crust, while typical values (1.6-1.8) indicate that there is little to no sediment beneath the station.

At many of the L.A. Basin stations we could not determine a Moho depth and

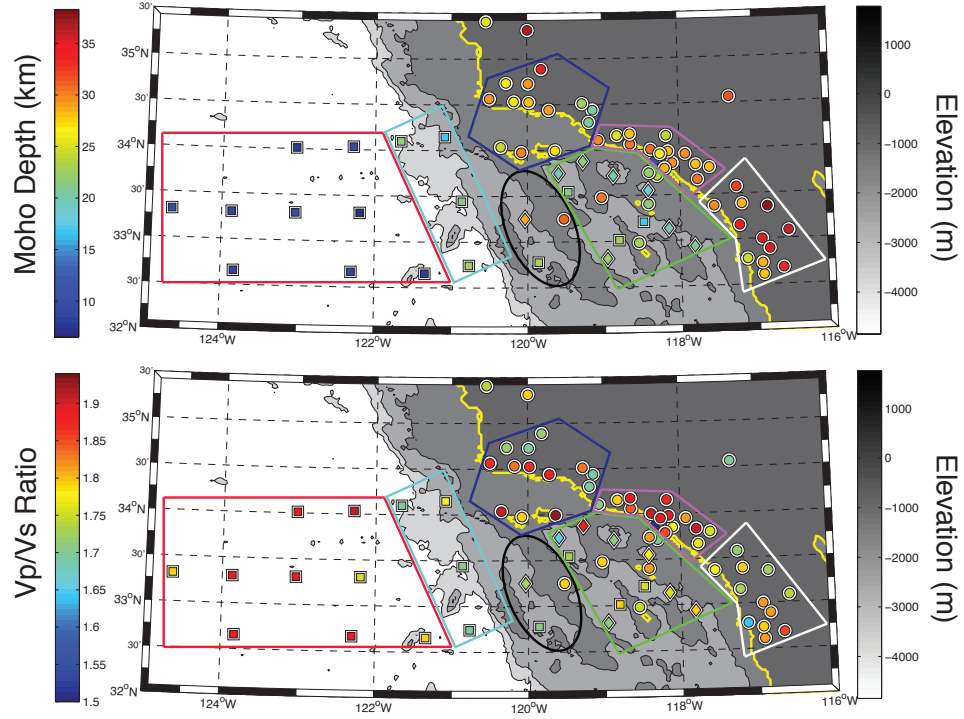


Figure 3.3: (top) Depths to the Moho for the ALBACORE (squares/diamonds) and island/coastal stations (circles) determined using H- κ stacking of Ps receiver functions. Note the systematic differences (relatively thin Inner Borderland, Outer Borderland and coastal stations roughly equal, expected thickness for abyssal plain) in crustal thickness across different structural blocks. (bottom) $\frac{V_p}{V_s}$ ratios corresponding to the Moho depths (top) chosen from the H- κ stacks. Outlines encompass stations used in average and standard deviation calculations by region (Table 3.2). Red = abyssal plain stations, Light Blue = stations inbetween the abyssal plain and Outer Borderland, Black = Outer Borderland stations, Dark Blue = western Transverse Range stations, Pink = L.A. Basin stations, White = Peninsular Ranges stations

Table 3.2: Average Moho depth and standard deviation by region
 IB = Inner Borderland OB = Outer Borderland AP = Abyssal Plain
 WTR = western Transverse Ranges PR = Peninsular Ranges L.A. =
 Los Angeles Basin

Region	Moho Depth		Vp/Vs	
	Average	Std. Dev.	Average	Std. Dev.
IB	21.86	3.13	1.77	0.06
OB	28.17	4.16	1.75	0.05
AP	8.37	0.87	1.86	0.06
WTR	27.92	3.55	1.81	0.09
PR	32.54	4.04	1.77	0.06
L.A.	28.5	2.12	1.84	0.06

$\frac{V_p}{V_s}$ ratio using H- κ stacking of Ps receiver functions because of contamination by multiples trapped by the sediment layer. Because Sp is not affected by multiples and sedimentary layers the way that Ps is, we could often observe clear Moho-related conversion on Sp receiver functions for these stations. Stations LGU and SCZ2 in the western Transverse Range block have two peaks in the H- κ stacks, either of which could be interpreted as the Moho, (LGU at 30.4 and 25.4 and SCZ2 at 27.4 and 20.4), but the deeper value has been chosen for both stations because the $\frac{V_p}{V_s}$ ratio associated with that choice seems more reasonable and these values coincide with the Moho depths found at surrounding stations. Other stations have alternate Moho choices, but this is typically because of poor data quality and/or contamination by multiples, and none of these stations have a double peak that looks like the one observed at SCZ2 and LGU. It should be noted that the possibilities for Moho depths for LGU and SCZ2 are ~ 7 km apart, which has led to them being interpreted as arising from oceanic crust from the Monterey microplate (Cheng, 2008; Nazareth and Clayton, 2003; Schmandt and Clayton, 2013). It is possible

that one of these peaks is a multiple from a shallow mid-crustal discontinuity, but this possibility seems unlikely at these two stations because of the large amplitude and deep but similar depths of the observed signals at these stations.

In addition to obtaining average crustal thicknesses and $\frac{V_p}{V_s}$ ratios, we also used H- κ stacking to characterize potential mid-crustal discontinuities. Figure 3.4 and Table 3.3 show the depths and average $\frac{V_p}{V_s}$ associated with mid-crustal discontinuities for stations at which they are observed. The first positive velocity increase was chosen at each land station because everything after it could (and typically was) a multiple of the first positive discontinuity. The estimates for $\frac{V_p}{V_s}$ ratios and uncertainty were given the broadest range that seemed reasonable because it is impossible to constrain the $\frac{V_p}{V_s}$ ratios to a narrow range using the H- κ stacks. It should be noted in Table 3.4, that stations that are not included in the average and standard deviation calculations include stations with no mid-crustal discontinuity, L07, L23, L24, PHL, and SMM. Beneath the L.A. Basin we find a generally consistent mid-crustal discontinuity, with an average depth of 5.3 ± 1.6 km and an average $\frac{V_p}{V_s}$ ratio of 1.85. Beneath the Inner Borderland, western Transverse Range block, and Peninsular Ranges, we observe the midcrustal discontinuity between 7 and 8 km depth, with $\frac{V_p}{V_s}$ ratios < 1.85 . The Outer Borderland only has one station (SNCC) where the mid-crustal discontinuity is observed, so there is no standard deviation or variance. At SNCC, the mid-crustal discontinuity is observed at a depth of 6.6 km with a $\frac{V_p}{V_s}$ ratio of 1.66. There are no abyssal plain stations with mid-crustal discontinuities except for two stations close to the Patton escarpment, which also have deep Mohos.

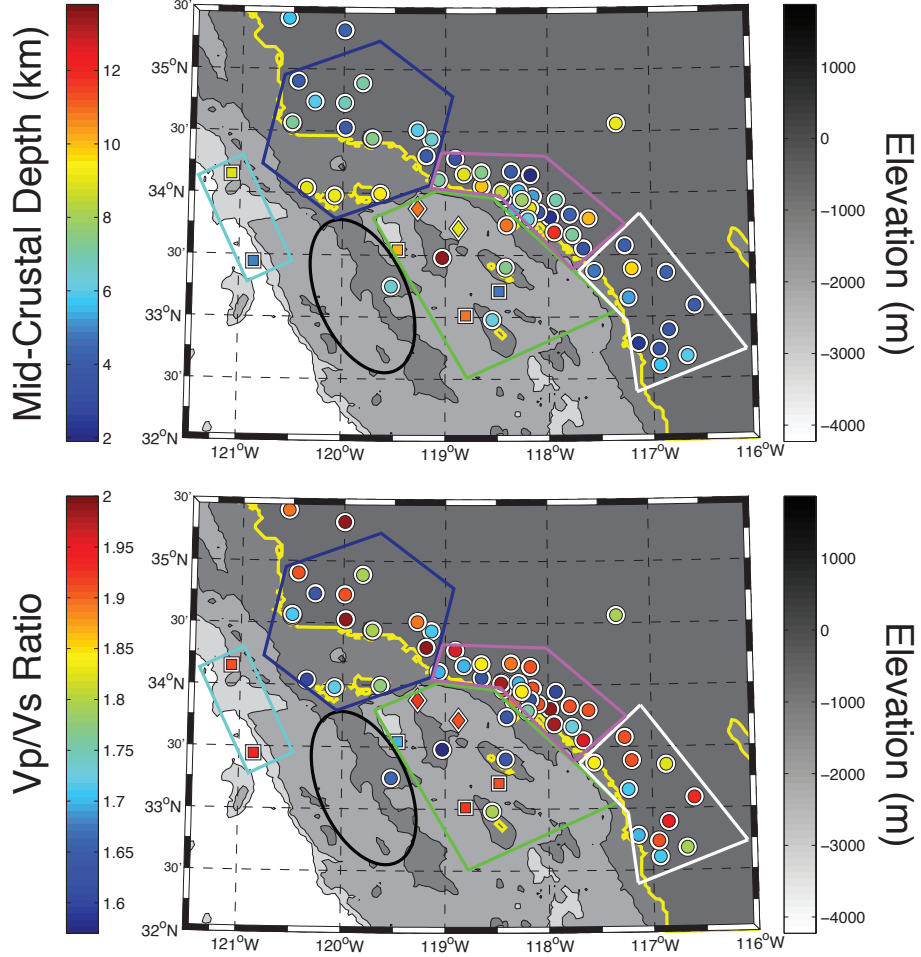


Figure 3.4: (top) Depths to the mid-crustal layer for land/OBS stations determined using $H-\kappa$ stacking of Ps receiver functions. Note the systematic differences between the L.A. Basin and the other regions. (bottom) $\frac{V_p}{V_s}$ ratios at stations corresponding to the mid-crustal depths chosen (top) from the $H-\kappa$ stacks. Shapes encompass stations used in average and standard deviation calculations by region. Light Blue = stations inbetween the abyssal plain and Outer Borderland, Black = Outer Borderland stations, Dark Blue = western Transverse Range stations, Pink = L.A. Basin stations, White = Peninsular Ranges stations

Table 3.3: Mid-Crustal Discontinuity Depths and Vp/Vs Selections

Station	H (km)	σ_H	Vp/Vs	Range
Island				
SCZ2	9.6	2	1.78	1.6-2.0
SMI	9.2	2	1.6	1.6-2.0
SRI	9.3	2	1.72	1.6-2.0
SNCC	6.6	2	1.66	1.6-2.0
SBI	13.8	3	1.57	1.55-1.8
CIA	7.5	2	1.64	1.6-1.9
SCI2	6.3	3	1.8	1.65-2.0
Land				
VTV	9.3	3	1.8	1.6-2.0
GATR	7.8	3	1.92	1.65-2.0
SBC	3.4	5	1.92	1.6-2.0
LGU	7.2	5	1.7	1.6-2.0
FMP	N/A			
SOF	4.7	2	1.85	1.6-2.0
GOR	5.2	3	1.71	1.6-2.0
SDG	2.2	5	1.7	1.6-2.0
NJQ	N/A			
SYP	10.1	3	1.62	1.6-1.9
STC	8.8	4	1.65	1.6-2.0
PHL	7.1	3	1.8	1.6-2.0
SMM	7.3	3	1.71	1.6-1.9
MPP	6	3	1.64	1.6-1.8
SMB	7.5	4	1.8	1.6-2.0
LCP	6.9	4	1.92	1.6-2.0
SMF2	3.4	3	1.97	1.7-2.1
PDR	3.5	3	1.97	1.7-2.1
FIG	4.2	3	1.92	1.6-2.0
RPV	1.9	2	1.92	1.6-2.0
SDP	4.3	4	1.83	1.6-2.0
STS	8.7	3	1.99	1.6-2.1
SBC	3.4	5	1.92	1.6-2.0
WGR	7.2	3	1.73	1.6-2.0
LLS	4	4	1.94	1.6-2.0
SES	11.1	3	1.64	1.6-2.0
MOP	13.2	4	1.59	1.59-1.9
OLP	12	4	1.99	1.7-2.1
BAR	11.1	3	1.93	1.7-2.1
SDR	10.2	3	1.92	1.6-2.0
EML	6	2	1.8	1.6-1.9

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Table 3.3 – *continued from previous page*

Station	H (km)	σ_H	Vp/Vs	Range
DPP	N/A			
JCS	9.9	2	1.92	1.7-2.1
PLM	7	2	1.6	1.6-2.0
CAP	10.2	3	1.7	1.6-2.0
BCC	9	4	1.92	1.6-2.0
SDD	5.7	3	1.9	1.6-2.0
TOV	4.3	4	1.92	1.6-2.0
SPF	6.6	3	1.71	1.6-2.0
WSS	6.3	3	1.76	1.6-1.9
HLL	2.1	3	2	1.6-2.0
USC	4	4	2	1.6-2.0
PAS	9.1	3	1.62	1.6-2.0
LGB	4	4	1.9	1.6-2.0
LAF	3.7	3	1.92	1.7-2.1
SMS	3.7	3	1.92	1.7-2.1
LCG	3.7	4	1.96	1.7-2.1
DLA	4.8	4	1.92	1.6-2.0
LTP	5.7	3	1.9	1.6-2.0
BRE	4.9	3	1.94	1.6-2.0
SRN	4	4	1.92	1.7-2.1
STG	4.3	3	2	1.6-2.0
PLS	5.7	4	1.71	1.6-2.0
OLI	5.2	4	1.92	1.6-2.0
WTT	4	4	2	1.7-2.1
OBS				
S02	N/A			
L03	9	3	1.7	1.6-2.0
L06	N/A			
L07	N/A			
L08	N/A			
L10	N/A			
L11	N/A			
L13	N/A			
L15	N/A			
L18	N/A			
L19	N/A			
L20	N/A			
L21	N/A			
L22	N/A			
L23	5.5	2	1.7	1.6-2.0
L24	5.9	3	1.9	1.6-2.0
S26	N/A			

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Table 3.3 – *continued from previous page*

Station	H (km)	σ_H	Vp/Vs	Range
S27	N/A			
L28	7.5	4	1.85	1.6-2.0
S29	9	3	1.9	1.6-2.0
S30	7.5	3	1.8	1.6-2.0
S31	N/A			
L32	8	4	1.85	1.6-2.0
S33	N/A			
S34	N/A			

3.3 Sp Common Conversion Point Stacks

Figure 3.5 shows cross-sections through our Sp CCP images, with warm (cool) colors indicating a rapid velocity increase (decrease) with depth. Slices C and D run in an east-west direction that is perpendicular to the coastline of southern California, and show lithospheric structure variations across the Continental Borderland. The Moho seen in these cross-sections is consistent with the observations made from H- κ stacking. Cross-sections A and B show a Moho at a depth of ~ 20 km in the Inner Borderland and ~ 30 km in the western Transverse Range block (cross-sections A and B) and L.A. Basin (cross-section B). Cross-section C shows a Moho at ~ 30 km depth in the Outer Borderland, ~ 20 km in the Inner Borderland and ~ 30 km in the Peninsular Ranges. Cross-section D shows the Moho transition between the abyssal plain and western Transverse Range block at about ~ 20 km depth and a depth of ~ 30 km in the western Transverse Range block. These cross-sections suggest an LAB that is roughly 70 km deep in the Peninsular Ranges and decreases

Table 3.4: Average mid-crust depth and standard deviation by region
 IB = Inner Borderland, OB = Outer Borderland, WTR = western Transverse Ranges, PR = Peninsular Ranges, L.A. = Los Angeles Basin

Region	Mid-Crust		Vp/Vs	
	Average	Std. Dev.	Average	Std. Dev.
IB	7.83	3.09	1.78	0.12
OB	6.6	0	1.66	0
AP	N/A	N/A	N/A	N/A
WTR	6.99	2.21	1.78	0.12
PR	7.95	3.12	1.82	0.13
L.A.	5.3	1.57	1.85	0.13

to 50 km depth in the Inner Borderland and increases back up to 80 km in the Outer Borderland. This is observable in the southern part (cross-section C) of the Continental Borderland and once the cross-sections reach the western Transverse Range block (5th hash mark from the left of D and 6th hash mark from the top of A), the LAB depth appears to be constant around 70 to 80 km depth. We find that the strength of the LAB signal in the cross sections weakens northward (cross-section D), in agreement with recent findings of Ford et al. (2014). There is also an unexpected positive phase in the southern part of the borderland that disappears as the cross-sections move northward at ~ 50 km depth. This phase is present in the southern part (hash marks 1-4 of Figure 3.5) of cross-section A, the most southern part (hash marks 1-2) of cross-section B, and the western portion (hash marks 4-7) of cross-section C. We found that the phase exists in every subset of the data, even when removing each island station one at a time or southern coastal stations (GOR,SDG,SOF).

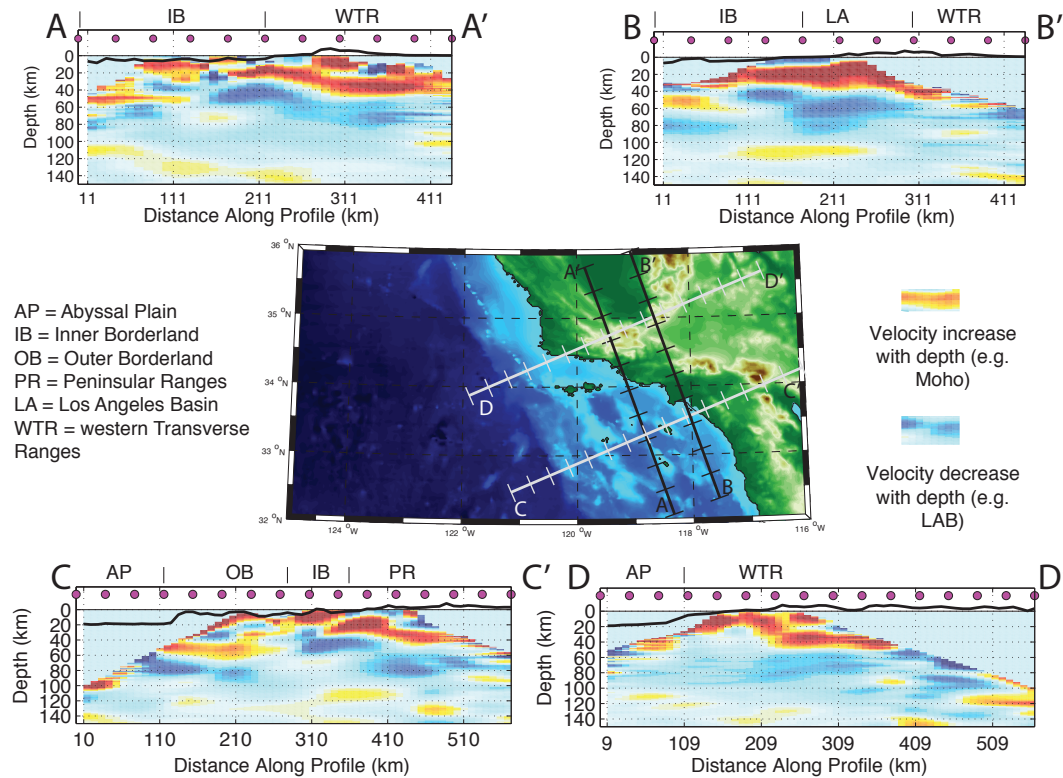


Figure 3.5: CCP stacks through cross-sections. Red indicates a velocity increase with depth, while blue indicates a velocity decrease with depth. The pink dots in the cross-sections correlate to the black hash marks on the map. Note the changes in the strength of the lithospheric signal between the south and north in cross sections C and D, as well as the strong positive phase at 50 km depth in the western part of cross-section C.

3.4 Discussion

3.4.1 Ocean Bottom Data Quality

A central aspect of this study was to obtain interpretable receiver functions from ocean bottom seismic data. Despite the relatively few usable waveforms gathered by stations of the ALBACORE deployment, we were able to obtain constraints on both crustal thickness variations - using Ps - and the depth to the lithosphere-asthenosphere boundary - using Sp. At some stations, no robust estimates could be made based on the little available data. For example, OBS02 shows a Moho depth of 22.6 km in a receiver function created from five events, which is shown by bootstrap analysis as insufficient to have confidence in the results. Unfortunately, there is little alternative with the data set being so small, except to take caution when using the results at these stations to draw interpretations.

Despite the fact that the transformation to get rid of water multiples was unable to be implemented, we were able to make observations using the OBS data. Ideally, this would have improved the data quality of the stations, though it is unlikely that the signal to noise ratios will ever be equivalent to high quality land stations. If the transformation to obtain the upgoing P-SV wavefield performed as expected, the large positive phase near zero lag time would have been mostly eliminated and the signature of the water multiples would have been diminished. This should give future experiments an increased ability to observe an LAB depth from Ps OBS data that would enable map variations in lithospheric thickness to be

made with higher confidence, similar to what we have done for crustal thickness.

The method used to improve the data quality was to handpick only the good events (seismograms with clear P wave, S wave, and surface wave arrivals), which for this project is in the range of $\sim 6\%$ of the waveforms requested at the broadband stations. One possible way of further improving the receiver function constraints in future analyses of the ALBACORE data would be to implement denoising based on transfer functions (Dolenc et al., 2007). Transfer functions are functions that represent the relationship between the pressure in the water column and the response of the seismometer. Because the OBS noise is partially due to these pressure variations, transfer functions can be used to remove these effects from the seismogram and improve the signal-to-noise ratio and quality. We were unable to take advantage of transfer functions because our analysis requires three component data and reliable transfer functions between the pressure measurements and horizontal components could not be reliably estimated.

The handpicked waveforms had to be filtered to long periods (.03 to .12 Hz) to avoid the natural frequencies of the ocean noise ($\sim .14$ Hz) and the sedimentary layer. The sedimentary layer has less influence on the long period waves because the ~ 40 km waves ($\text{wavelength} = \frac{\text{speed}}{\text{frequency}} \approx \frac{5 \text{ km s}^{-1}}{.125 \text{ s}^{-1}}$) are not nearly as effected by the thin sedimentary layer as are waves with shorter wavelengths, which we found in synthetic tests to become readily trapped in the sedimentary pile. Similar filtering can be applied to other ocean bottom research relying on teleseismic waves because it reduces noise from the ocean environment, and avoids many of the complications due to the sedimentary layer, therefore making the use of OBS data more practical.

It should be noted that by filtering to long periods, the resolution is degraded laterally, as well as in depth.

3.4.2 Comparison to Previous Work

Scarcity of data aside, we were able to successfully interpret the Ps H- κ stacks and obtain crustal thickness estimates. Though our results generally confirm previous inferences of crustal structure (Table 3.5), especially for the onshore regions, we note a number of consequential discrepancies between the current results and previous work. Previous Ps receiver function studies of this region to obtain crustal thickness estimates by Zhu and Kanamori (2000) and the IRIS EarthScope Automated Receiver Survey (IRIS EARS) program (Crotwell and Owens, 2005). When comparing our findings, we focus our attention to coastal and island stations because they provide information on the structure - and therefore the formation and deformation - of the Continental Borderland.

Zhu and Kanamori (2000) performed H- κ stacking of Ps receiver functions from the stations (and earthquakes) available to them throughout the L.A. Basin, Peninsular Ranges, and western Transverse Range block; however, along the coastline, they did not analyze many of the stations we were able to use. In fact, the only relevant stations we have in common along the coast are SBC, CIA RPV, SCI (on the same island as SCI2), and SNCC. We also compared our VTV with theirs as our control station. Their observations show a similar crustal structure, which gets thinner as it proceeds from the L.A. Basin and Peninsular Ranges into the Inner

Table 3.5: Comparison our crustal thickness and Vp/Vs ratio estimates with those from Zhu and Kanamori (2000) (Z&K) and IRIS EARS.

Station	This Moho	Study σ_H	Z&K Moho	Z&K σ_H	IRIS Moho	EARS σ_H
BAR	33.7	2	34.2	1.6	40	2.4
CIA	21.9	2	22	1.7	24	1.1
JCS	36.3	3	34.9	1.3	36	1.6
PAS	24.4	6	28	1	25	1.2
PHL	25.6	2	24.3	1.1	24	0.2
PLM	38.3	5	34	0.8	29	1.2
PLS	28.9	1.5	28	0.7	N/A	N/A
RPV	22.3	1.5	21.5	0.7	20	0.3
SBC	29.8	2.5	33.3	1.5	28	5.3
SCI(2)	23.4	1.5	21.8	0.5	28	7.5
SNCC	31.5	3	21.1	0.9	32	8.1
VTV	32.4	1.5	30.9	0.9	32	1.7

Borderland. While we both found similar results for CIA, RPV, SCI(2), and VTV, they found a very slightly thicker crust at SBC (33 vs 29.8 km) and significantly thinner crust at SNCC (21.1 vs 31.5 km). The fact that signal to noise ratios are highest at CIA, RPV, SCI(2), and VTV may account for our agreement at these stations, as clear signals make for easily interpretable H- κ stacks. This is because Zhu and Kanamori (2000) used a much smaller dataset than is available to us thirteen years later. At SNCC the differences may be due to misidentification of shallow crustal multiples. It should be noted that Zhu and Kanamori (2000) only had three island stations available at this time, compared to the seven used in this study.

IRIS EARS also only takes advantage of three of the island stations and some of the coastal stations, but the values they observe in the three islands we have in common are comparable to the observed values in this study. IRIS EARS (Crotwell and Owens, 2005; Trabant et al., 2012) did find similar (within 3 km) results for

CIA, FMP, LGU, SBC, SNCC and VTV even though it assumed a V_p of 6.048 km/s (instead of 6.4 km/s, which should make their estimates shallower). However, IRIS EARS found a thicker crust under SCI2 (28 km, almost exactly the same as our alternative depth) and a significantly thicker (44 km) crust under SCZ2. The IRIS EARS values show generally similar continental crust along the coastline (LGU, FMP, SBC) that thins by $\sim 30\%$ in the Inner Borderland and then thickens to values comparable to the coastline in the Outer Borderland.

We dedicated relatively more analysis to SNCC (see Figure 3.6) than any other station because it is the only island station in the Outer Borderland, and provides the vast majority of the data for this structural block. The ALBACORE stations in this region that returned usable data (L06 and S26) had very high noise and not enough waveforms to allow high confidence interpretation. Therefore, it was crucial for the interpretation of H- κ stacks of SNCC to be reliable in order to quantify differences between the crustal structure of this and surrounding blocks, which may help to understand the history of deformation in this region. In order to ensure that our conclusions were self-consistent and reproducible across different methods, we analyzed SNCC using Ps and Sp receiver functions, as well as Ps and Sp H- κ stacking.

A previous study (Zhu and Kanamori, 2000) concluded that the Moho in the Outer Borderland was $21.1 \pm .9$ km with a V_p of 6.4 km/s and a $\frac{V_p}{V_s}$ ratio of 1.74. The result of IRIS EARS (Crotwell and Owens, 2005) changed during the course of our study, but as of March, 2014, was a crust of 32 ± 8.1 km with a V_p of 6.048 km/s and a $\frac{V_p}{V_s}$ ratio of 1.6. The parameters chosen by Zhu and Kanamori (2000) appear

to be the best choice when first looking at the receiver functions and H- κ stacks for this station, because they fall on a prominent peak in the Ps receiver function and the H- κ stack (see Figure 3.6). However, SNCC receiver functions unequivocally show a strong conversion (both Ps and Sp) at small lag times, consistent with a thin, shallow crustal layer. When H- κ parameters are put in for this shallow (most likely sedimentary) layer, it becomes apparent in the receiver functions and H- κ stacks that the first multiple from the mid-crustal phase (the strongest positive phase) arrives at the same time as the direct conversion from the ostensible Moho at around 21 km depth (see Figure 3.6). Furthermore, the second multiple from the shallow layer arrives at the same lag-time as the third positive phase. This second multiple (PpSs+PsPs) from the Ps conversion in the mid-crustal layer will have a negative amplitude and therefore reduce the apparent amplitude of the positive third arrival, resulting in weak signal. Based on this analysis, we believe that the third positive arrival is likely the signal of the actual Moho beneath SNCC.

The H- κ product stacks (see Figure 3.6) support our parameter choice because no peak is present for the Zhu and Kanamori (2000) choice of crustal parameters. This absence implies that the sign of the Ps receiver function is opposite to that predicted for crustal multiples by their crustal thickness and Vp/Vs ratio. On the other hand, a small, but perceptible peak is present in both the sum and product H- κ stacks for our choice of crustal parameters. We further verified our results by analyzing the Sp receiver function, which is not affected by multiples from shallow structures, and should therefore highlight the true Moho. Indeed, a clear positive phase is present in the Sp receiver function at a depth of ~ 34 km, which is compati-

ble with, though slightly deeper than our preferred Moho depth derived from the Ps H- κ stacks. The small discrepancy between the Sp and Ps results falls within our uncertainty estimate, and may be due to damping in the frequency domain deconvolution skewing to a slightly deeper Moho, or simply the effects of higher noise levels in the Sp results. Our final test for validating the interpretation of crustal structure at SNCC involved incorporating Sp receiver function within the H- κ stacks. These stacks confirm that no peak is present for a crustal thickness of 21 km. Instead, a clear peak is present for a crustal thickness of 30.0 km and a reasonable $\frac{V_p}{V_s}$ range (1.7-1.9), in agreement with the preferred crustal structure parameters. We also use the correlation of the stations to support these results, both qualitatively and quantitatively. The pattern of three positive peaks with the third peak being the Moho or one large sedimentary peak and then another peak at a similar depth is prevalent in the Peninsular Range and southern coastal stations. The quantitative correlation that minimized misfit to choose which stations were the most similar also put SNCC with the Peninsular Ranges and southern coastline (as well as SBI).

Figure 3.6: The left panels show the Ps and Sp receiver functions for SNCC, with the dashed line going through the Moho selection. The six panels on the right represent the H- κ stacking done for $\frac{V_p}{V_s}$ ratio and Moho constraints with an assumed Vp of 6.4 (Zhu and Kanamori, 2000) (top left) Ps with the sediment and multiples represented to show that the parameters selected in previous studies (Zhu and Kanamori, 2000) fall on the sedimentary multiple, adding to the strength of this phase, while the second Ps multiple arrives at the closely to our Moho, decreasing its amplitude. (middle and bottom left) An H- κ stack with Sp included in the stack (with parameters reduced to avoid selection of sediment), so that the phase is consistent across all of the phases used for the H- κ stack and it can be clearly seen that the phase chosen (red phase) in previous studies (Zhu and Kanamori, 2000) has no positive signal in the Sp stack. (top right) shows the product of the H- κ stack without the Sp phase taken into account. It should be noted that the parameters used in previous studies do not exist here. (middle right) the sum of the H- κ stack, which was the only method used previously and the previous parameters chosen do exist here. (bottom right) the product of the Sp H- κ stack (focused again to avoid the sediment), which shows that the phase chosen in this study is the only one that exists with reasonable parameters for a $\frac{V_p}{V_s}$ range and Moho depth.

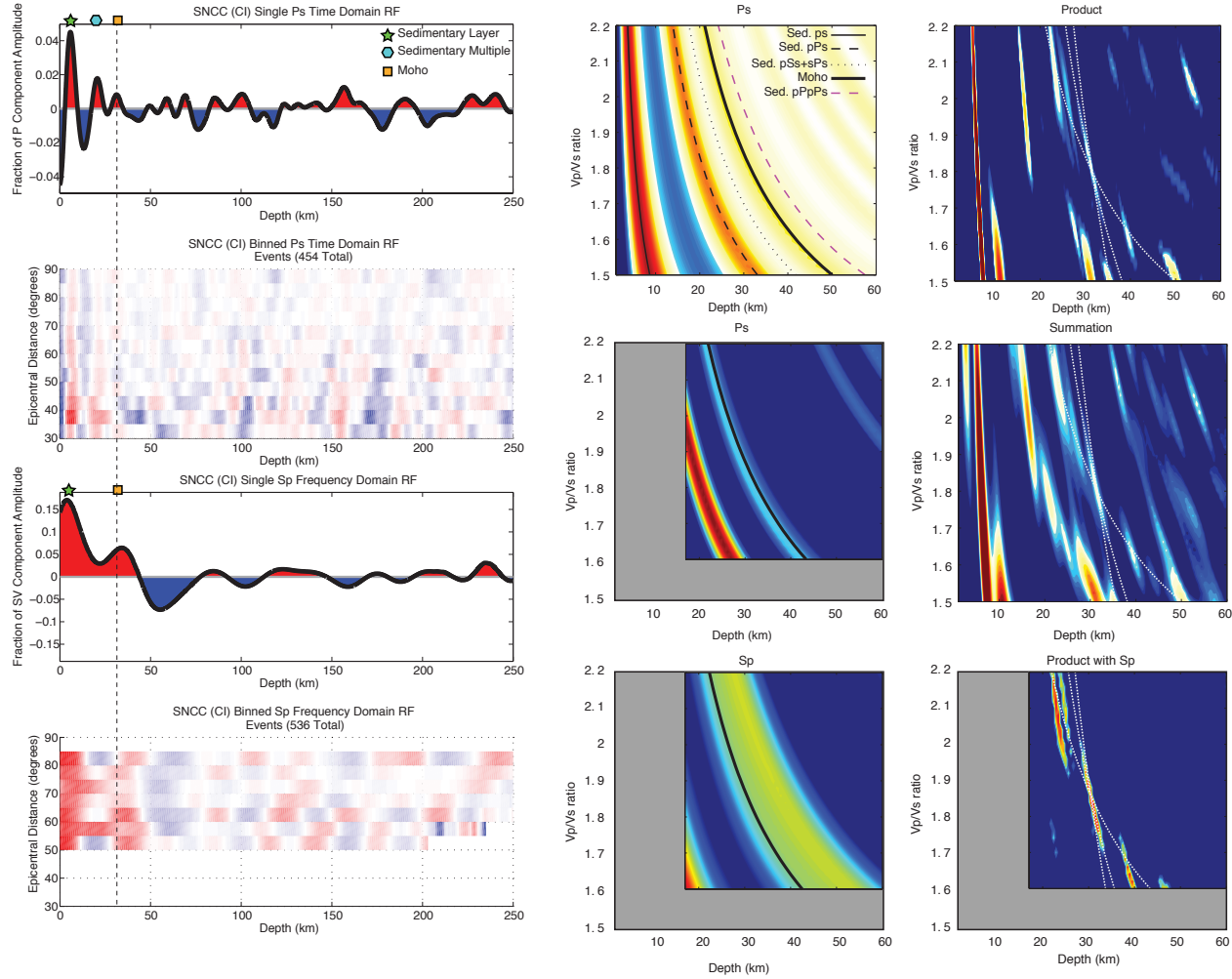


Figure caption on preceding page.

This correlation between SNCC and the Peninsular Ranges suggests that the Outer Borderland was translated and slightly rotated, as opposed to being rifted and thinned to compensate for the rotation of the western Transverse Ranges because of the similarity in thickness. The varying thickness of the mid-crustal layers between SNCC, the southern coastal and Peninsular Range stations can be explained by the geography in each region (mountains, base of mountains, island). The fact that we used multiple methods gives us confidence to give different results than previous studies obtained. It is likely that substantial extension played a key role in the formation of the Continental Borderland because of the significant crustal thickness difference (~ 22 percent) between the Inner and Outer Borderland and the Inner Borderland and WTR block (Figure 3.7). There is a ~ 50 percent difference in thickness between the Peninsular Ranges and the Inner Borderland. These findings can be interpreted as supporting the Crouch and Suppe (1993) model for the formation of the Inner Borderland because this model relies heavily on extension and the uplift of a metamorphic core complex beneath it.

It should also be noted that there is a relatively rapid change in crustal thickness between stations FMP/RPV and the L.A. Basin. Interestingly, this abrupt change agrees with previous analysis of two completely independent datasets, which argue for a rapid transition between the Inner Borderland and the L.A. Basin near the Newport-Inglewood Fault Zone (Nazareth and Clayton, 2003; Schmandt and Clayton, 2013). In one study, high density array data allowed Schmandt and Clayton (2013) to use teleseismic imaging to show a very sudden boundary change to the Inner Borderland rifting and a Moho dip of $\sim 65^\circ$ at this boundary. Nazareth and

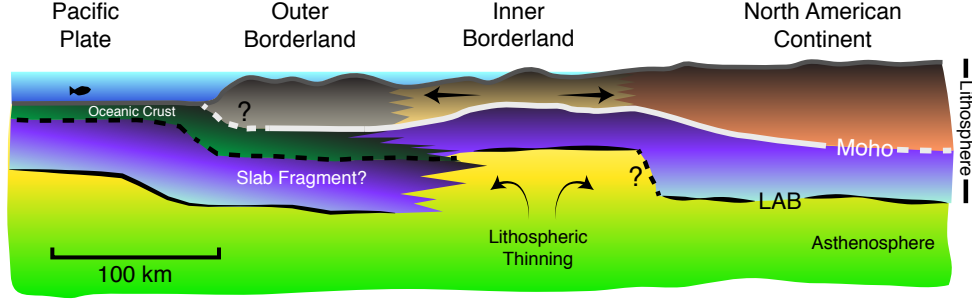


Figure 3.7: A possible interpretation of the borderland based on cross-section C of Figure 3.5. The Inner Borderland has been formed from rifting and thinning of the lithosphere, while there is possibly a slab fragment of oceanic lithosphere that allowed the Outer Borderland and western Transverse Range block to be translated/rotated without substantial thinning.

Clayton (2003) used a continuous Pn phase to show that if the crust in the Inner Borderland is 22 km thick, which it appears to be in our study, the transition zone is constrained to initiate within a 2 km wide region beneath the southwest L.A. Basin, and have a width of 20-25 km.

It is possible that the positive phase observed beneath the Outer Borderland in Sp CCP stacks (Figure 3.5) can be due to a fragment of trapped oceanic lithosphere that was partially subducted and then proceeded to translate with the overriding plate, but this interpretation is preliminary and warrants further investigation. We attempted to ascertain the robustness of this phase by constructing CCP stacks using different subsets of the data.

3.4.3 Mid-crustal Discontinuity

At some, though not all stations, we observed a Ps conversion across a mid-crustal discontinuity. The depth and average $\frac{V_p}{V_s}$ ratio associated with this discon-

tinity varies systematically across our region. Because sedimentary basins tend to have higher $\frac{V_p}{V_s}$ ratios than most igneous rocks, we can use the maps in Figure 3.4 to determine in which areas the mid-crustal discontinuity is likely due to a sedimentary layer, and in which areas is it more likely to result from other structural or lithological interfaces. Because the $\frac{V_p}{V_s}$ ratios of the mid-crustal phase observed at stations in the L.A. Basin tend to be high (>1.85) and the average depth of the discontinuity observed here is ~ 5.3 km (Table 3.4), we conclude that it occurs across the base of the sedimentary basin. Indeed, the average depth of the mid-crustal phase yields a reasonable average thickness for the sedimentary layer in the L.A. Basin, considering that the maximum depth of the sediment is ~ 10 km (Yerkes et al., 1965). The depth and V_p/V_s ratios of mid-crustal discontinuities detected at stations in surrounding areas and islands do not appear to be similar to the L.A. Basin because they generally have lower $\frac{V_p}{V_s}$ ratios (<1.85) and depths that range from 2 km up to 13 km. This suggests that the discontinuity observed in the islands, western Transverse Ranges and Peninsular Ranges is different than the L.A. Basin and could be some sort of lithological interface, as opposed to the base of the sediment. This conclusion is supported by previous studies. For example, Fuis et al. (2003) observed a reflector in the upper crust at a depth of ~ 3.5 km beneath the eastern portion of the WTR block, using Los Angeles Region Seismic Experiment 2 (LARSE-2) stations. He suggested that the reflector is caused by lithostatically-pressured fluids in pores and cracks, which may be released during earthquakes. The mid-crustal phase we observe in the same area (stations SMS, PDR, LCG, SMF2 and HLL) has a similar depth to this reflector (average of ~ 3.3 km). Further west, Namson and Davis

(1988) demonstrated that there is a fold system that has a detachment fault at a range of depths, which could be the large positive phase showing up in our receiver functions at stations WGR (7.2 km), SES (11.1 km), STC (8.8 km), MOP (13.2)km and TOV (4.3 km). Interestingly, we observe a mid-crustal phase at two of the three ALBACORE stations located in the abyssal plain but just west of the Patton Escarpment. The depth (~ 8) and $\frac{V_p}{V_s}$ ratios (~ 1.85) of these phases are consistent with oceanic crust, which may suggest that the deeper phase we identified as the Moho might result from a sub-crustal interface.

3.4.4 Lithosphere-Asthenosphere Boundary

Variations in the depth to the LAB can also provide information on the structure and history of deformation in this region. For example, Lekic and Fischer (2013) performed common conversion point (CCP) stacking of Sp receiver functions to estimate LAB depths in this region, but used a substantially smaller dataset than that available currently. Though Lekic et al. (2011) did analyze Sp receiver functions at almost all of the island stations that we use in this study, many additional waveforms from earthquakes that have occurred since that study are available to us. While that study provided constraints on LAB depth in the Inner and, to a lesser extent, the Outer Borderland, substantial gaps in data coverage prevented inferences on LAB structure in about half of the Inner Borderland, almost all of the Outer Borderland and some of the WTR block. Furthermore, no observations were made west of the Patton Escarpment, since those only became possible as

a result of the ALBACORE deployment. The areas that they did cover show an Outer Borderland with similar thickness (~ 70 km) to the coastline and a relatively shallow LAB (~ 50 km) in the Inner Borderland. Our results generally confirm the findings of Lekic et al. (2011), especially regarding the substantial lithospheric thinning beneath the Inner Borderland. However, we obtain a more detailed picture of structure beneath the Outer Borderland and confirm that the lithosphere of this block is very different, consistent with our interpretation based on crustal structure that the Outer Borderland has experienced a very different history of deformation.

West of the Patton Escarpment, Sp data density is insufficient to enable CCP stacking, so instead, we analyze single station Sp receiver functions. We find a clear, strong, negative phase, consistent with a velocity drop elsewhere interpreted as due to a seismic LAB, at stations L07, L08, L10, L11, L13, L15, L18 and L19. The depths of this phase, along with uncertainties estimated by bootstrapping, are shown in Table 3.6, plotted at the plate age from Müller et al. (2008). If the age and LAB depth for each individual station is considered in isolation, the relatively large uncertainties make it difficult to make any argument regarding how strongly the depth of the seismically-observed LAB depends on age. This is because, within uncertainty, some of our stations fall on the LAB depths predicted by both an age-dependent model (e.g. Kumar and Kawakatsu, 2011), as well as models in which LAB depth is weakly dependent on age (e.g. Evans et al., 2005; Karato, 2012). However, by taking the mean age and depth of all stations in the abyssal plain (Table 3.6), we can calculate the standard deviation of the mean (which assumes that uncertainty on the LAB depth measurement at each station is independent

Table 3.6: Estimates of plate age and depth to the LAB for select AL-BACORE stations. Uncertainty on average LAB depth is calculated assuming uncorrelated errors on individual station estimates.

Station	Age (Mya)	Depth (km)
L07	17.6	57 ± 20
L08	19.3	60 ± 13
L10	21.1	64 ± 15
L11	21.3	65 ± 15
L13	23.8	45 ± 17
L15	25.5	51 ± 16
L18	24.1	66 ± 17
L19	22.6	56 ± 15
Avg.	21.9 ± 2.6	58 ± 2.93

of the others) and reduce the uncertainty substantially. By taking the standard deviation of the mean, it becomes apparent that the LAB depths we observe are similar to those found beneath much older areas (e.g. Kumar and Kawakatsu, 2011; Schmerr, 2012), whereas age-dependent models predict a significantly thinner ($\sim 30\%$) lithosphere for our relatively young ages (not within two standard deviations). Therefore, the inferred LAB depths in the Pacific abyssal plain, while not disproving models in which the depth of the seismically observable LAB, do not seem to support these models (e.g. Kumar and Kawakatsu, 2011).

Instead, our results are more compatible with models in which dehydration controls the depth of the seismic LAB (e.g. Evans et al., 2005; Olugboji et al., 2013). Our findings are particularly similar to those of Evans et al. (2005), who used images of the conductivity structure near the East Pacific Rise to claim that the oceanic lithosphere should be ~ 60 km thick everywhere because its thickness is dependent on dehydration due to melting of dissolved hydrogen olivine, rather

than age. They are also compatible with a model in which the seismic LAB is due to anelastic effects resulting from elastically accommodated grain boundary sliding (EGBS) controlled by hydration (Karato, 2012). While this suggests that there is not a relationship with age and lithospheric thickness, it should be noted that this is a small sample size (8 stations) from a small range of ages. To have confidence in these results, there need to be more ocean bottom stations deployed and observed over a wide range of ages.

3.5 Conclusion

The goals of this project were to develop a method for using low signal-to-noise OBS data applicable to future OBS experiments, better understand large scale relationships between oceanic and continental crust and/or lithosphere, give evidence that supports or disputes a correlation between lithospheric thickness and age, observe depth to the LAB variations between oceanic/continental conditions, and map variations in crustal thickness in order to draw interpretations regarding crustal deformation history in this region. We were able to use ocean bottom data for receiver functions by selecting the best waveforms (clear P, S, and surface wave arrival) and bandpass filtering these waveforms from .03-12 Hz. This method should be applicable to future OBS experiments and there is room for improvement if a transfer function could be developed that takes both, the vertical and horizontal components into account. We were able to observe what seems to be a fairly typical interaction between the oceanic and continental crust and lithosphere in subduction

zones where we observed the possibility of a captured piece of partially subducted oceanic crust, as well as some sort of high velocity anomaly at ~ 60 km depth in the Outer Borderland. Our small set of data is more consistent with a model where oceanic lithosphere thickness is not strongly correlated with age and is more likely related to dehydration, but this is not strong support for either model. We are able to conclude with reasonable certainty that the depth of the seismic LAB appears to be much deeper under the continent, shallower in the Inner Borderland, and a generally constant depth in the abyssal plain. The final conclusion we are able to make from this experiment is that with our updated receiver functions, the Outer Borderland appears to have been translated and slightly rotated to accomodate for the rotation of the western Transverse Range block, while the Inner Borderland was rifted. We were able to accomplish the goals set for this project, although we have higher confidence in some results (Moho depths) than others (abyssal plain LAB depths).

More studies of this region using OBS data would be beneficial to reducing uncertainties and having more confidence in the results found here. In particular, the southern Inner Borderland, and the whole Outer Borderland, would greatly benefit from some permanent OBS stations or, possibly, repeat occupation of a single site by temporary stations. While a substantial number of new constraints on the structure of the Continental Borderland and oceanic lithosphere were obtained in this study, the robustness and confidence can always be improved upon with more data. Indeed, the relatively short length of the ALBACORE deployment was the main hindrance in achieving high confidence and low uncertainties in this study. This is because

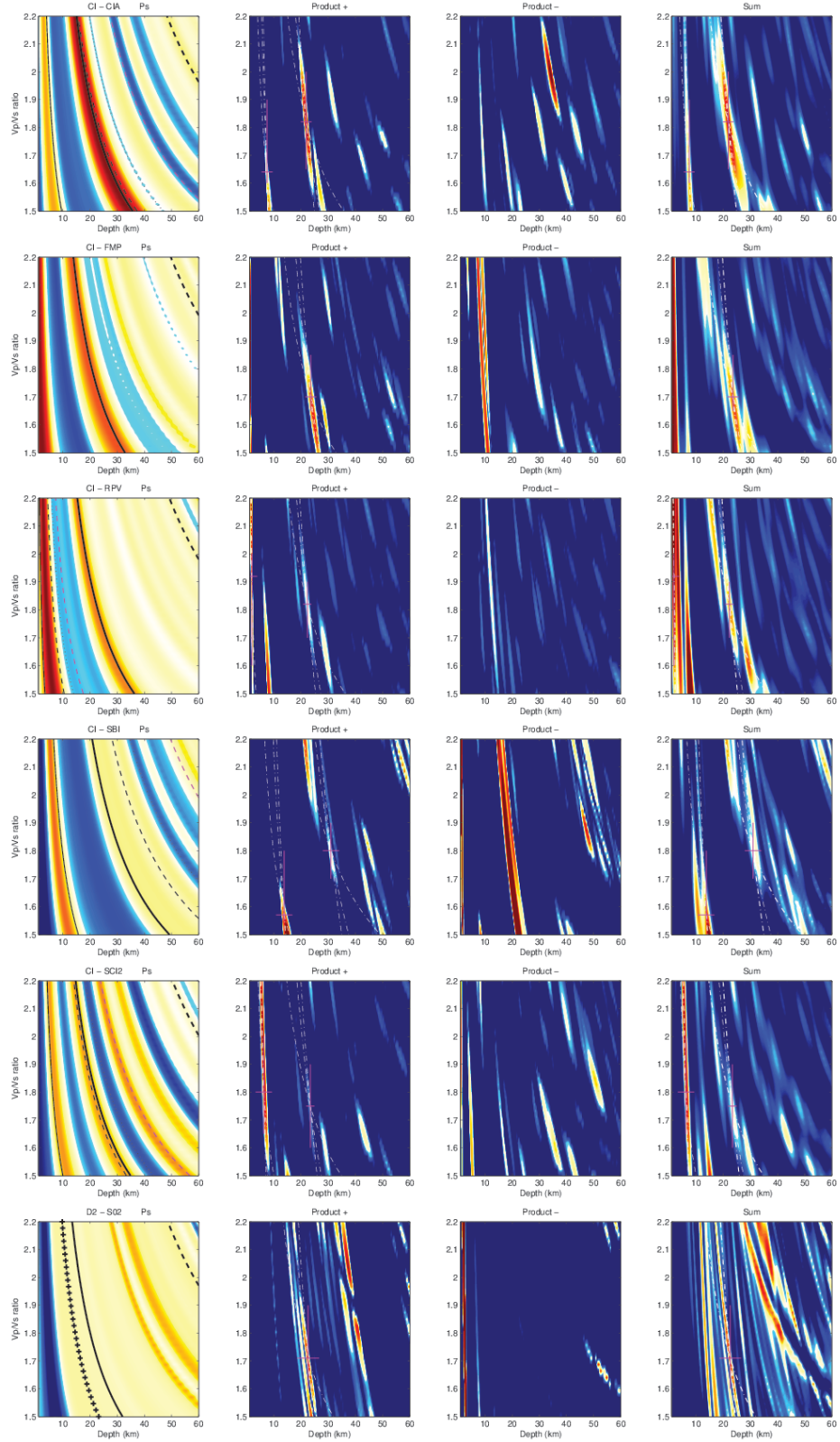
even if the quality matched that at the quietest land stations, it would still be hard to have confidence in receiver functions calculated from only a handful of events.

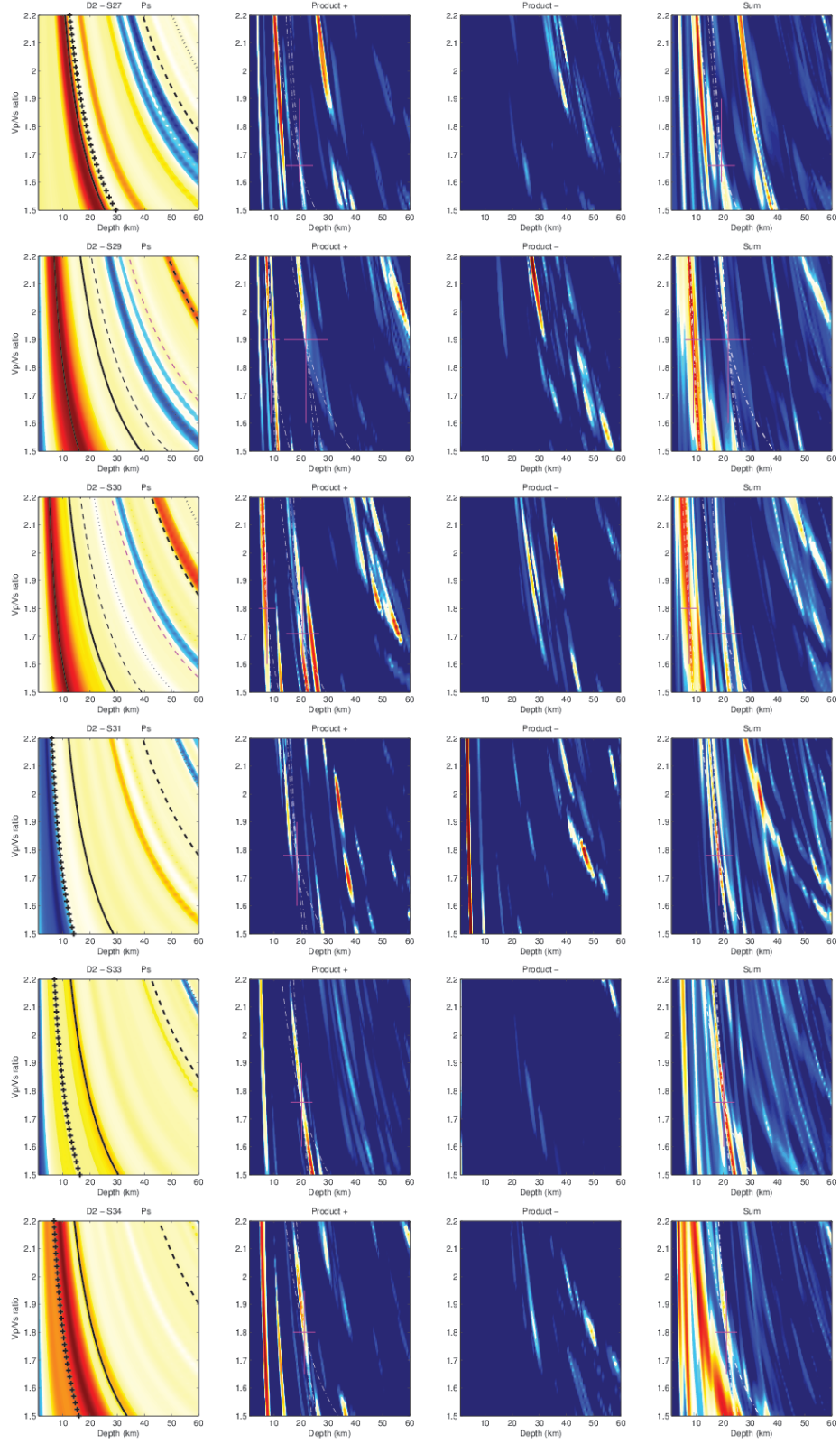
Chapter A: Appendix-H- κ Stacks

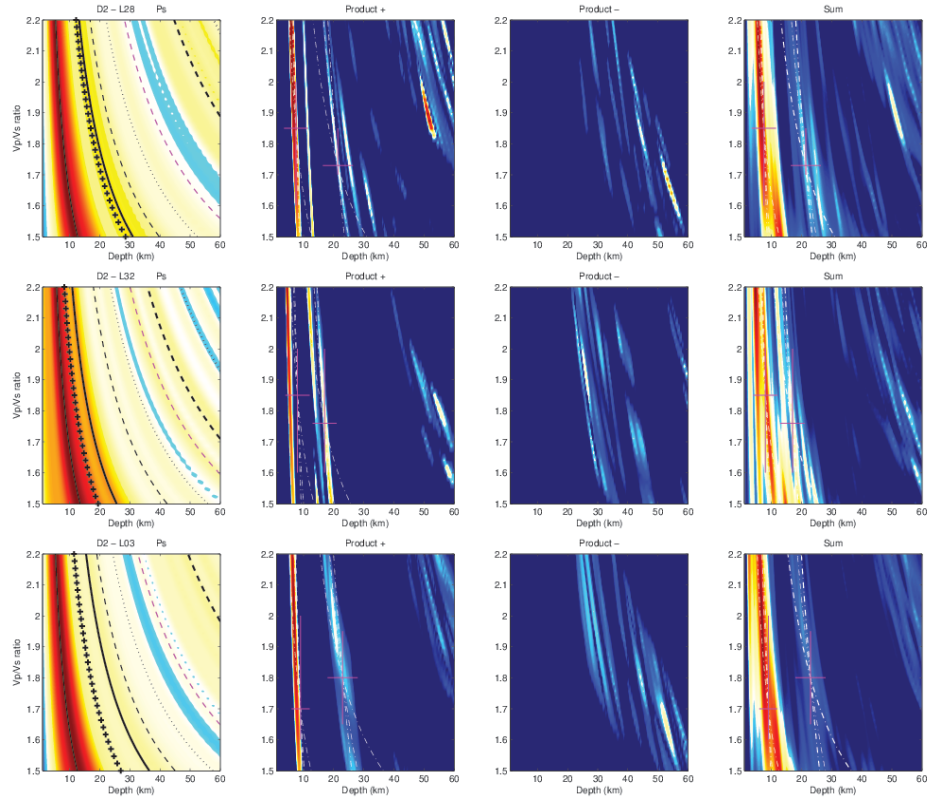
In this Appendix, we present the H- κ summation and product stacks for all stations in our study.

Inner Borderland

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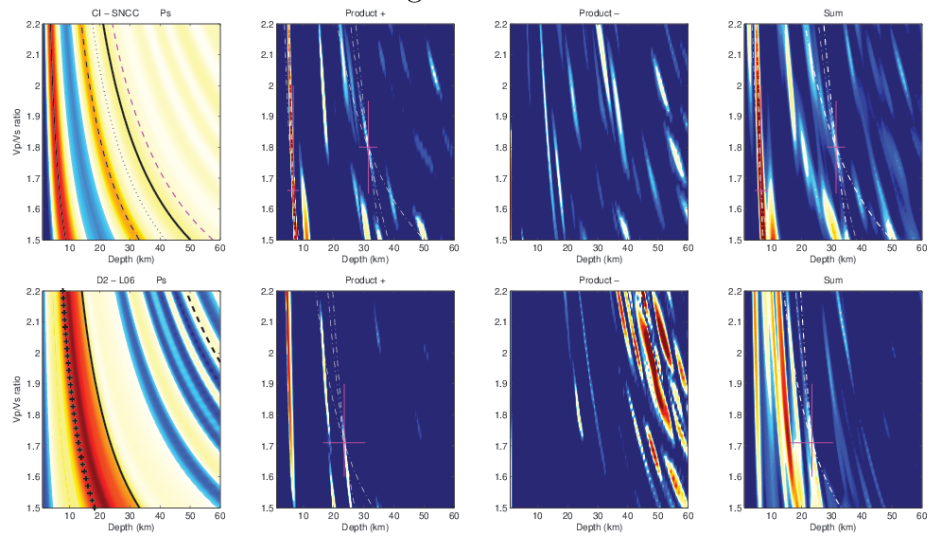


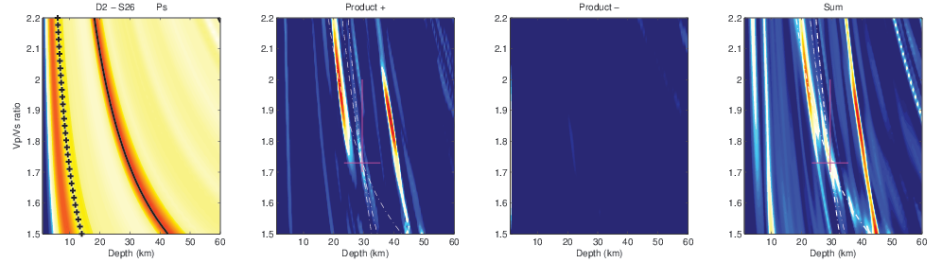




Outer Borderland

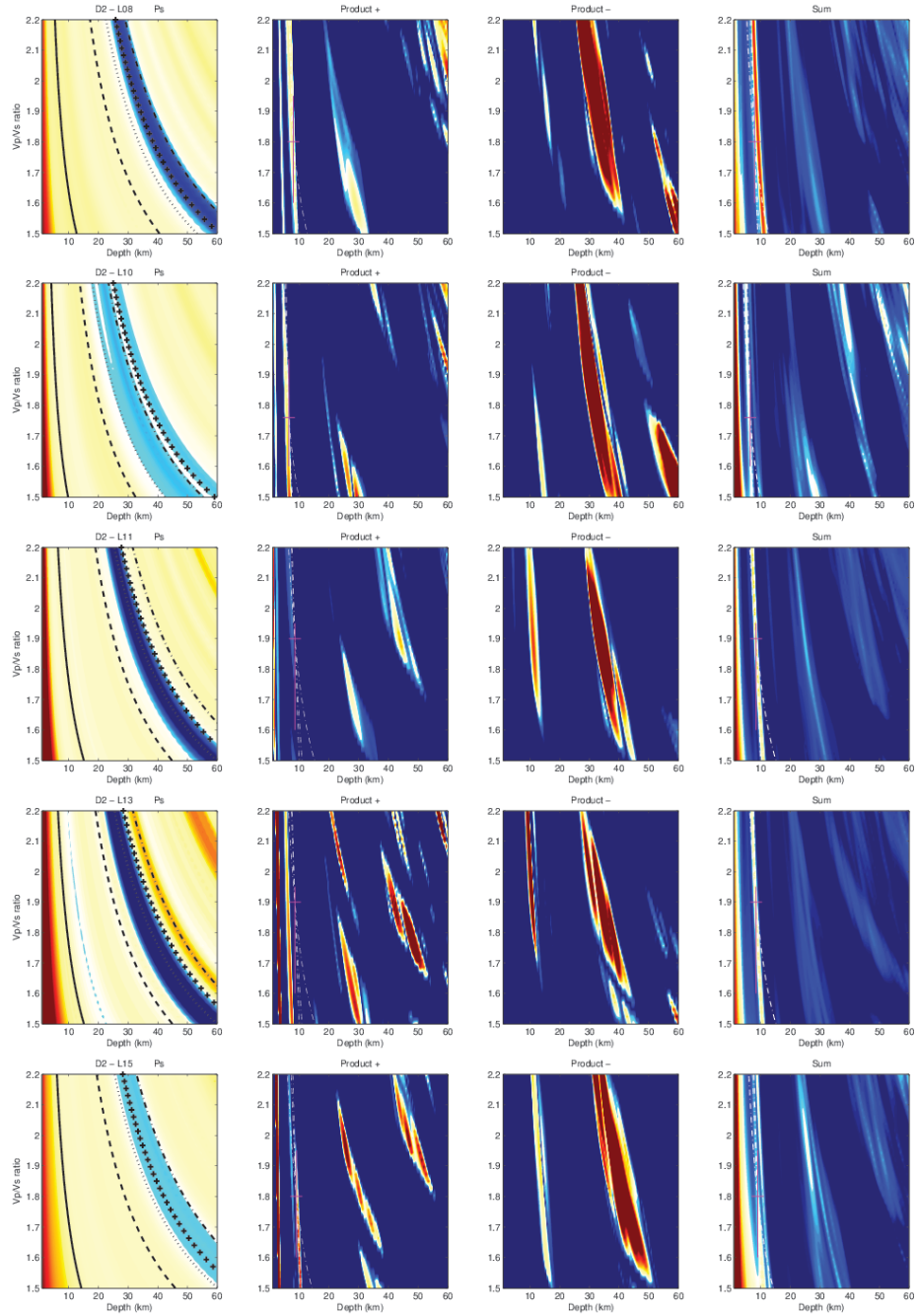
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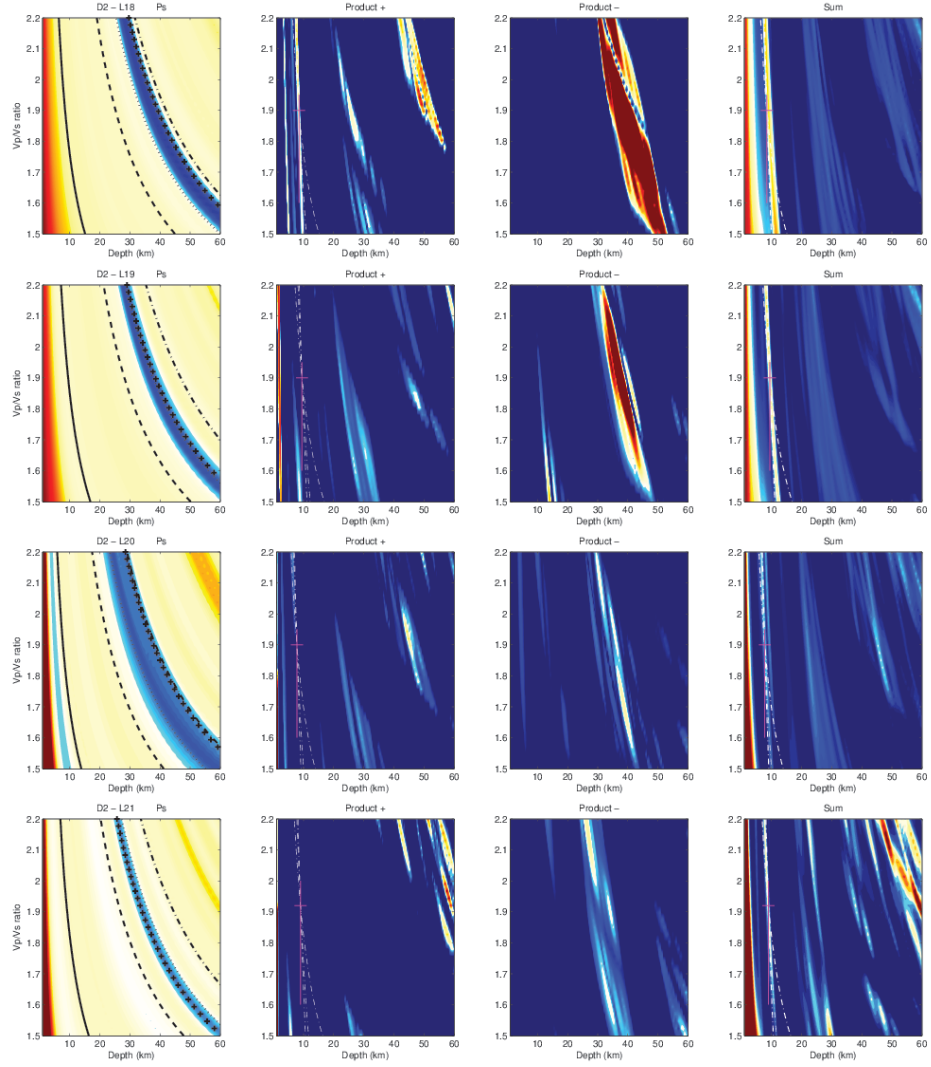




Abyssal Plain

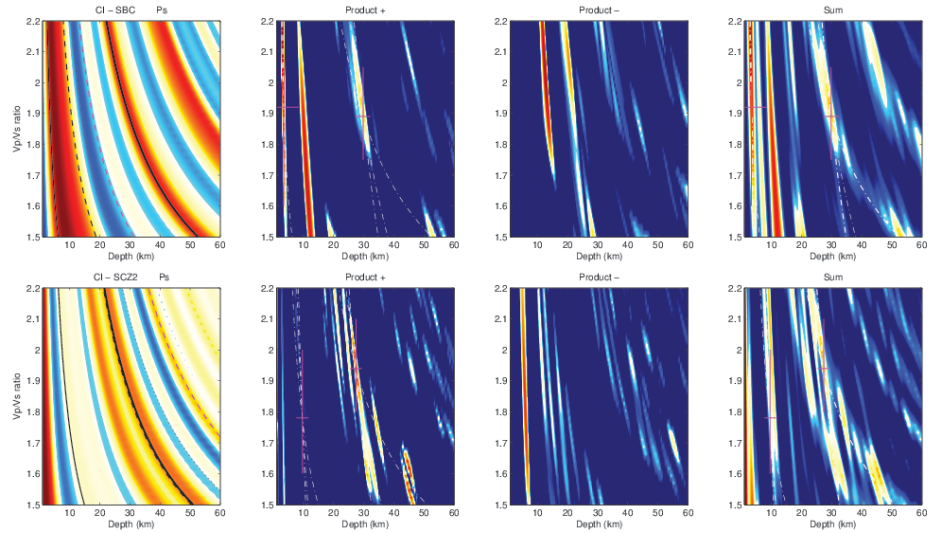
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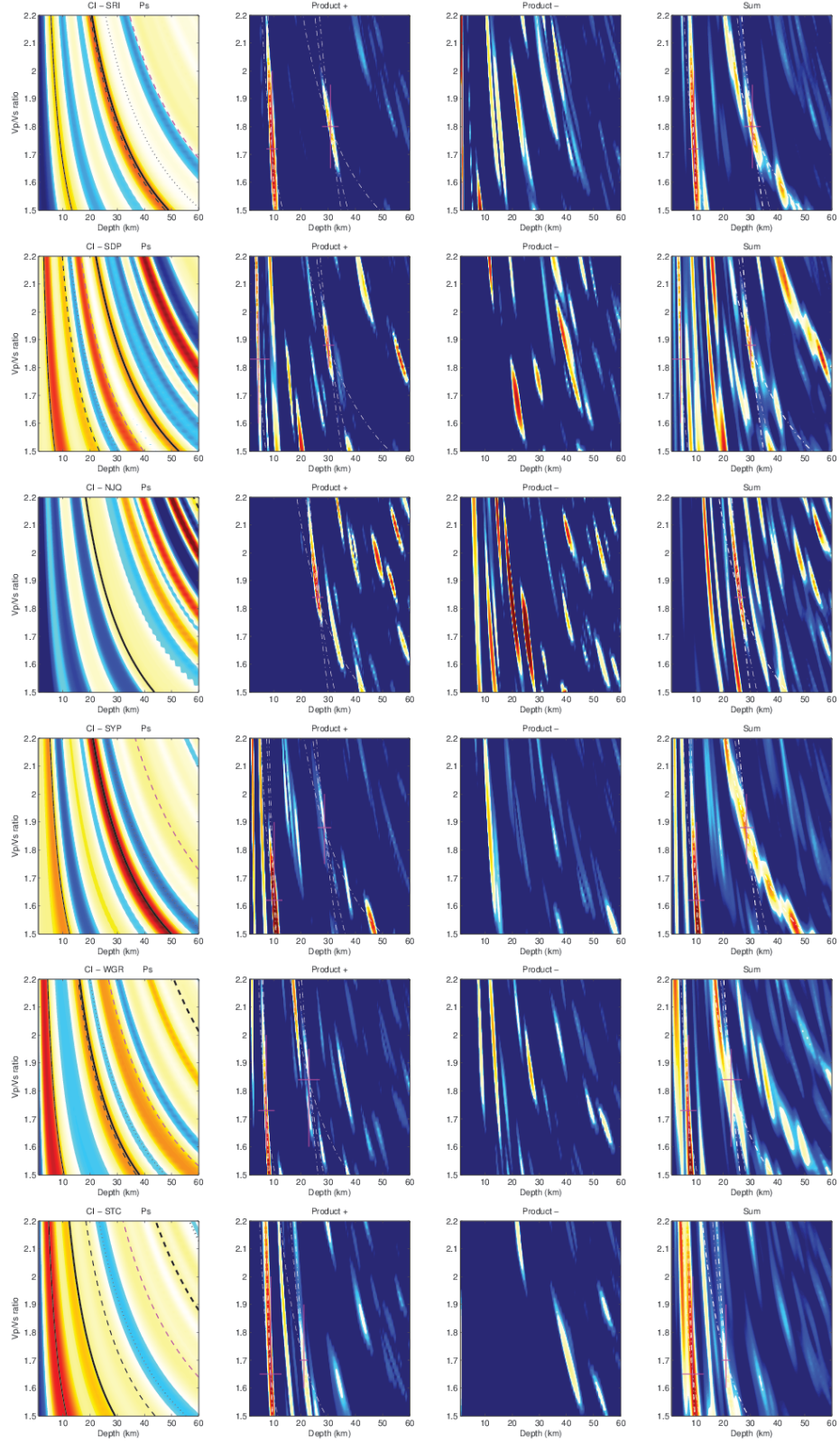


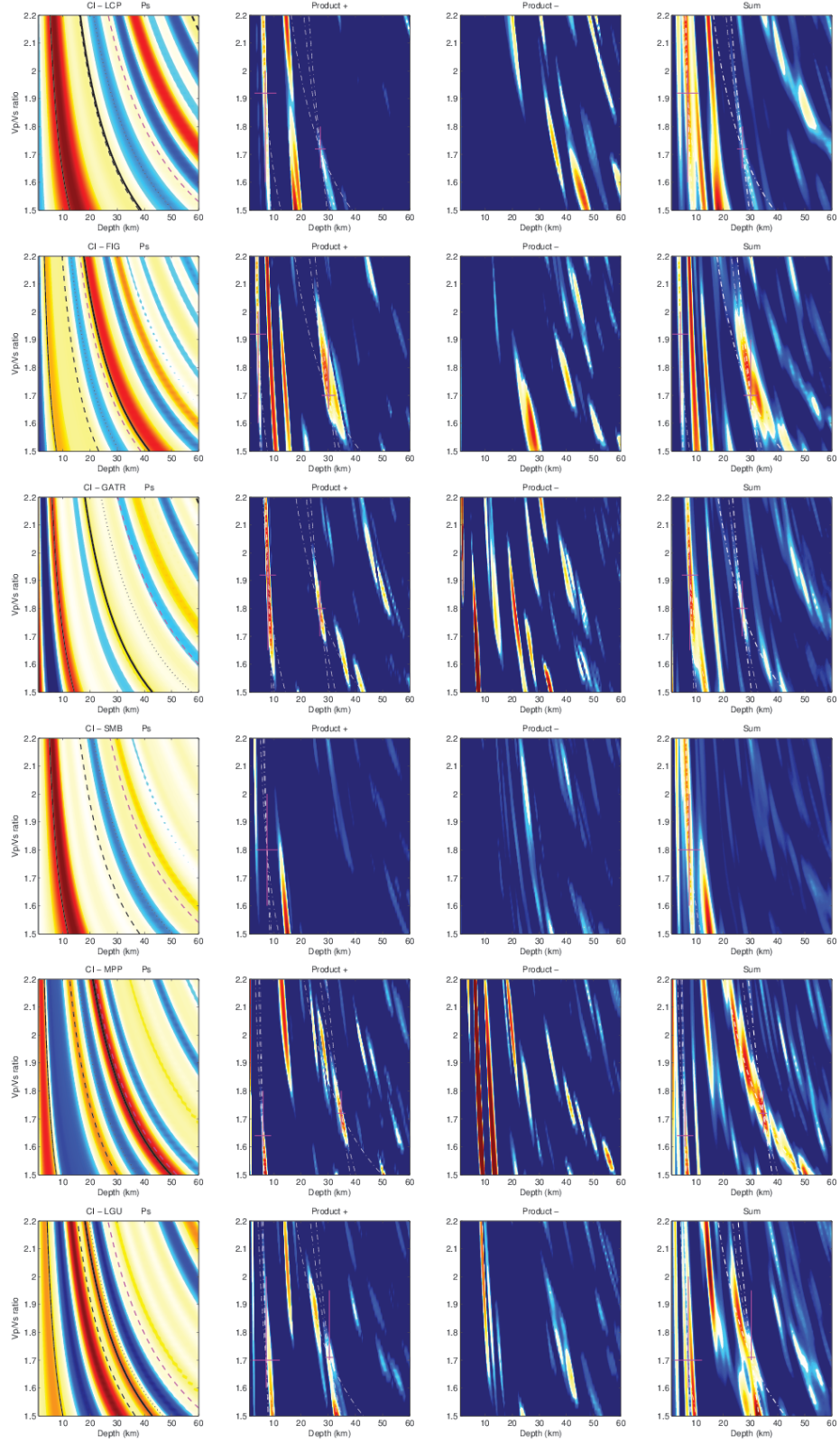


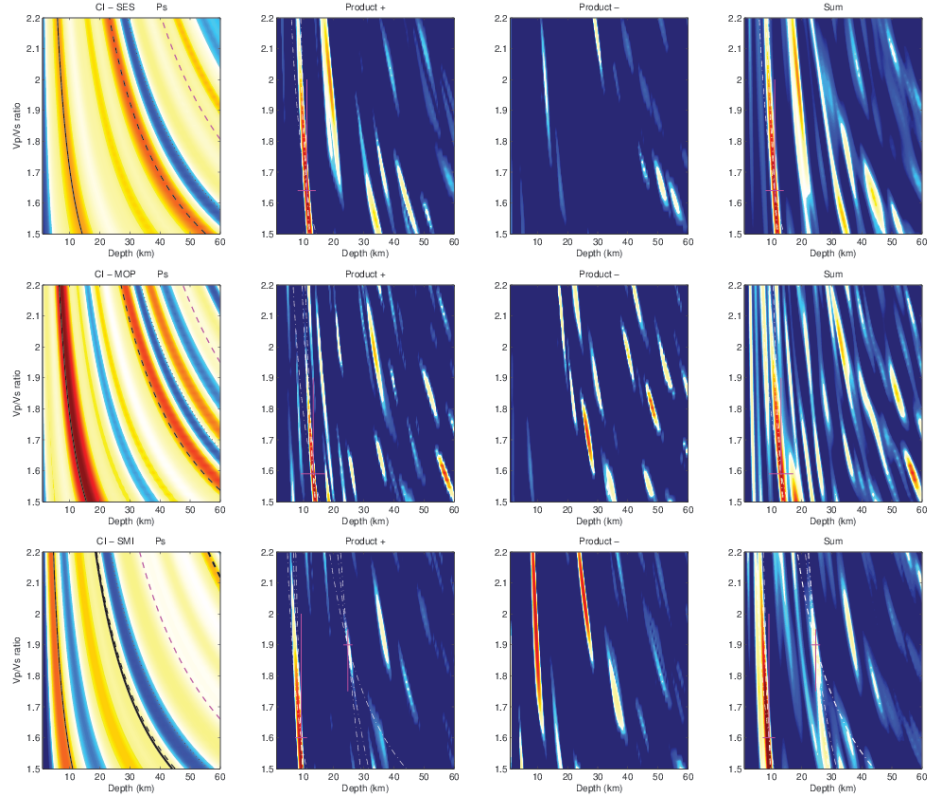
Western Transverse Range

Figure A.4:



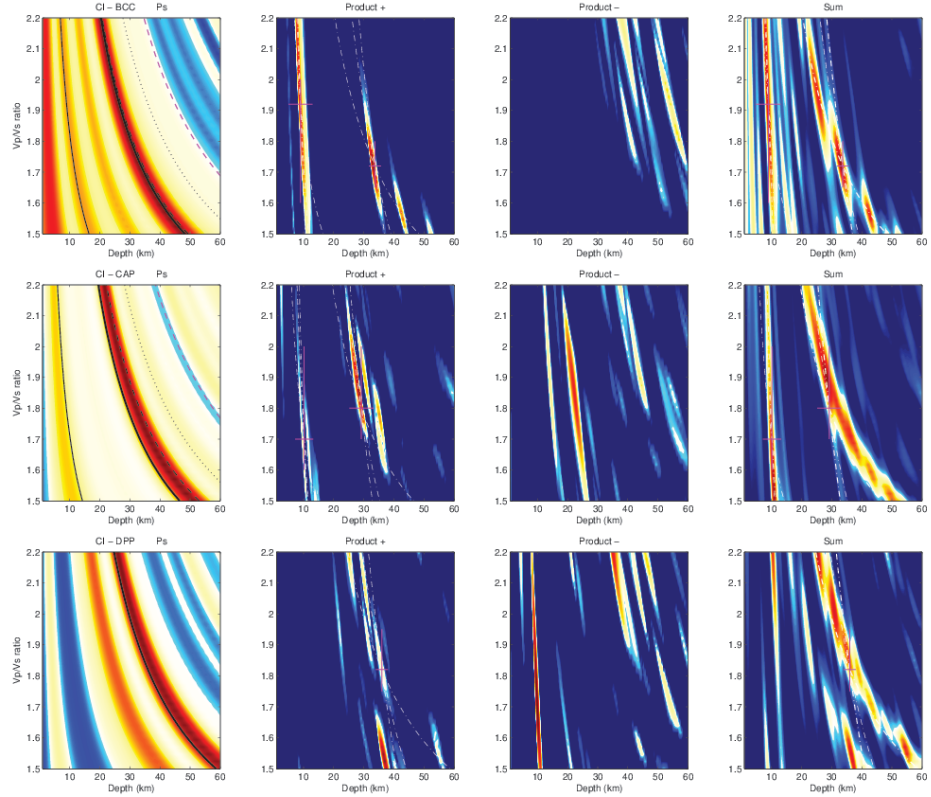


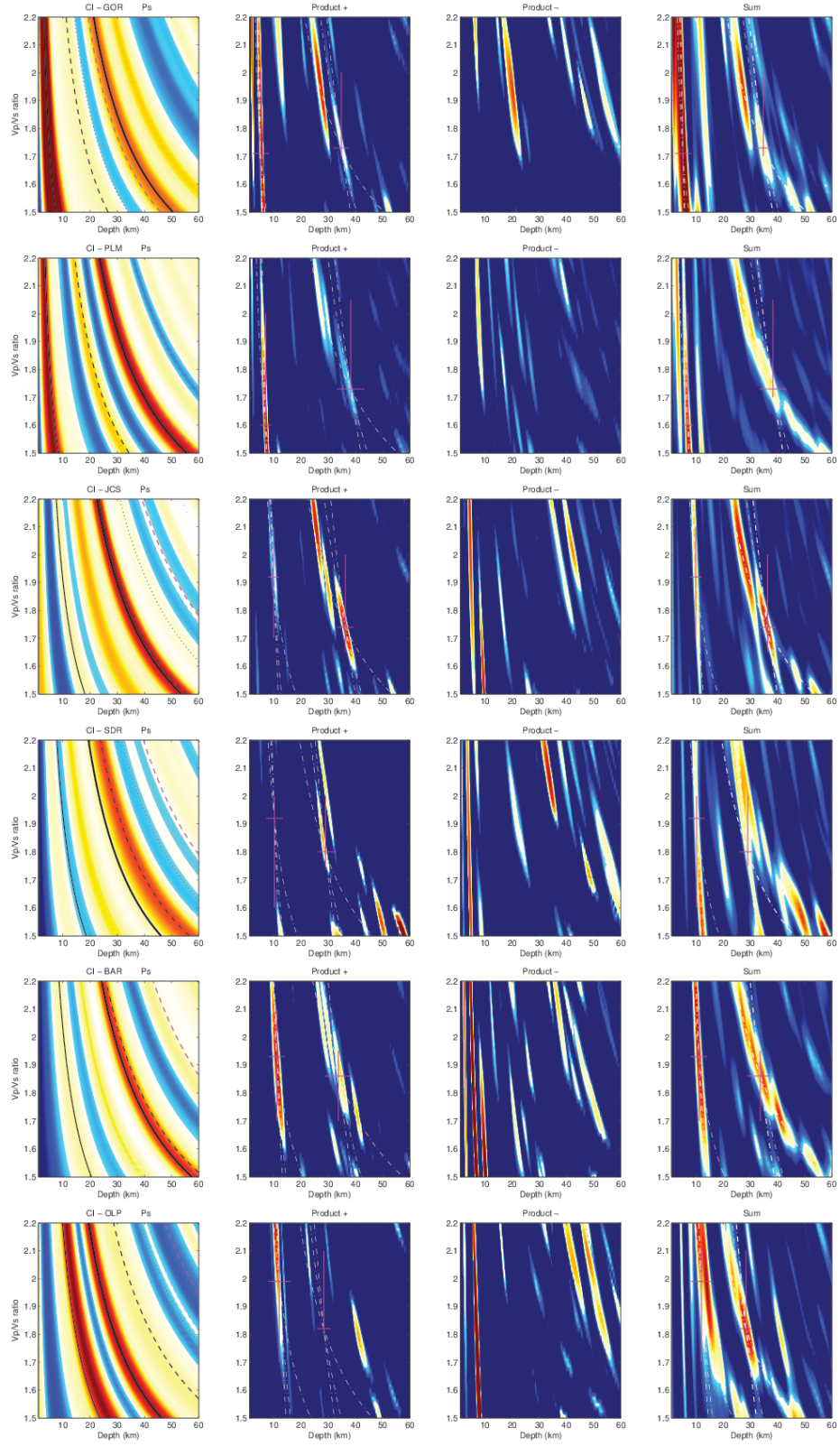


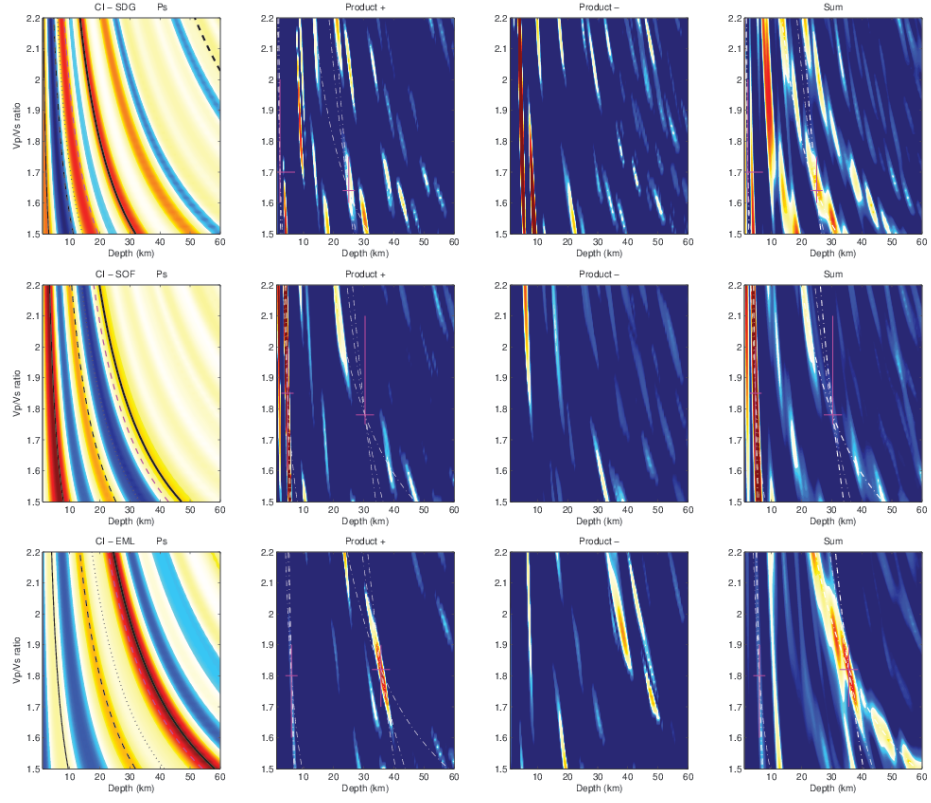


Peninsular Ranges

Figure A.5:

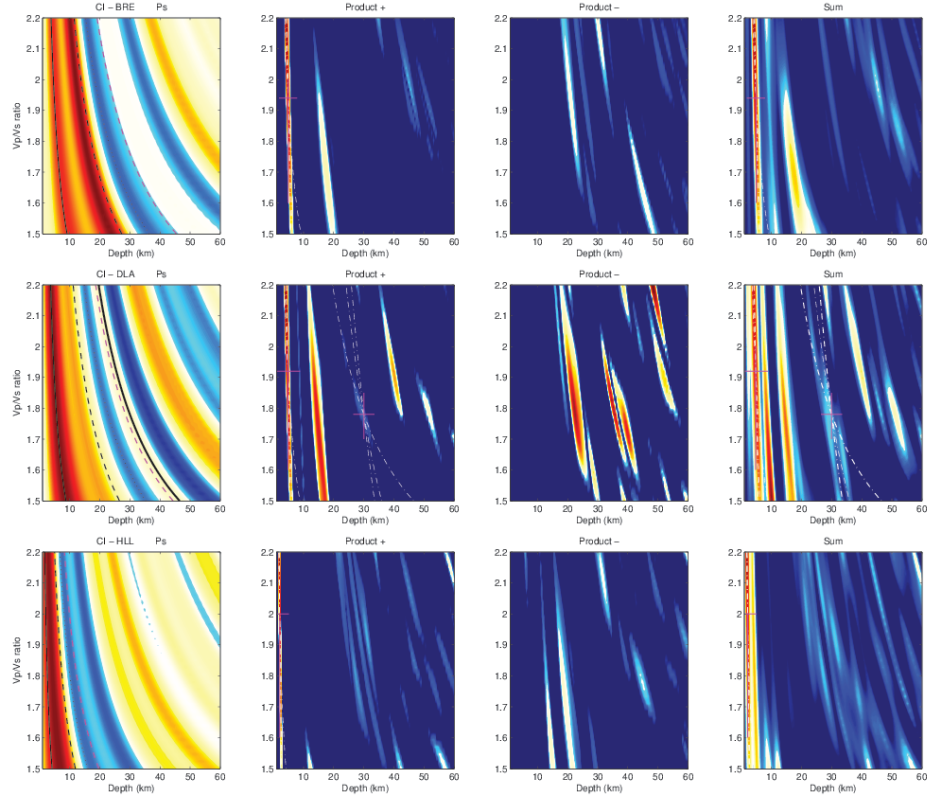


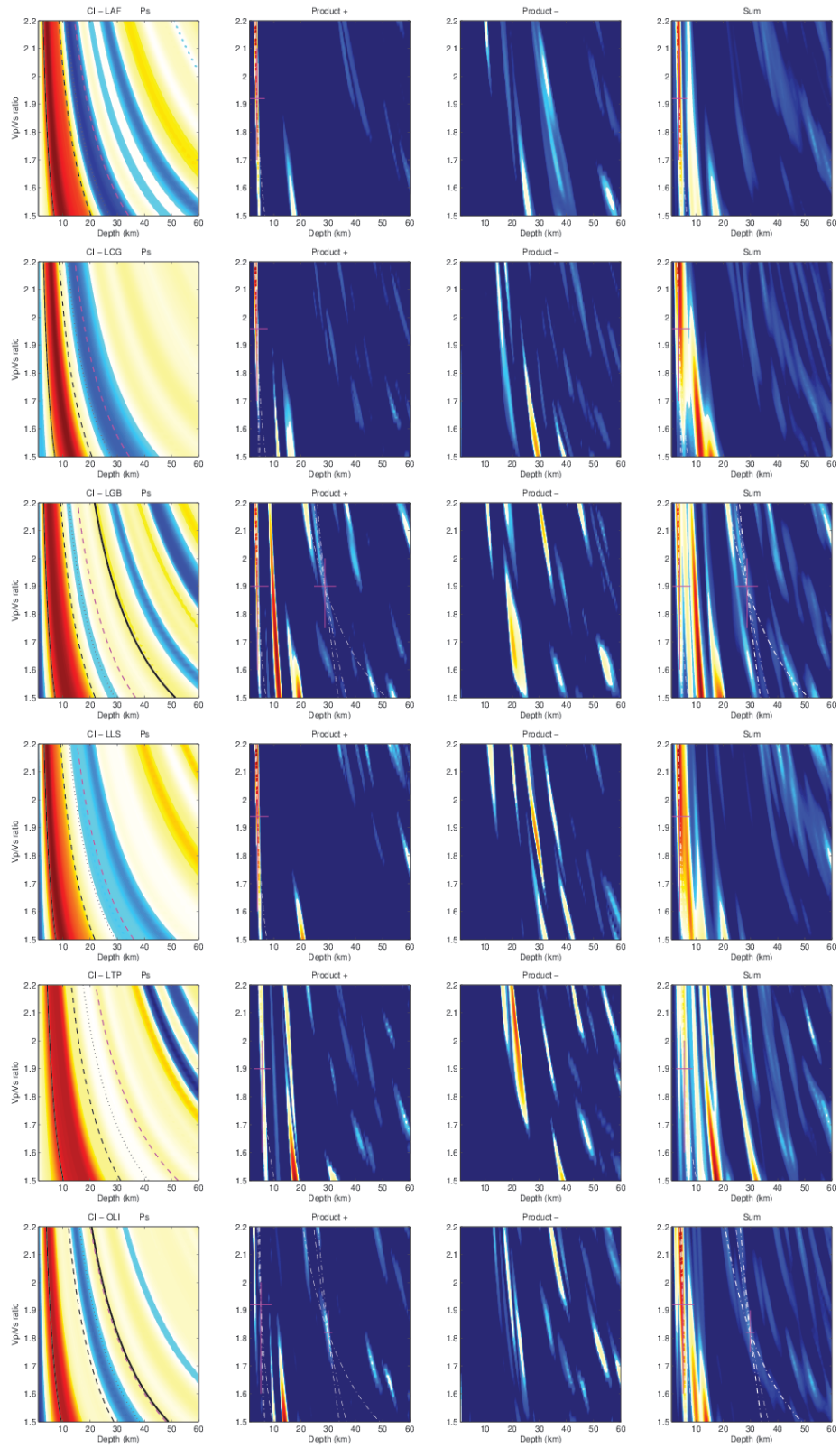


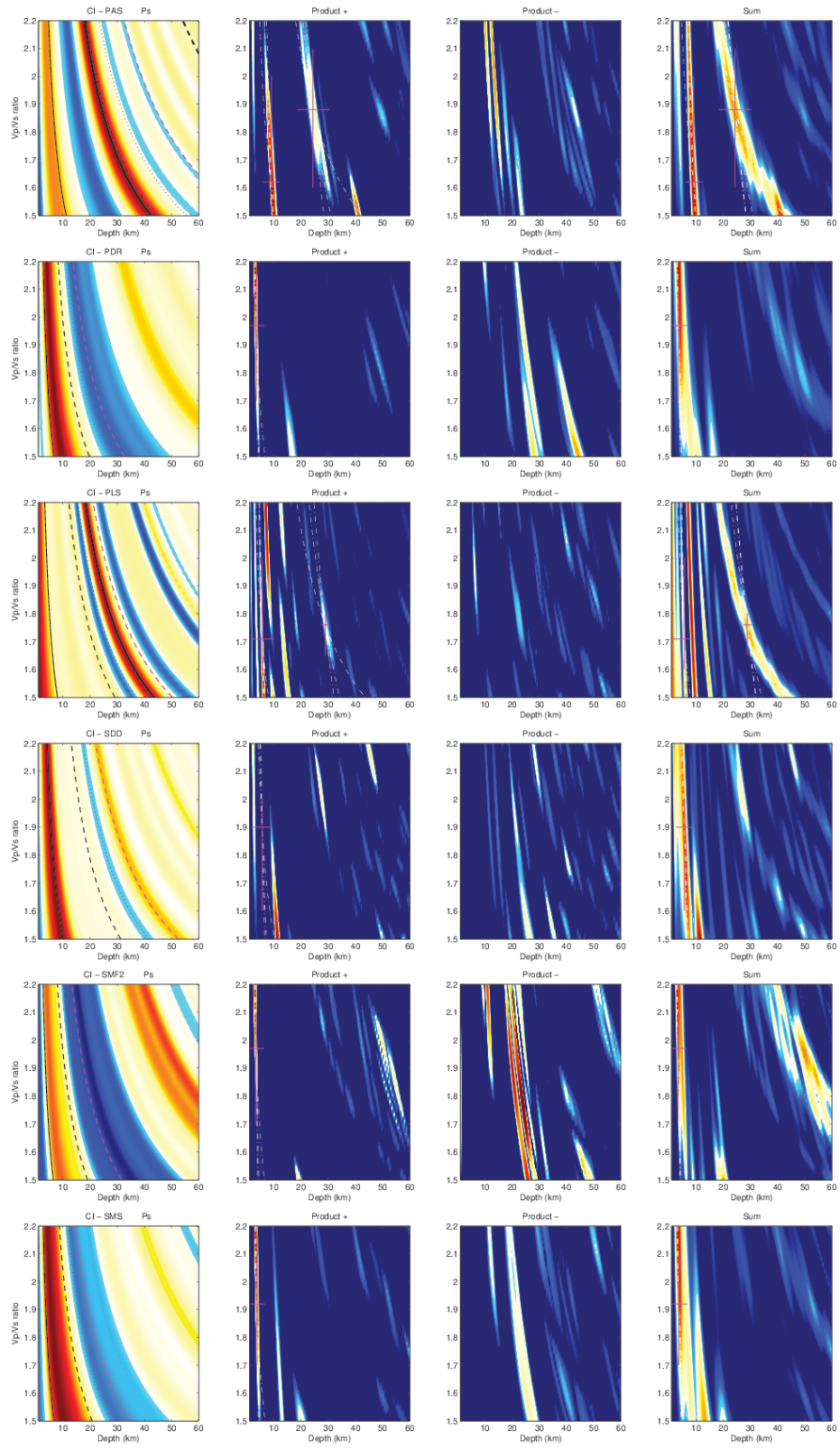


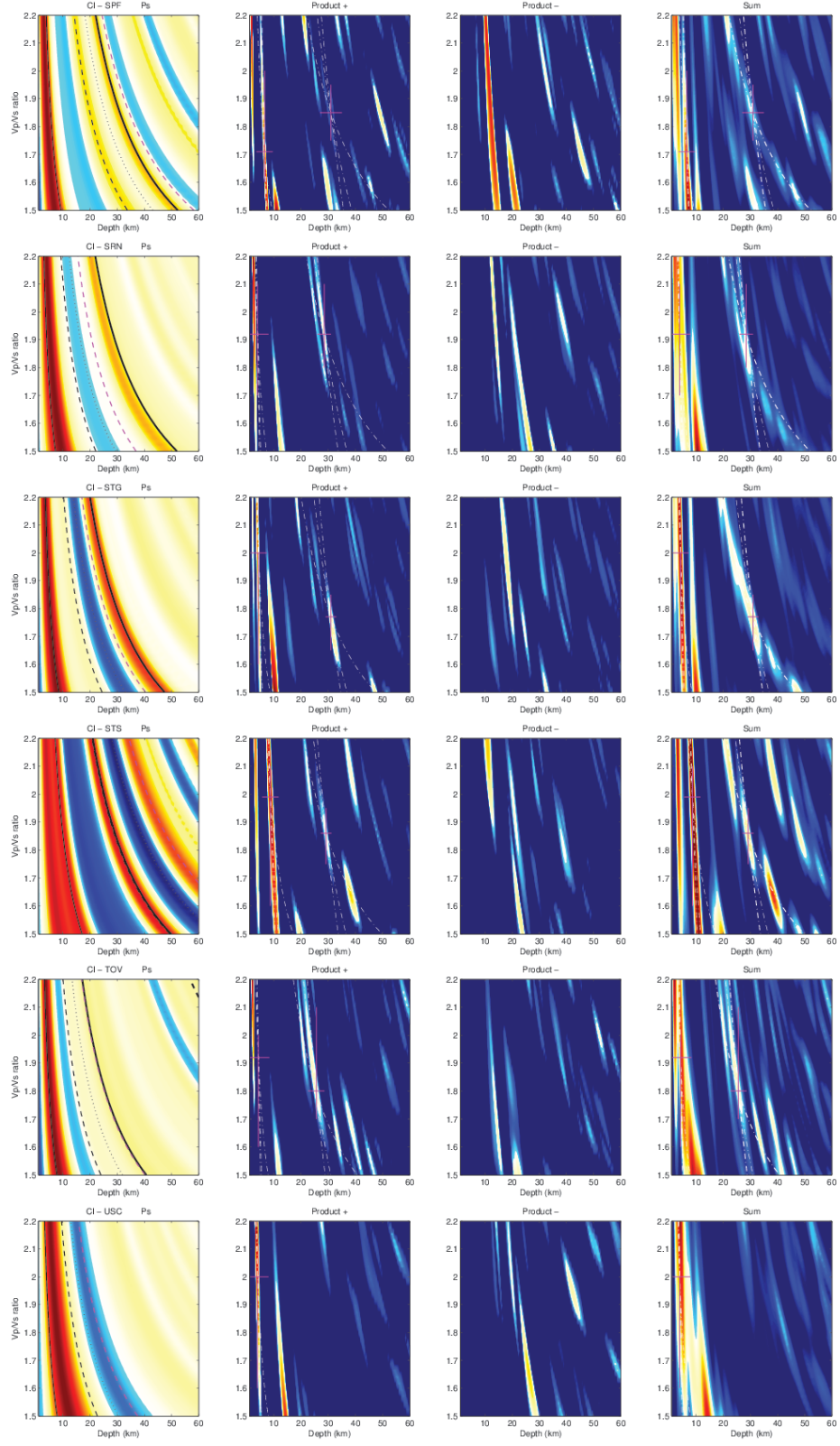
L.A. Basin

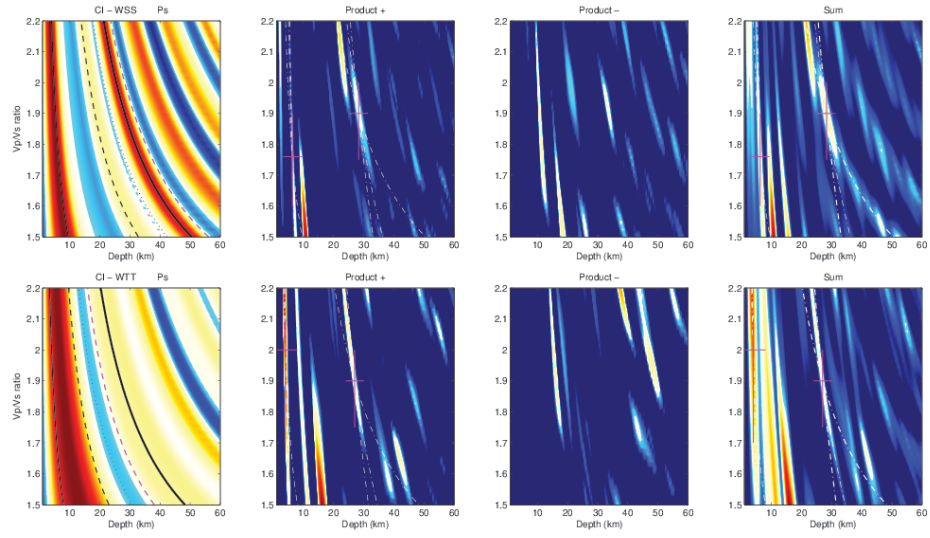
Figure A.6:





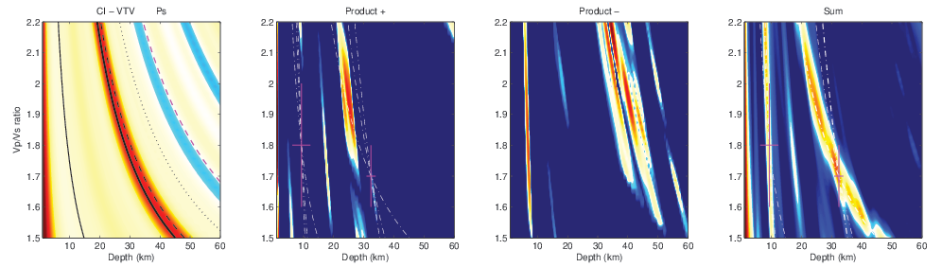






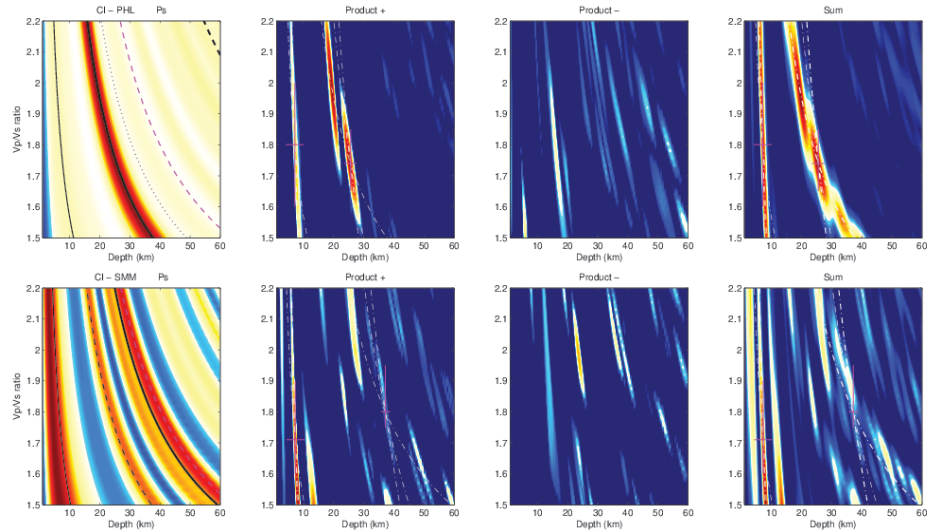
Control Station

Figure A.7:



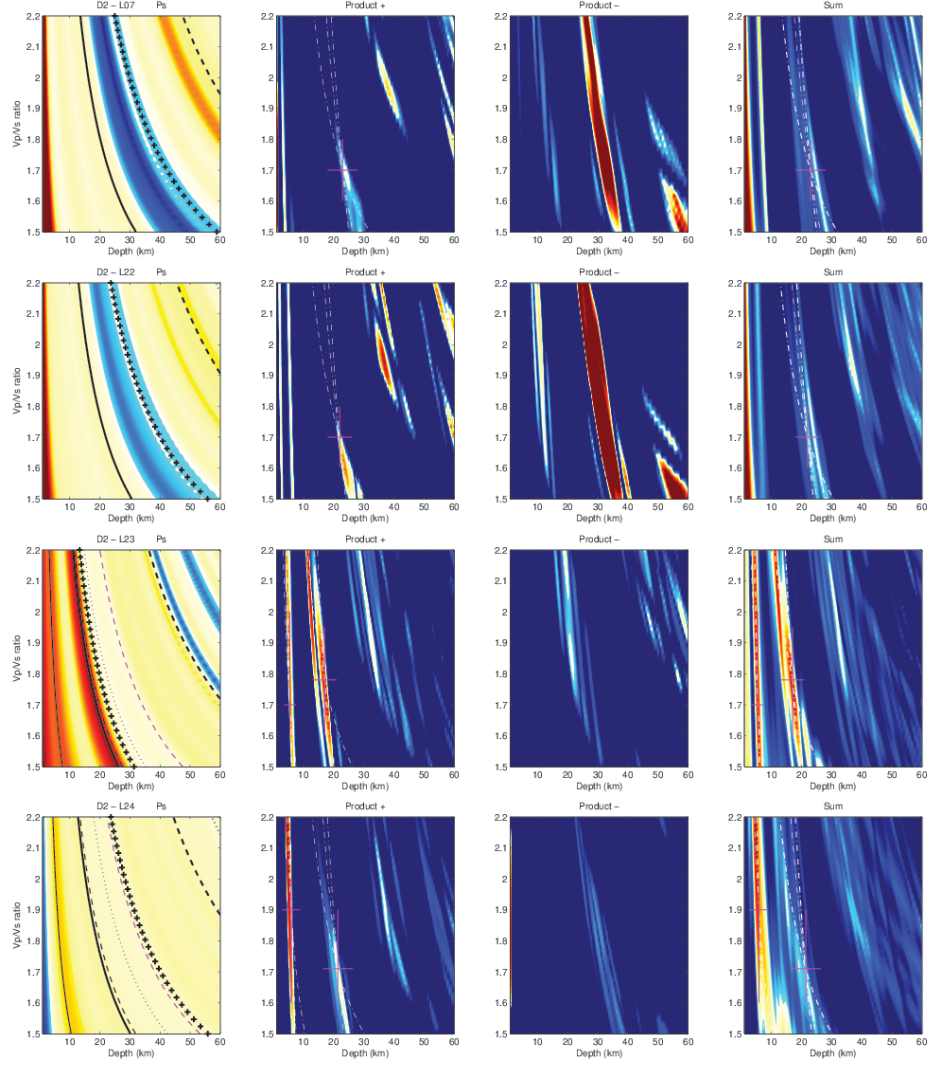
North of Western Transverse Range

Figure A.8:



Stations inbetween abyssal plain and Outer Borderland

Figure A.9:

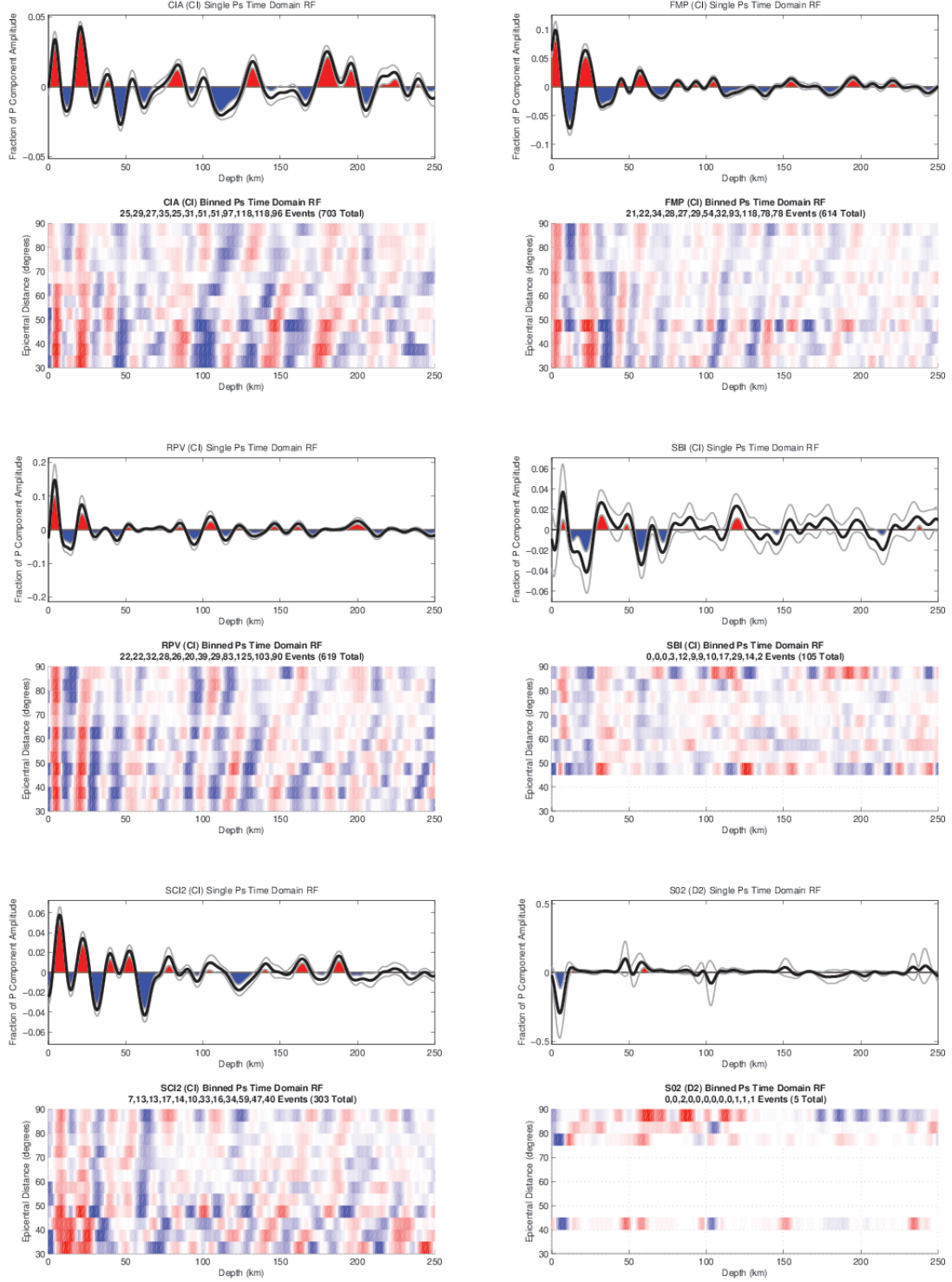


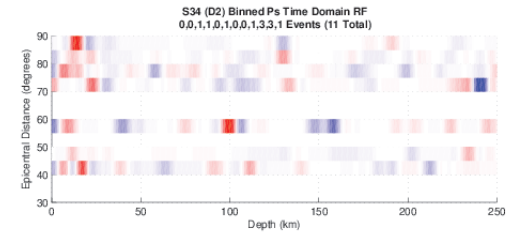
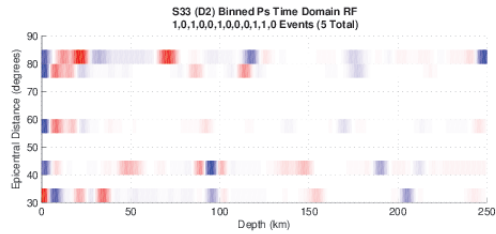
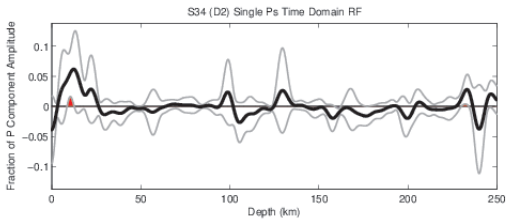
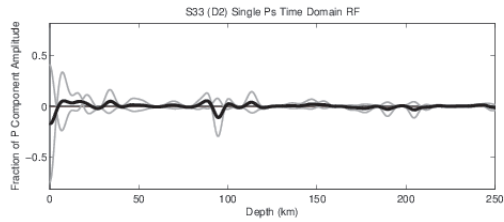
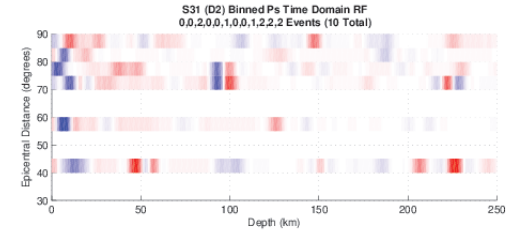
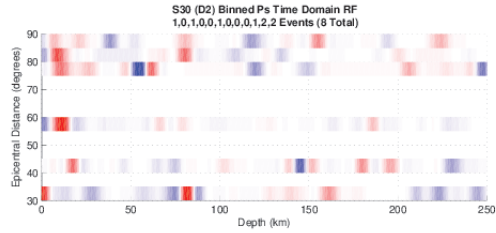
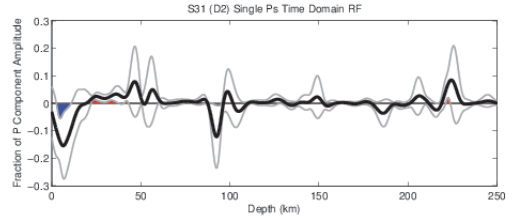
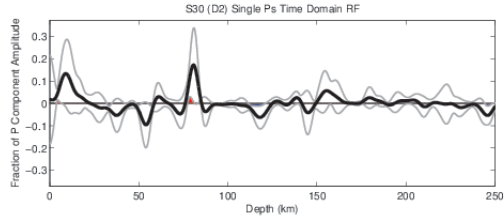
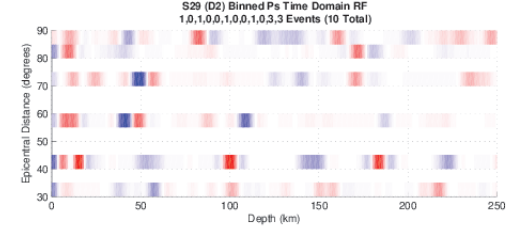
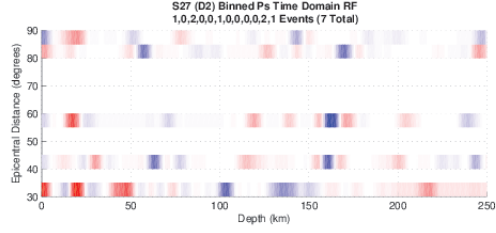
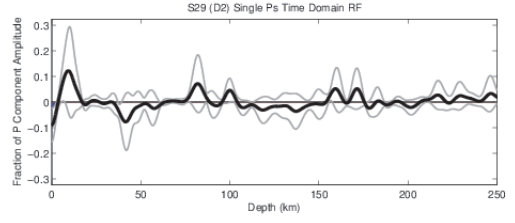
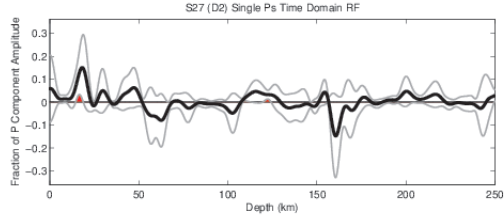
Chapter B: Appendix-Receiver Functions

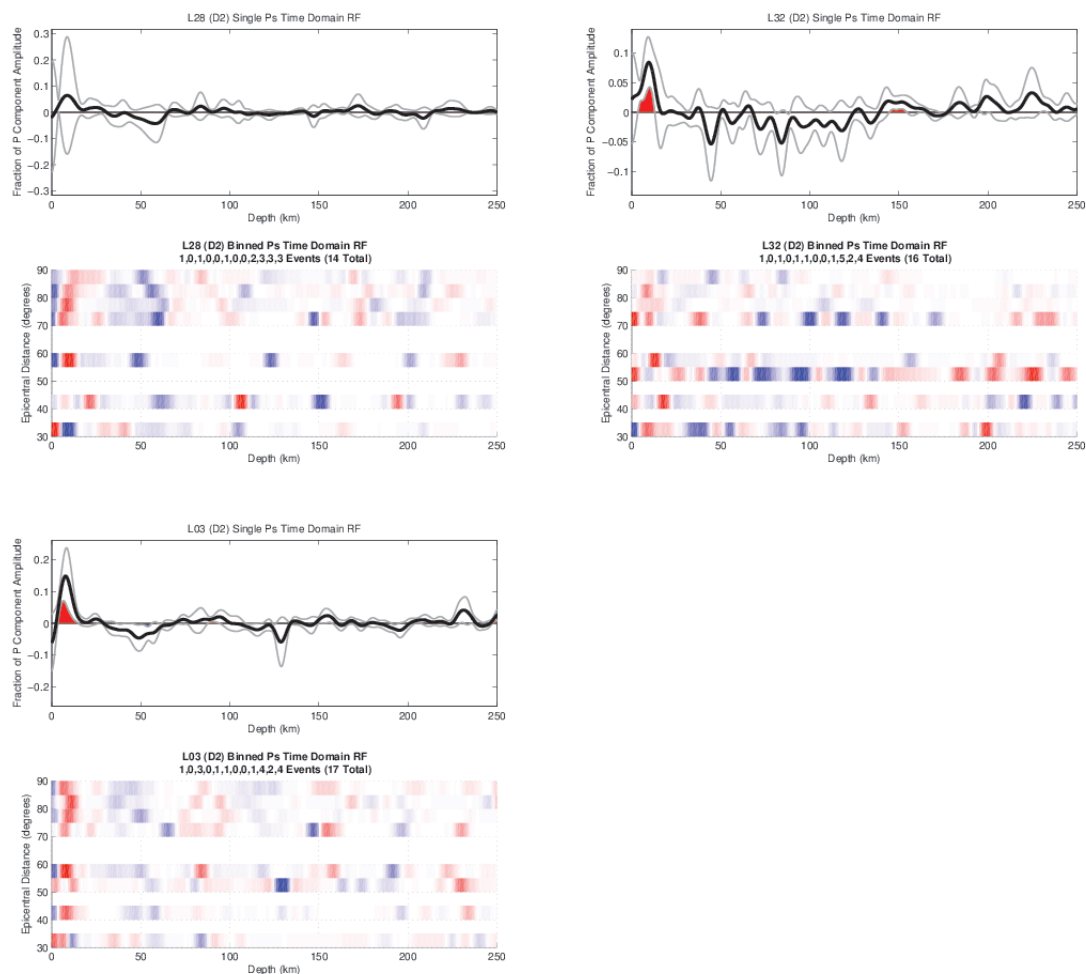
In this Appendix, we present the Ps and Sp receiver functions for all stations in our study.

Inner Borderland

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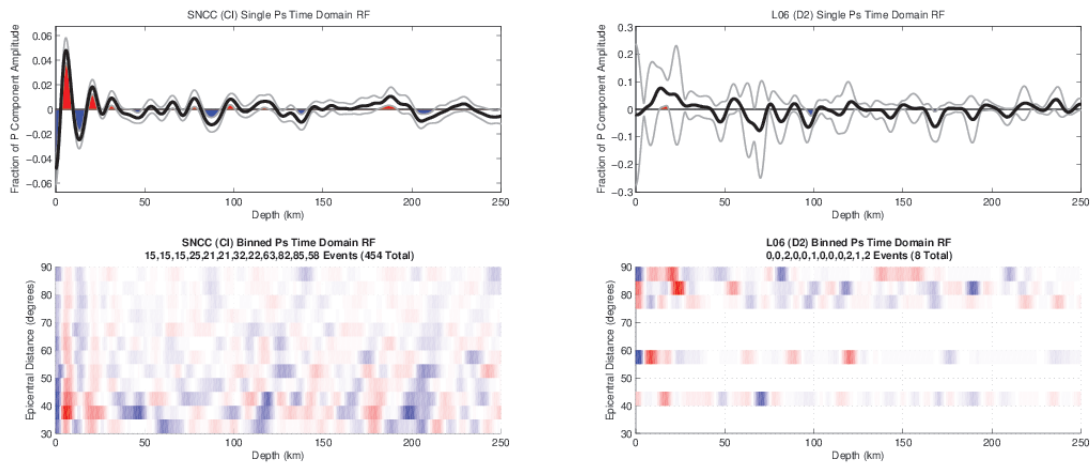


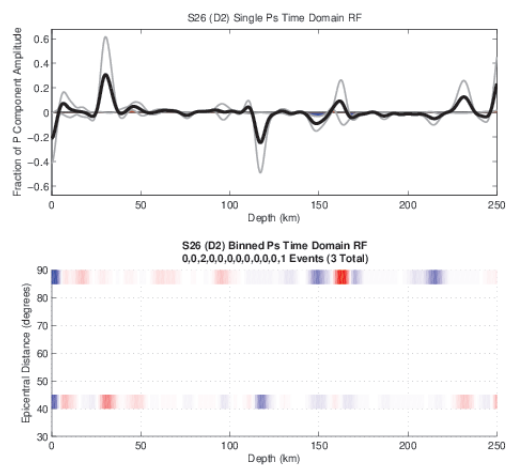




Outer Borderland

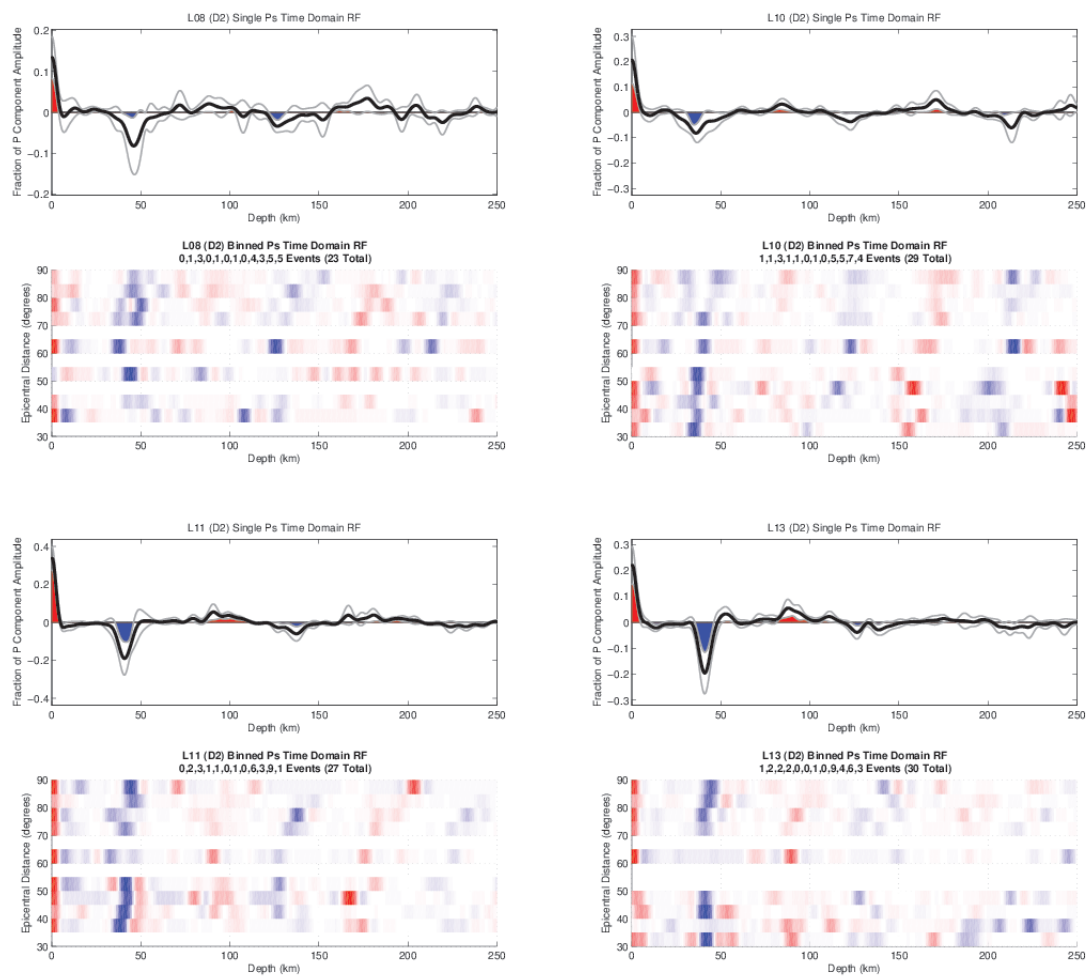
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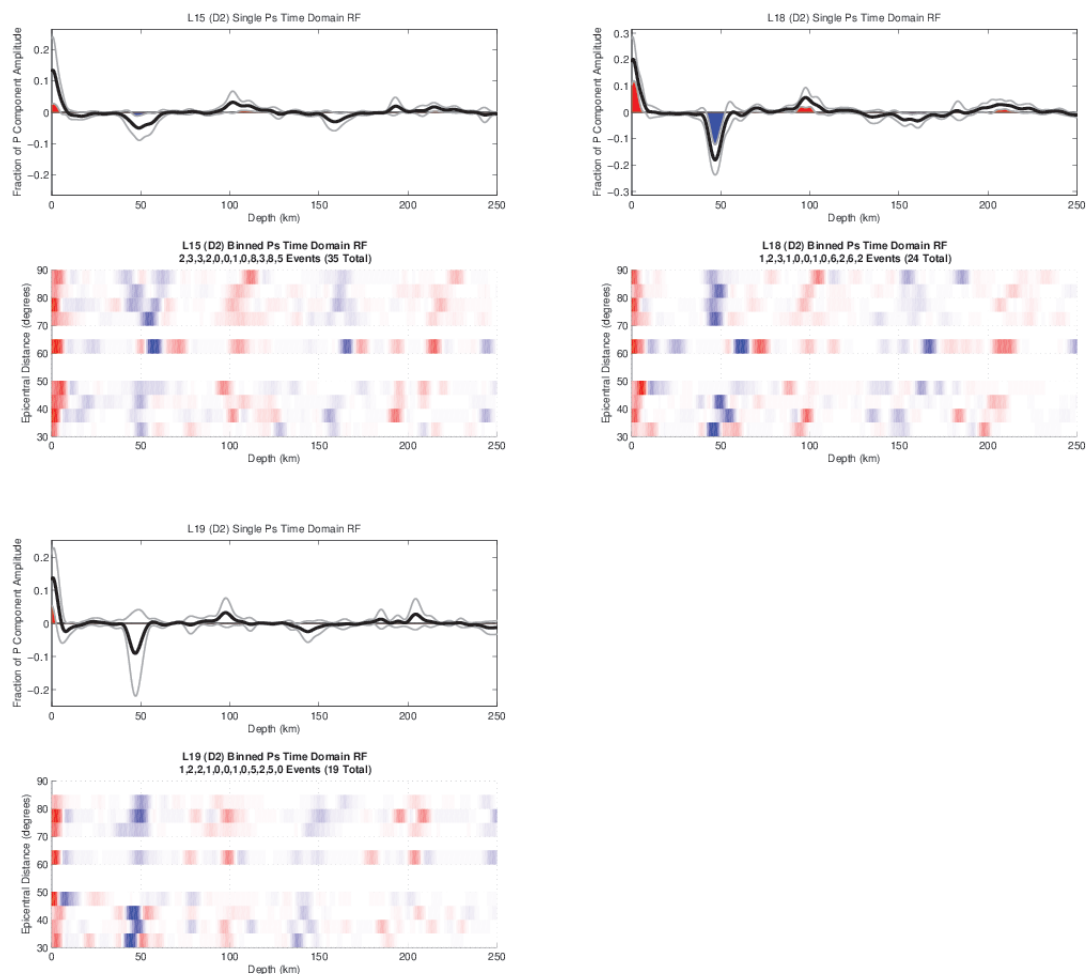




Abyssal Plain

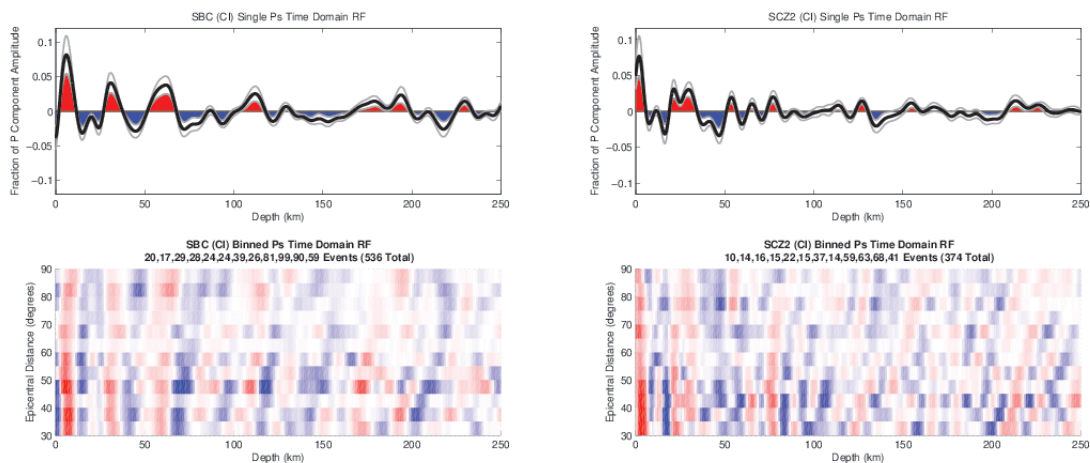
Figure B.3:

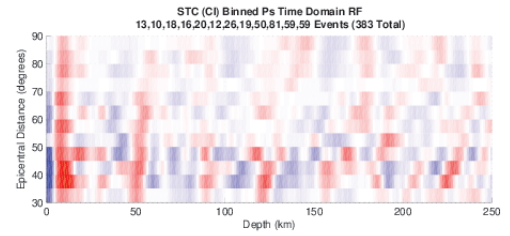
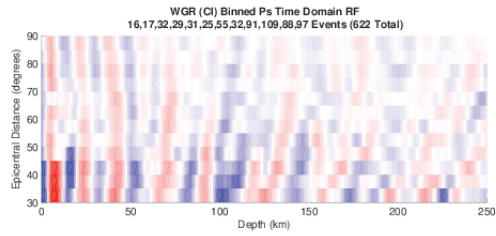
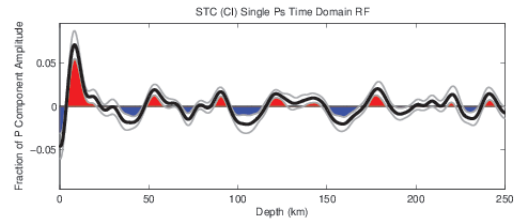
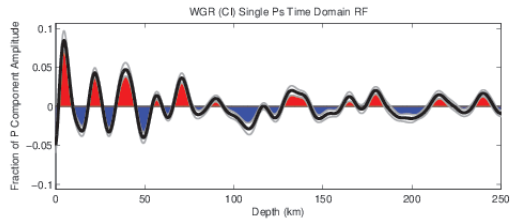
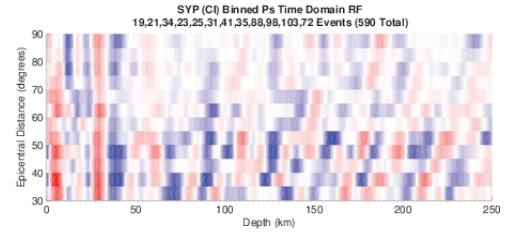
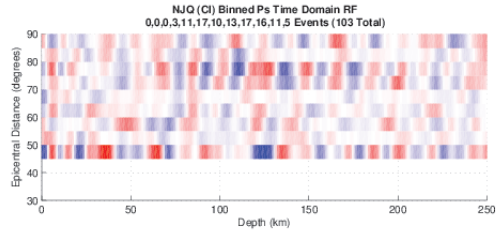
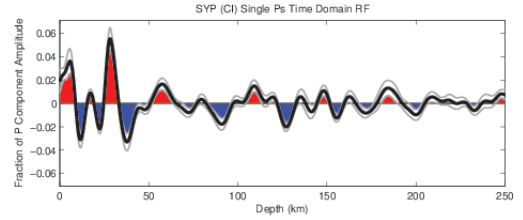
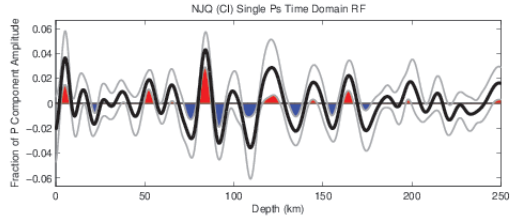
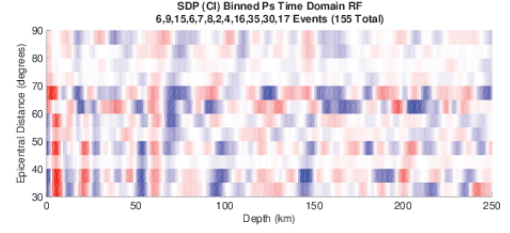
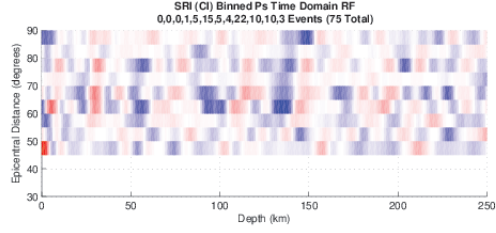
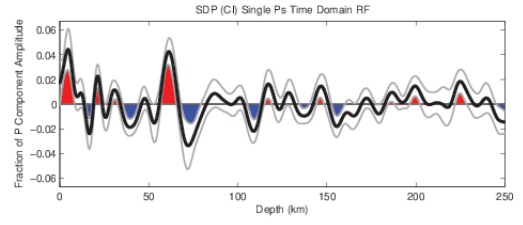
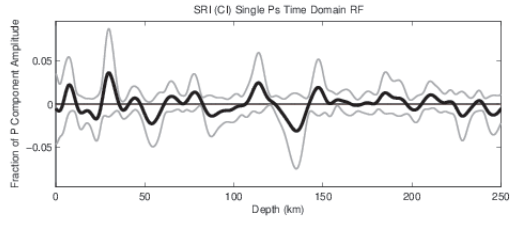


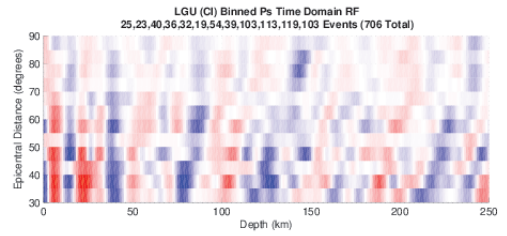
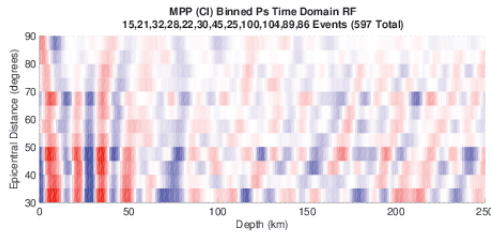
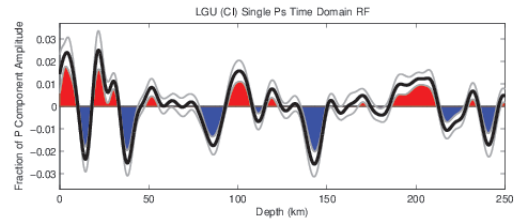
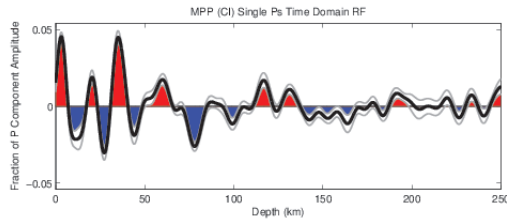
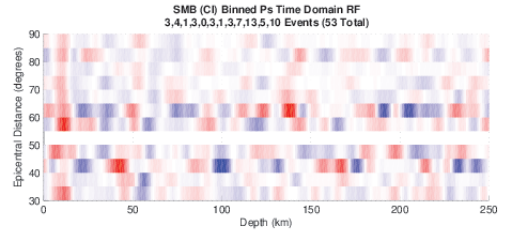
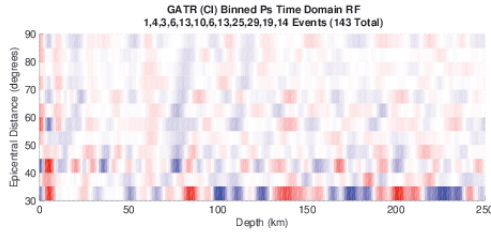
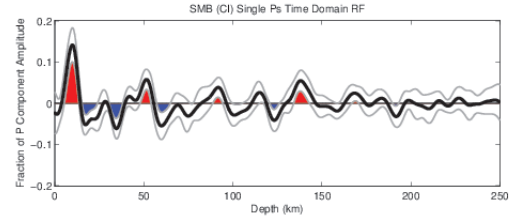
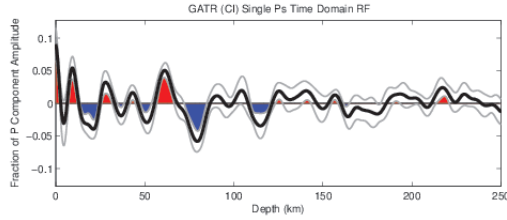
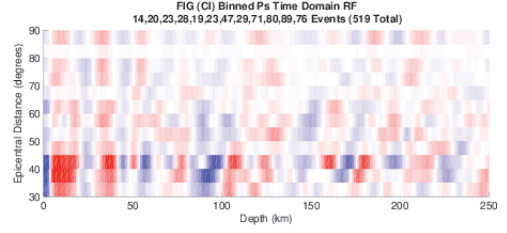
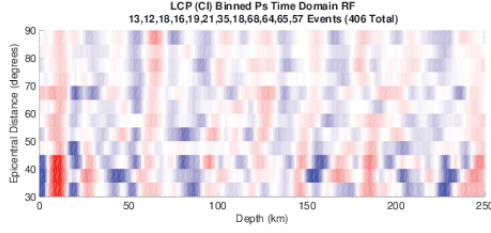
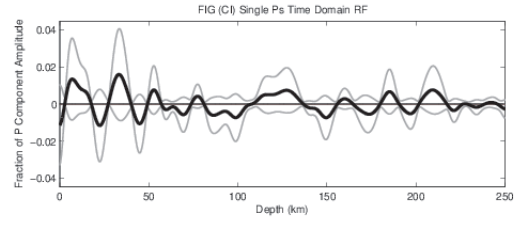
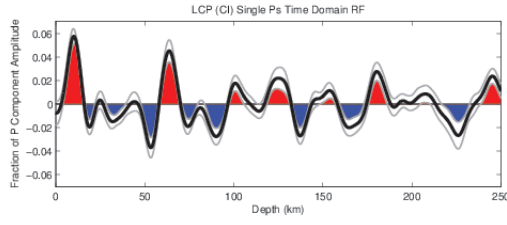


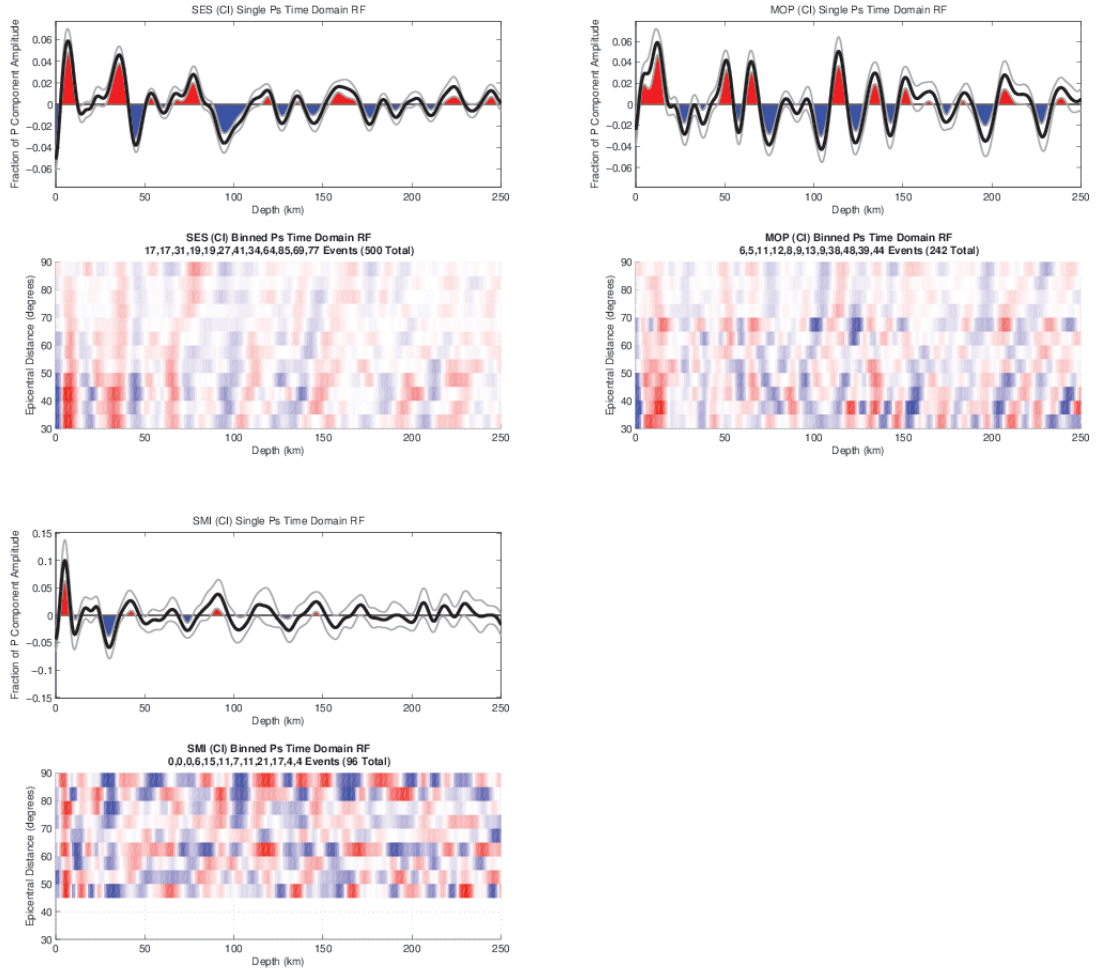
Western Transverse Range

Figure B.4:



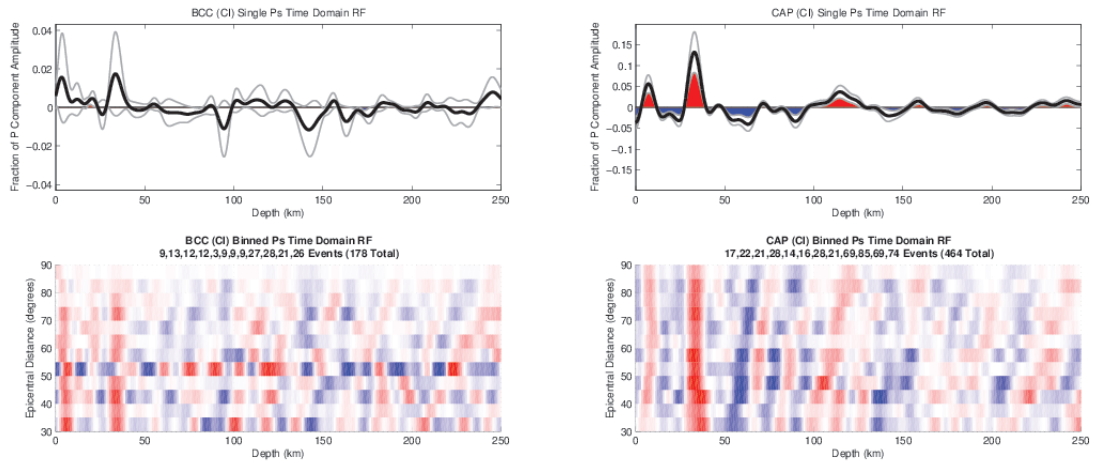


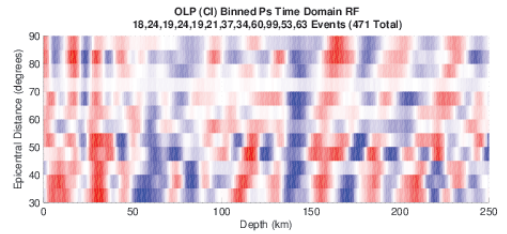
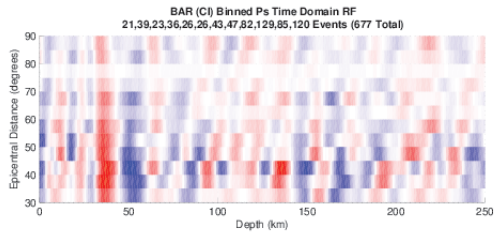
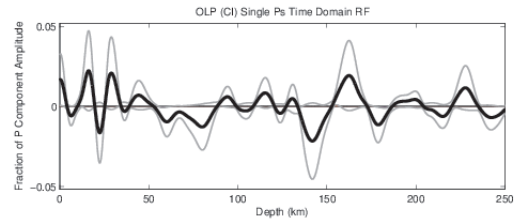
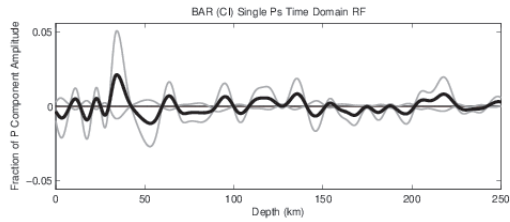
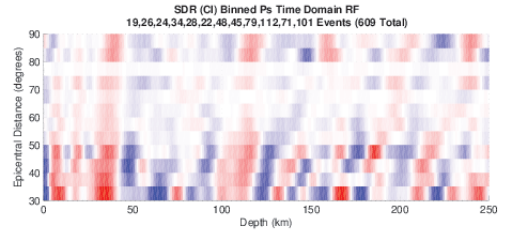
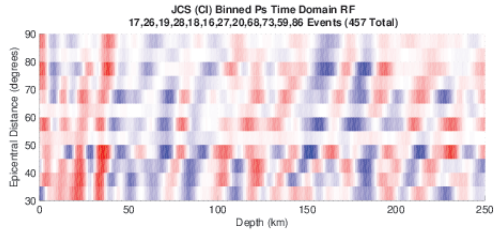
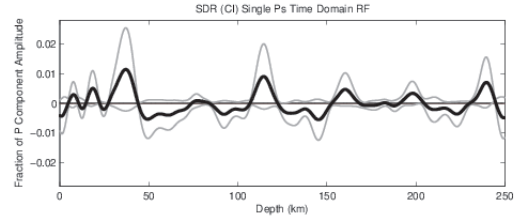
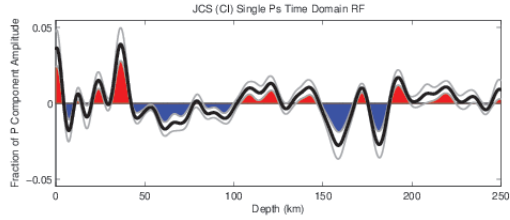
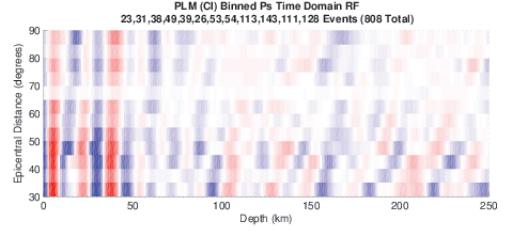
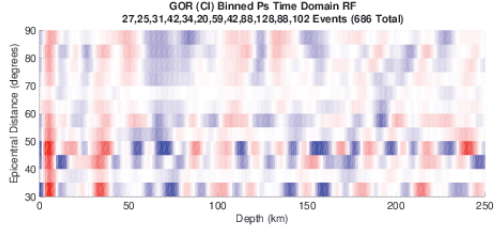
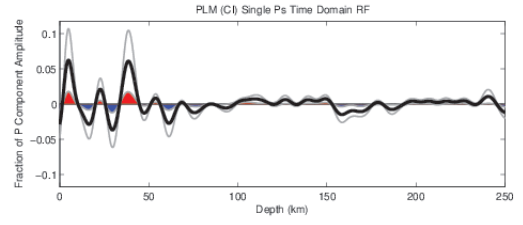
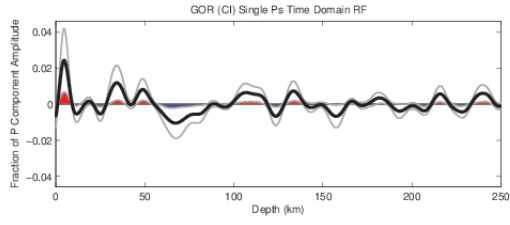


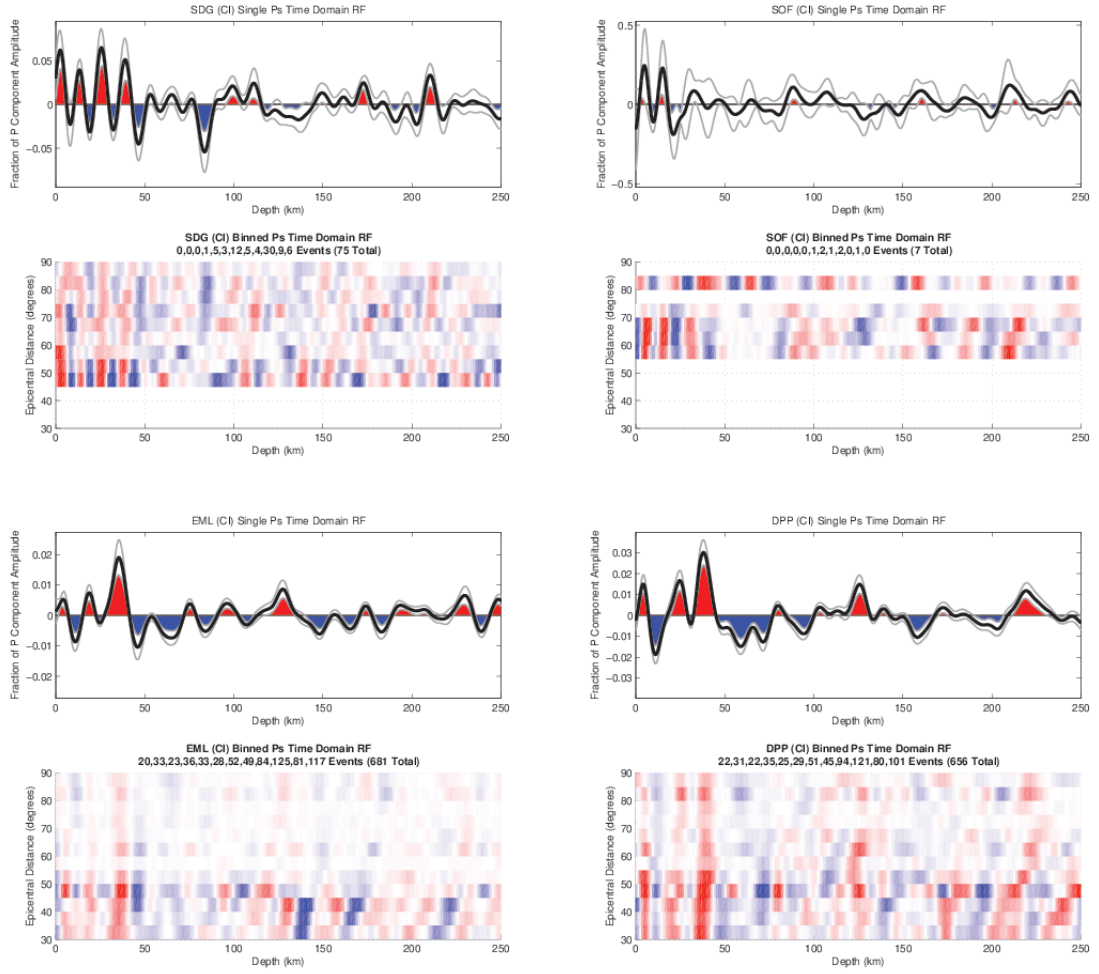


Peninsular Ranges

Figure B.5:

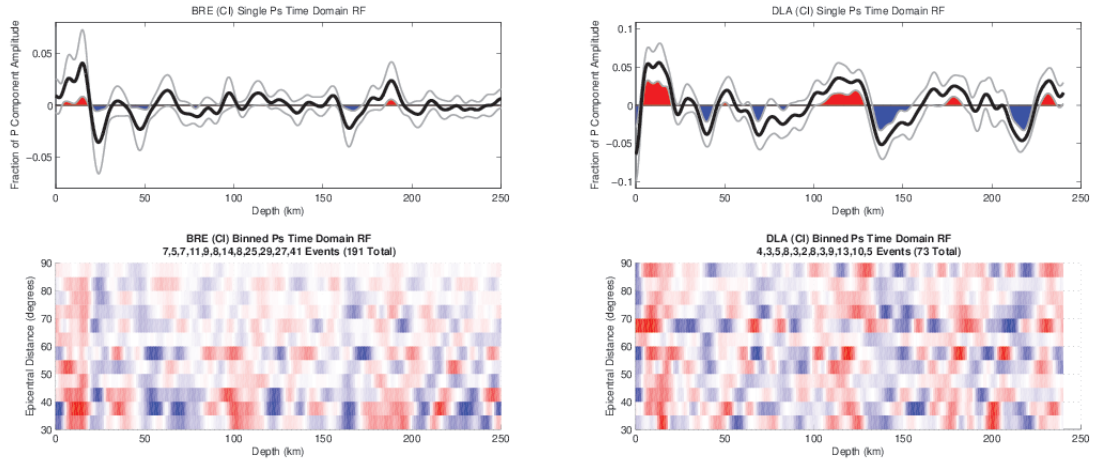


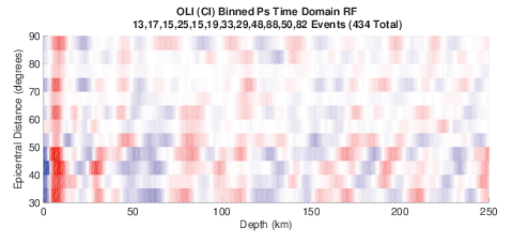
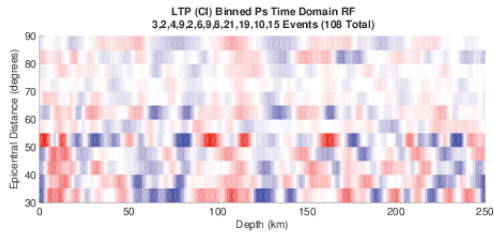
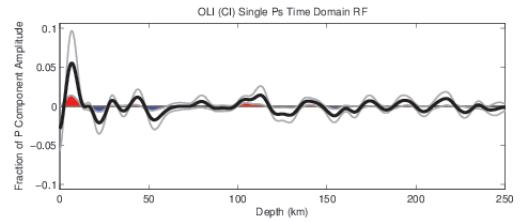
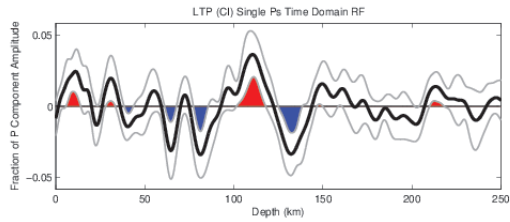
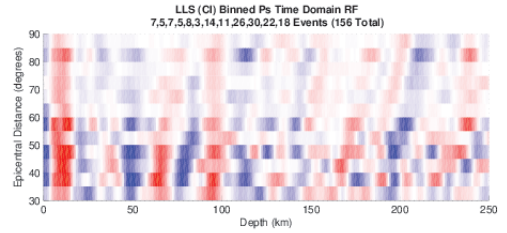
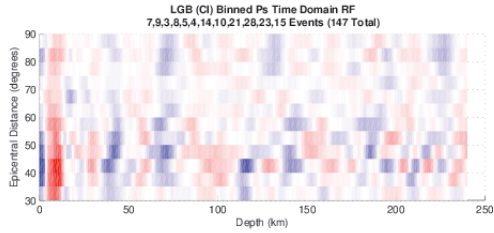
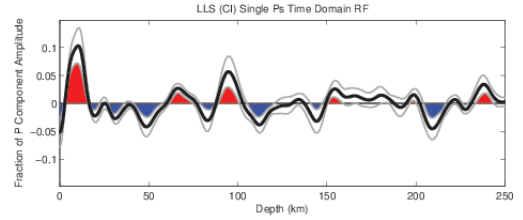
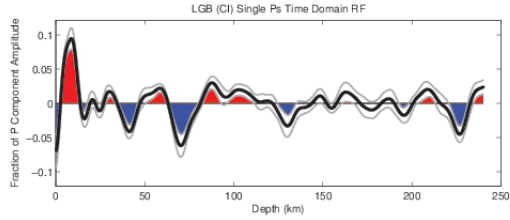
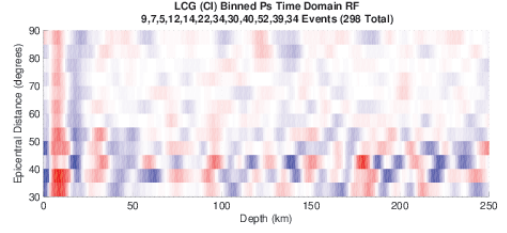
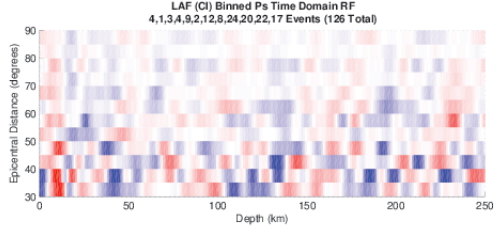
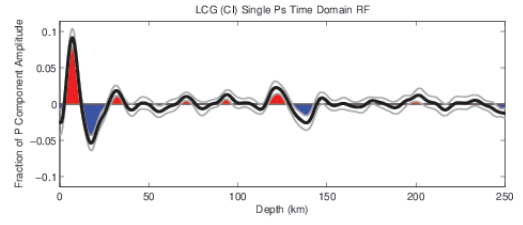
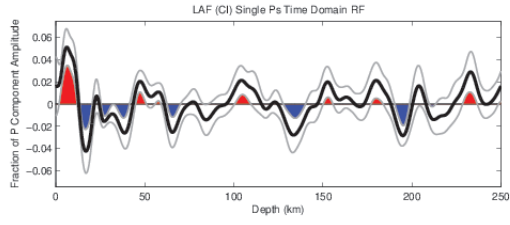


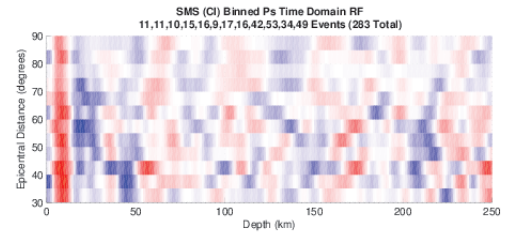
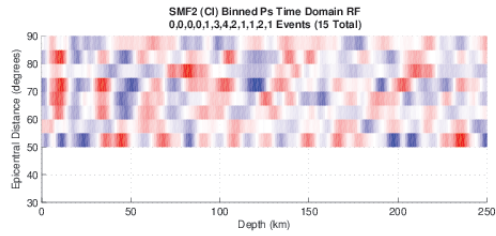
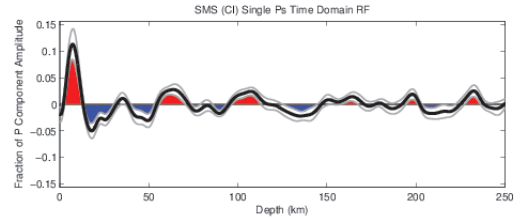
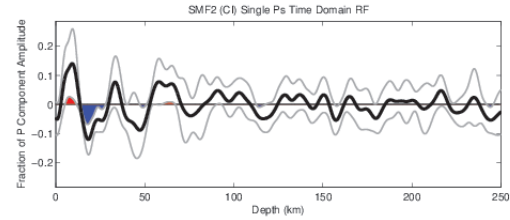
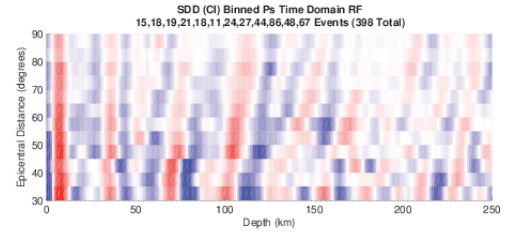
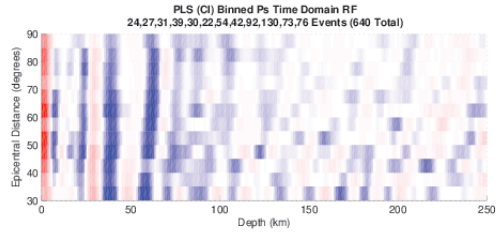
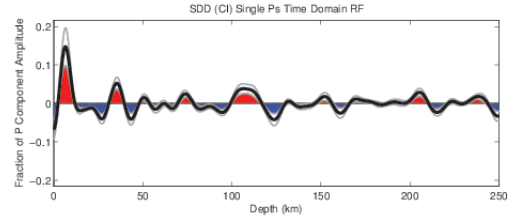
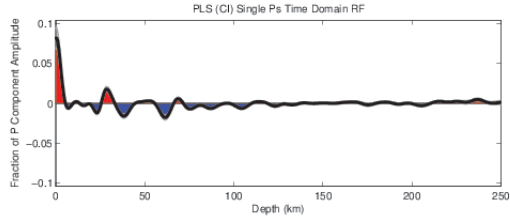
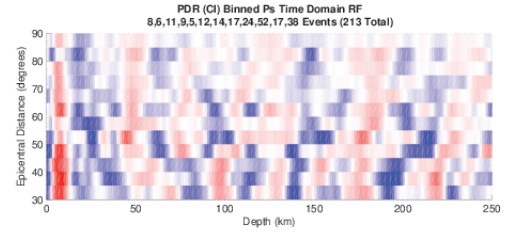
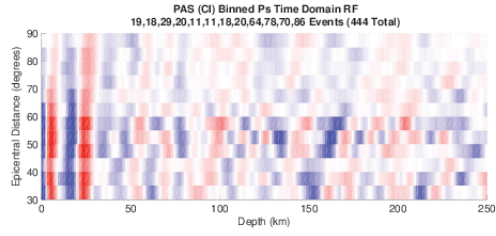
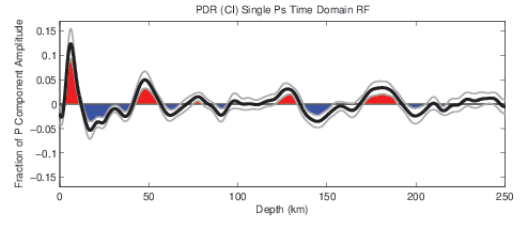
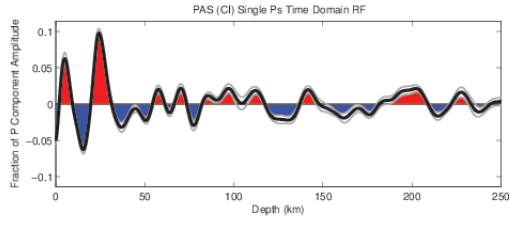


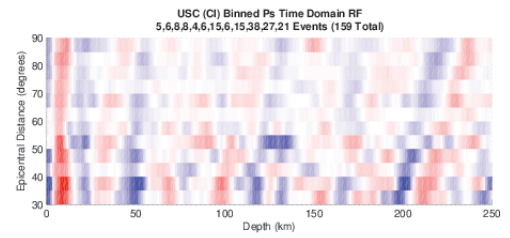
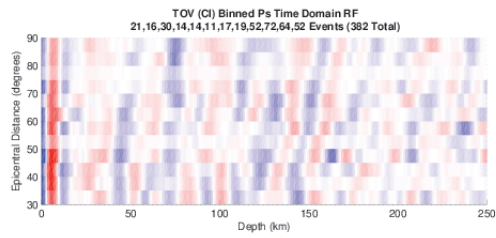
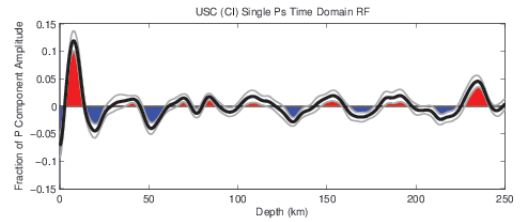
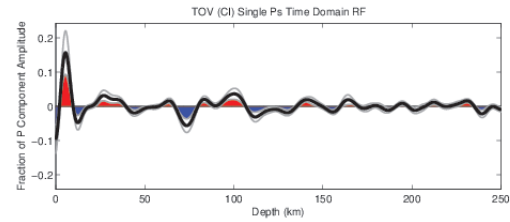
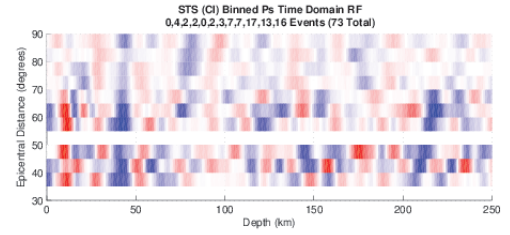
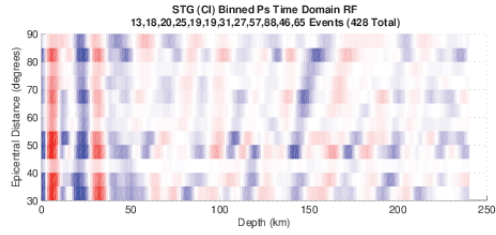
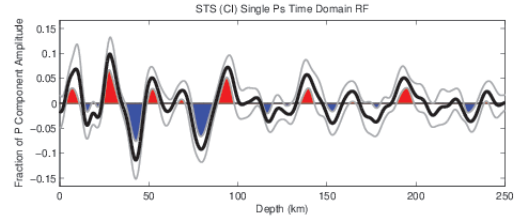
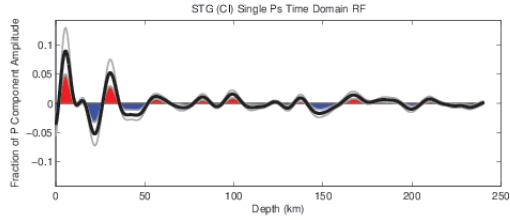
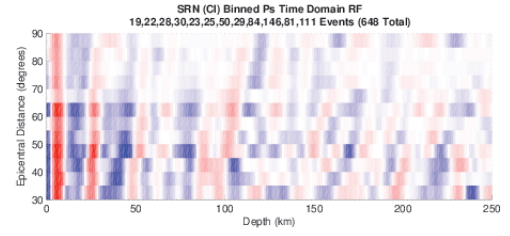
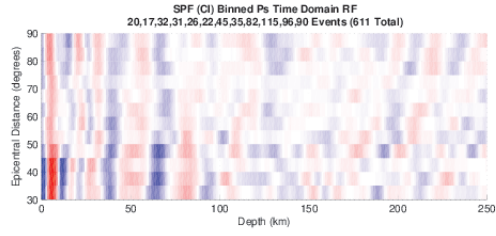
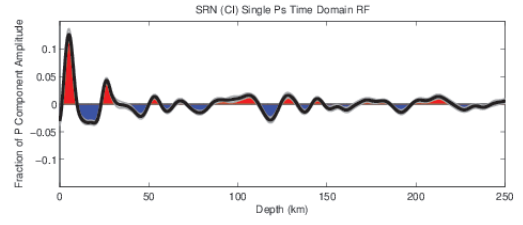
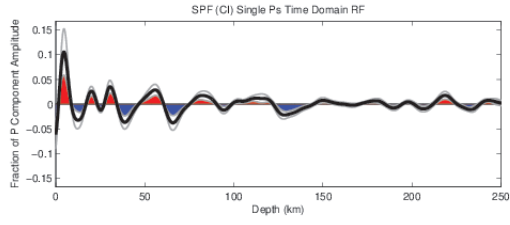
L.A. Basin

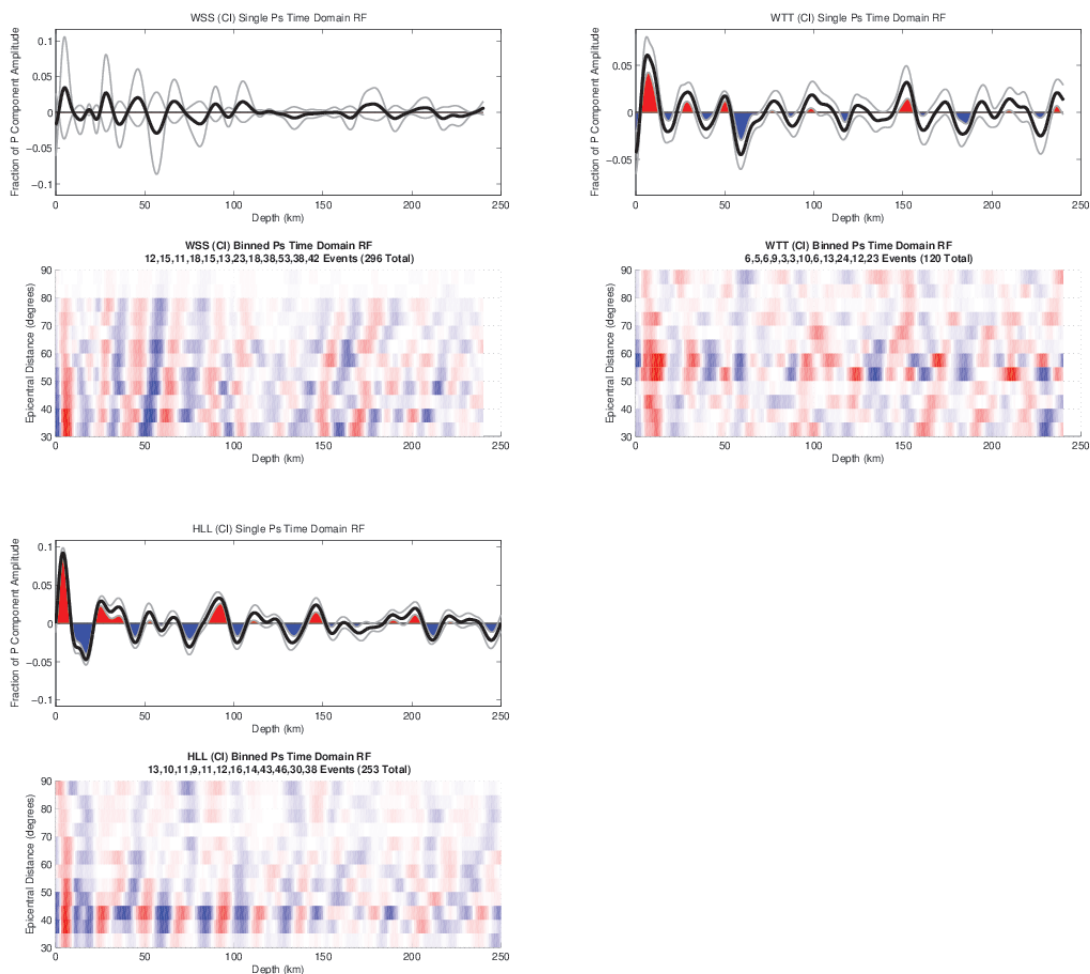
Figure B.6:





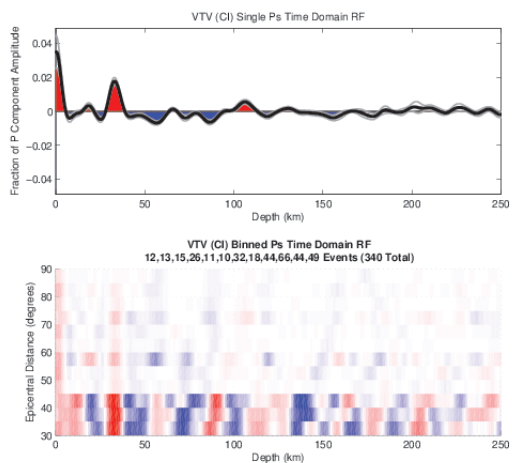






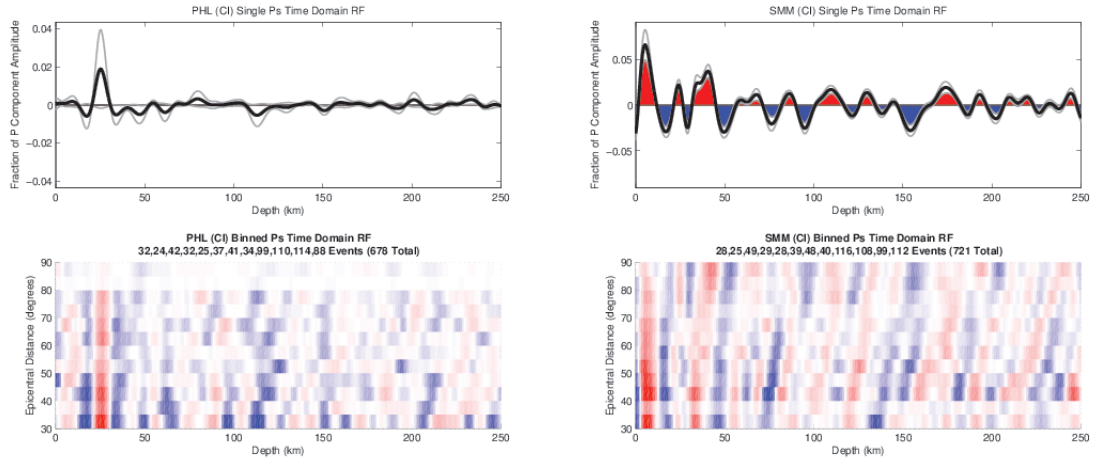
Control Station

Figure B.7:



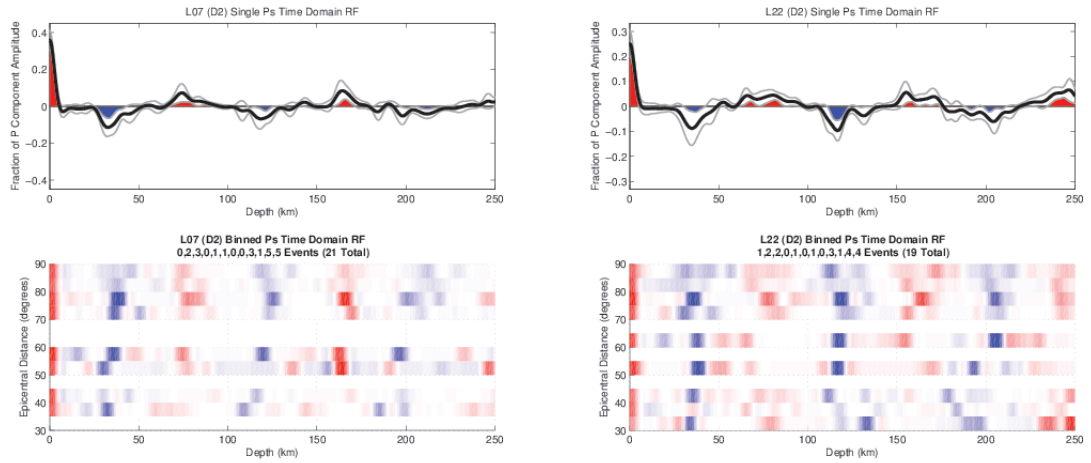
North of Western Transverse Range

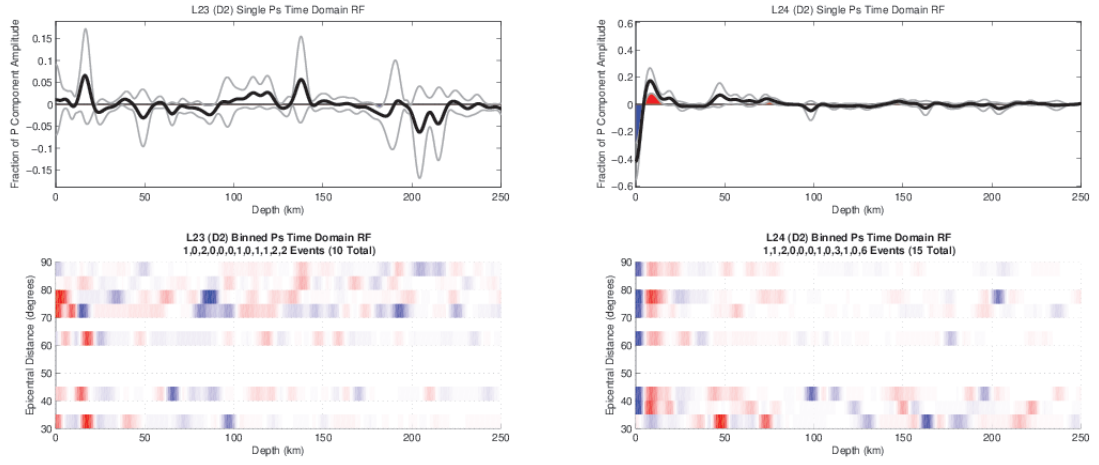
Figure B.8:



Stations inbetween abyssal plain and Outer Borderland

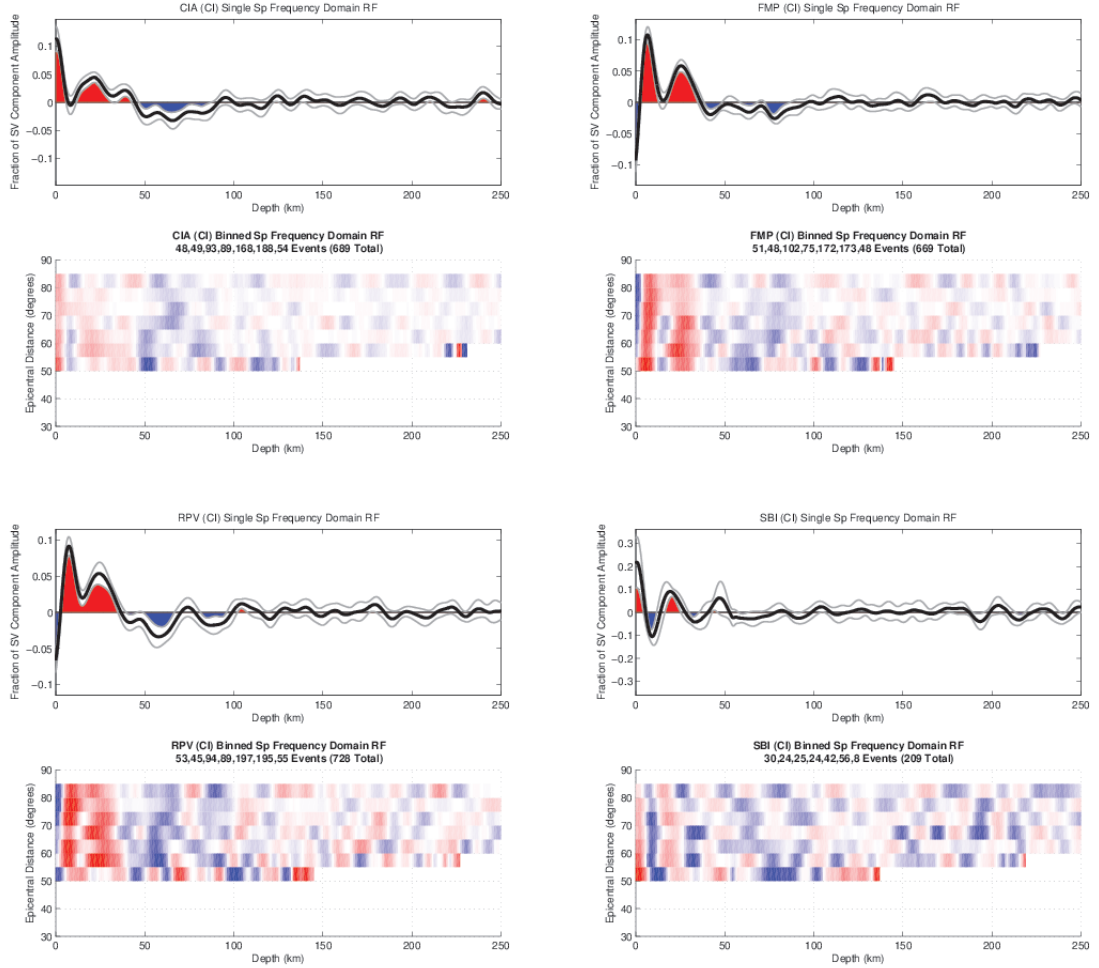
Figure B.9:

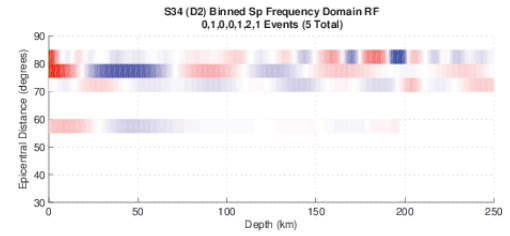
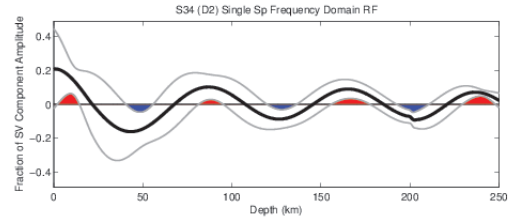
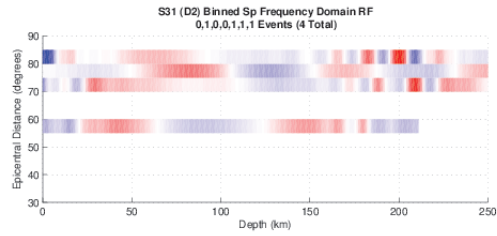
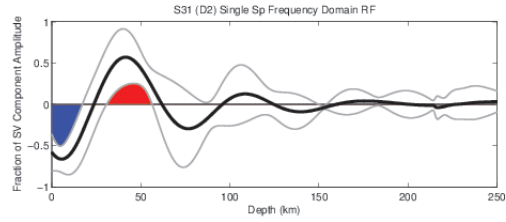
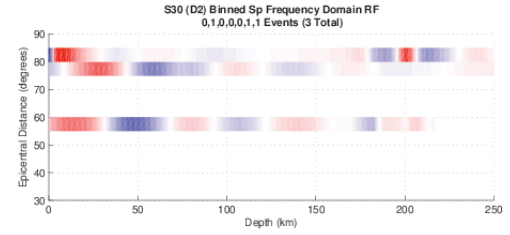
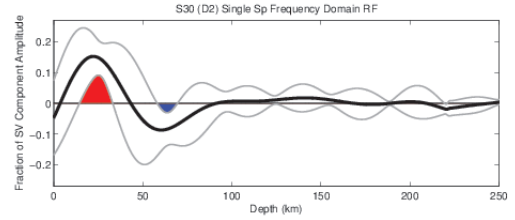
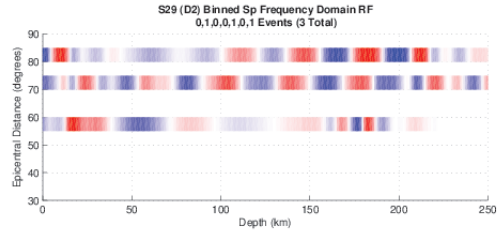
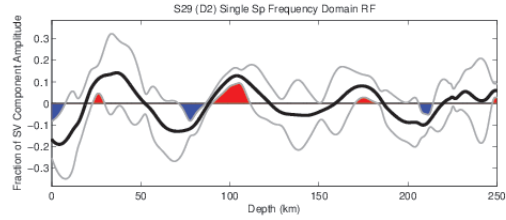
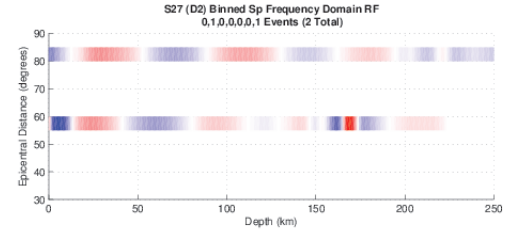
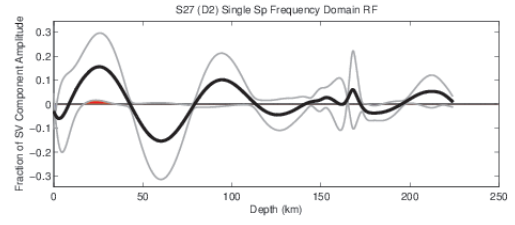
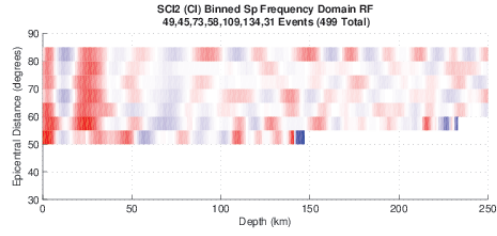
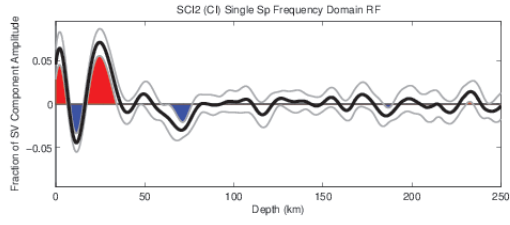


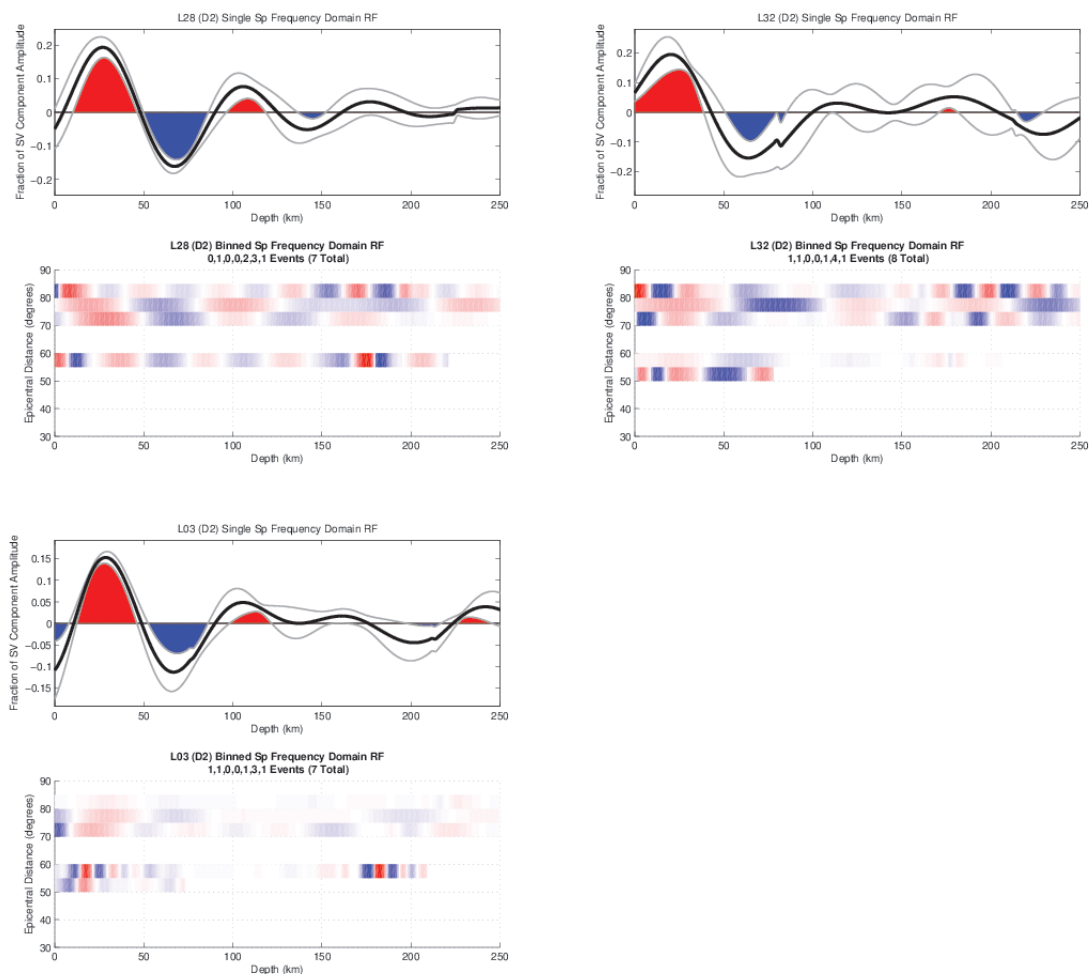


Inner Borderland

Figure B.10:

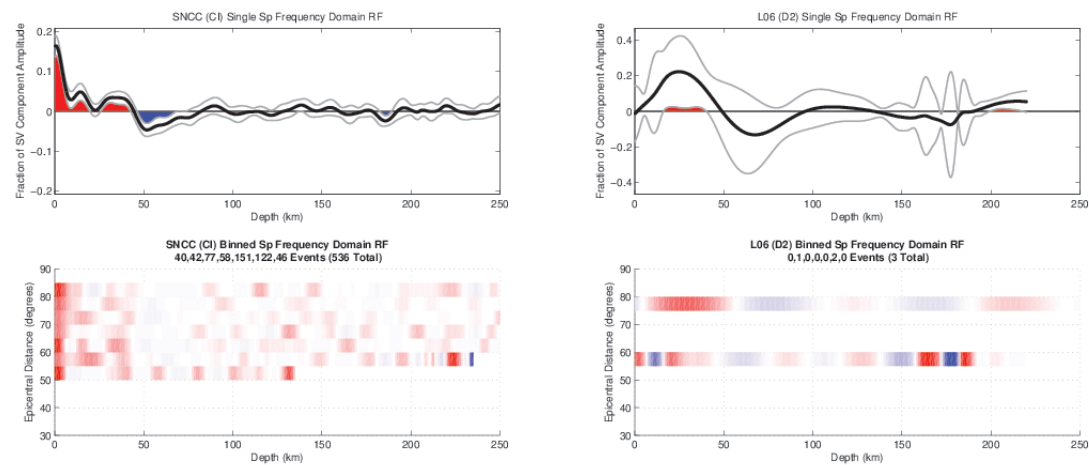






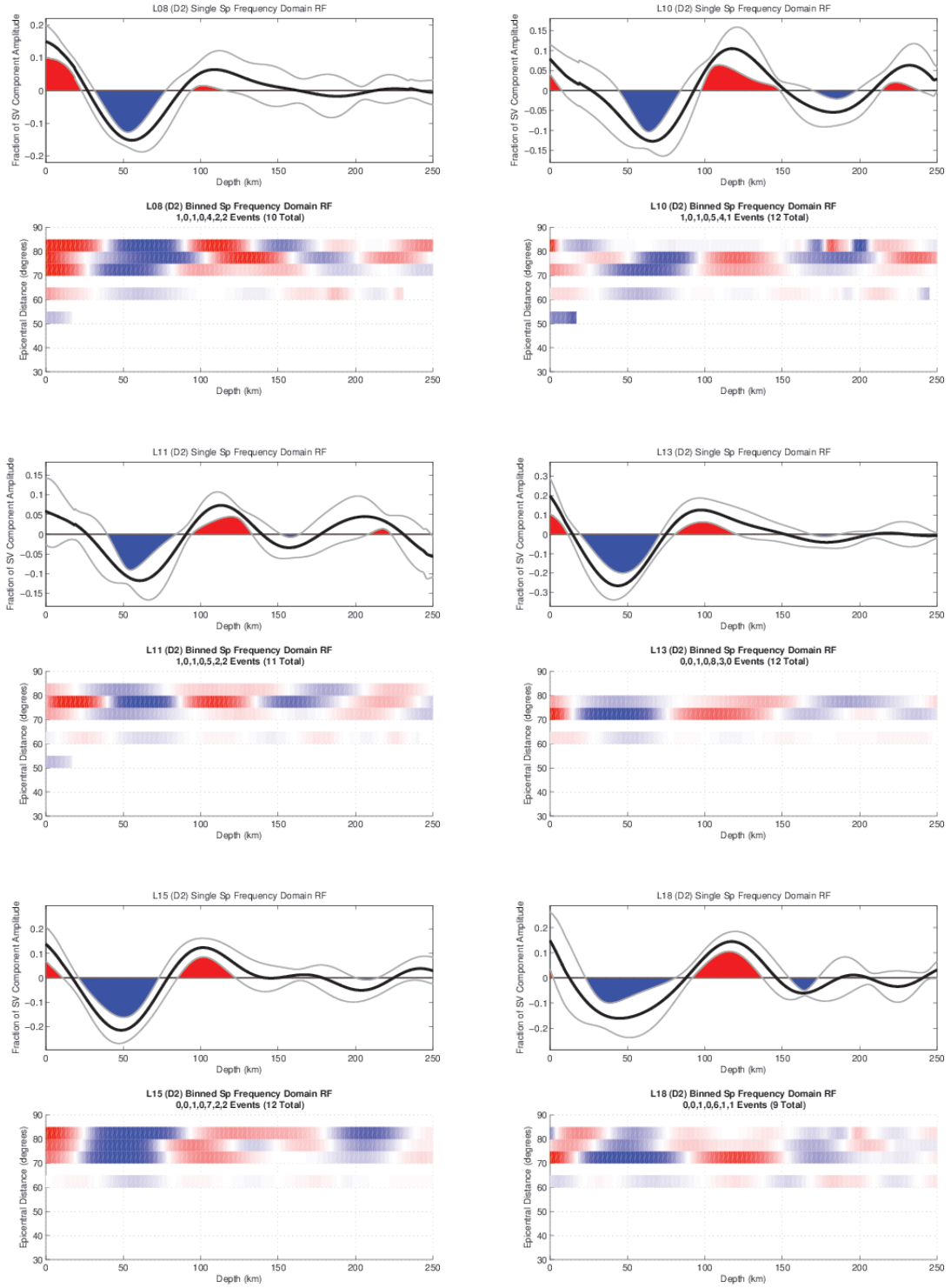
Outer Borderland

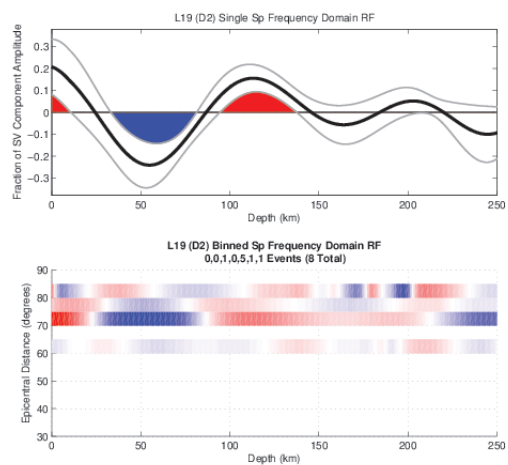
Figure B.11:



Abyssal Plain

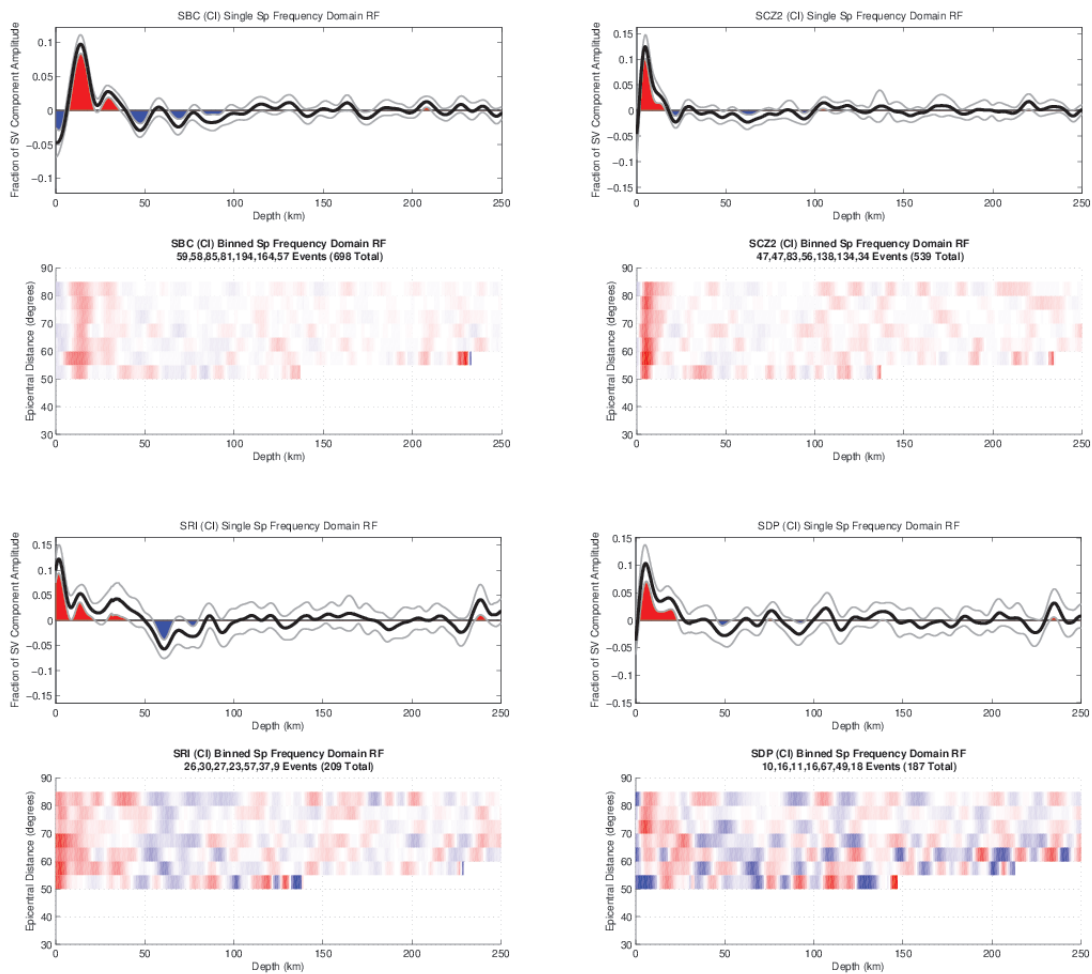
Figure B.12:

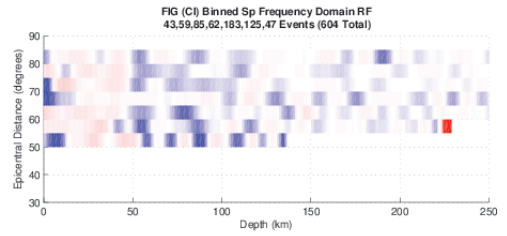
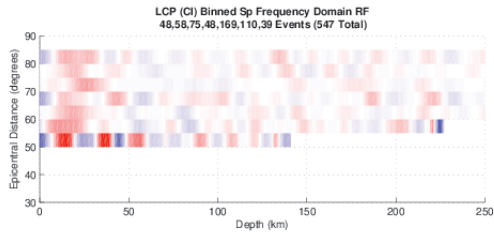
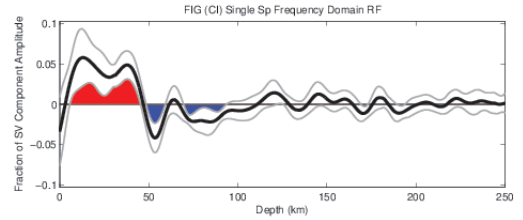
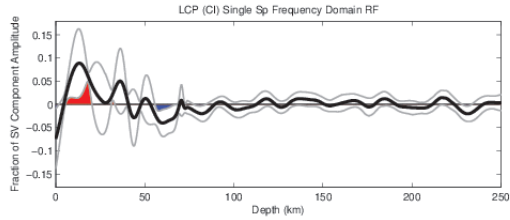
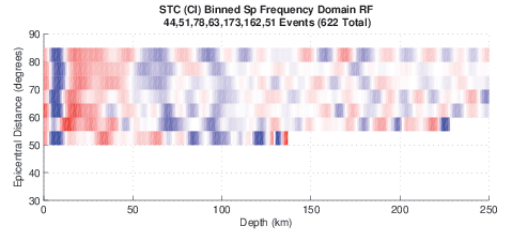
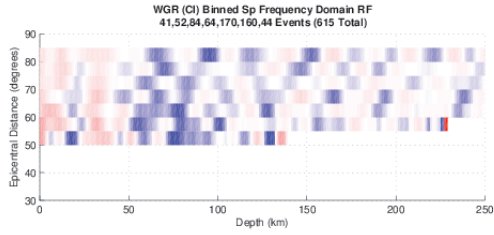
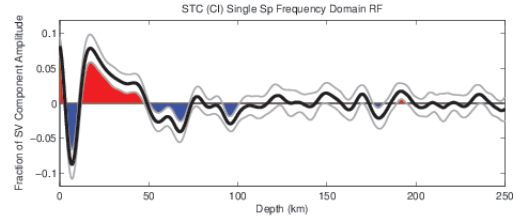
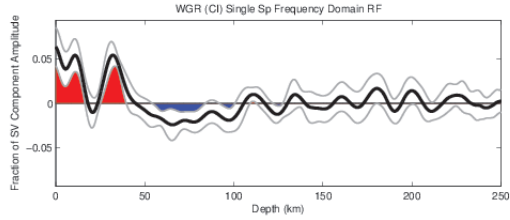
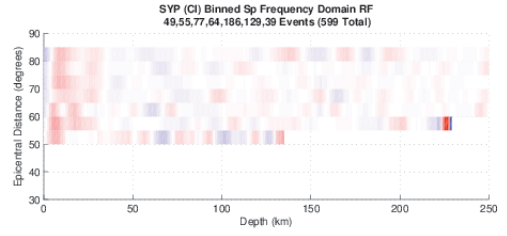
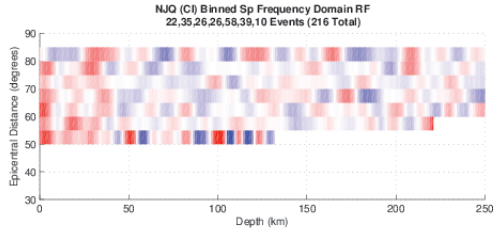
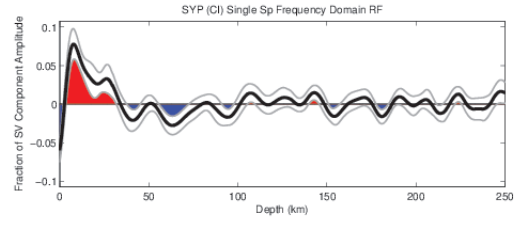
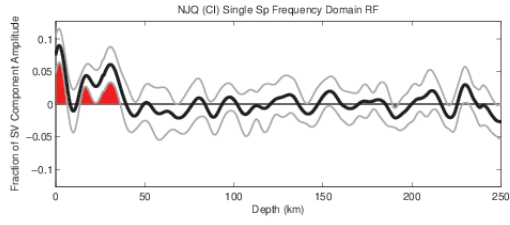


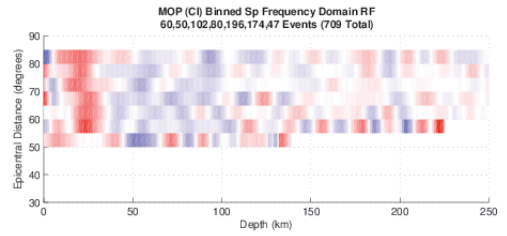
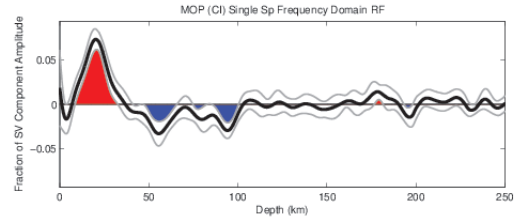
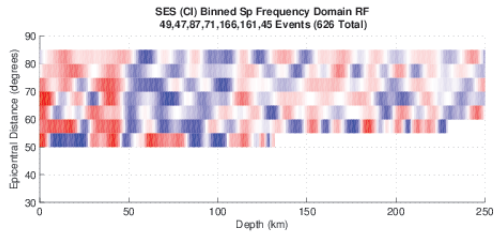
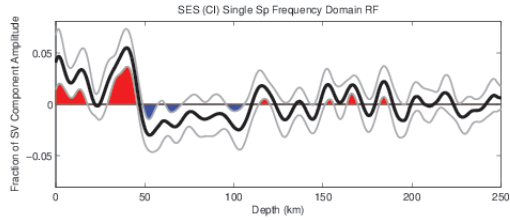
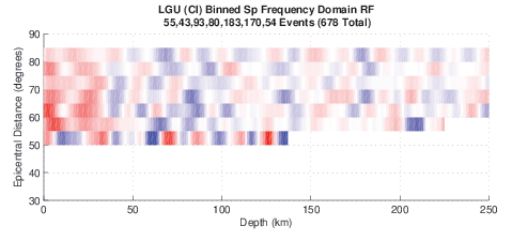
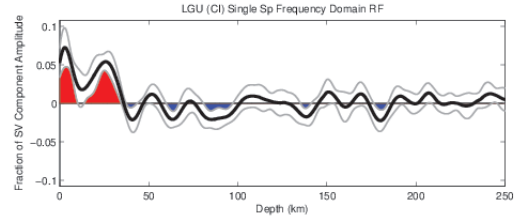
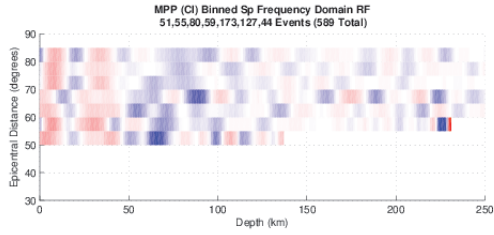
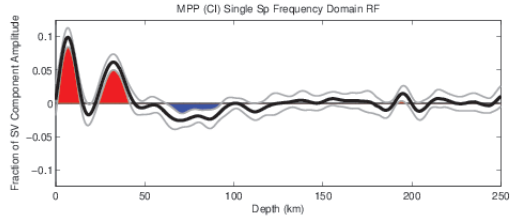
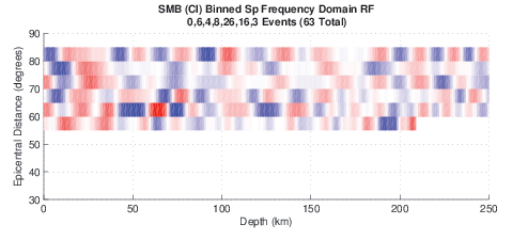
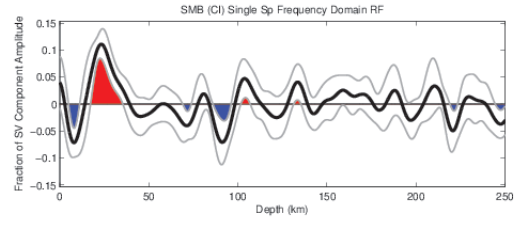
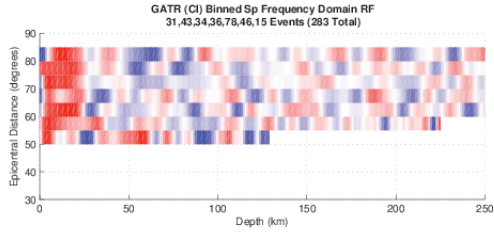
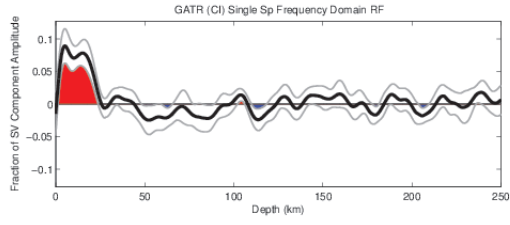


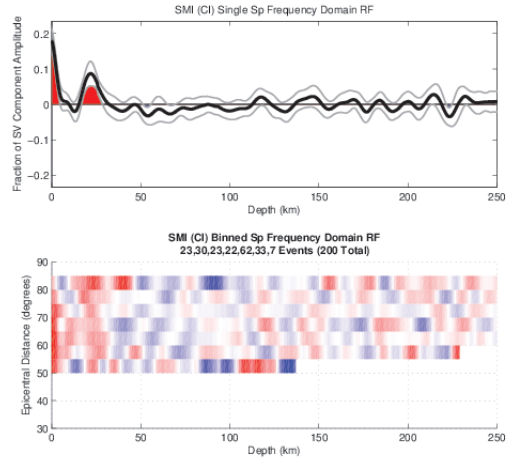
Western Transverse Range

Figure B.13:



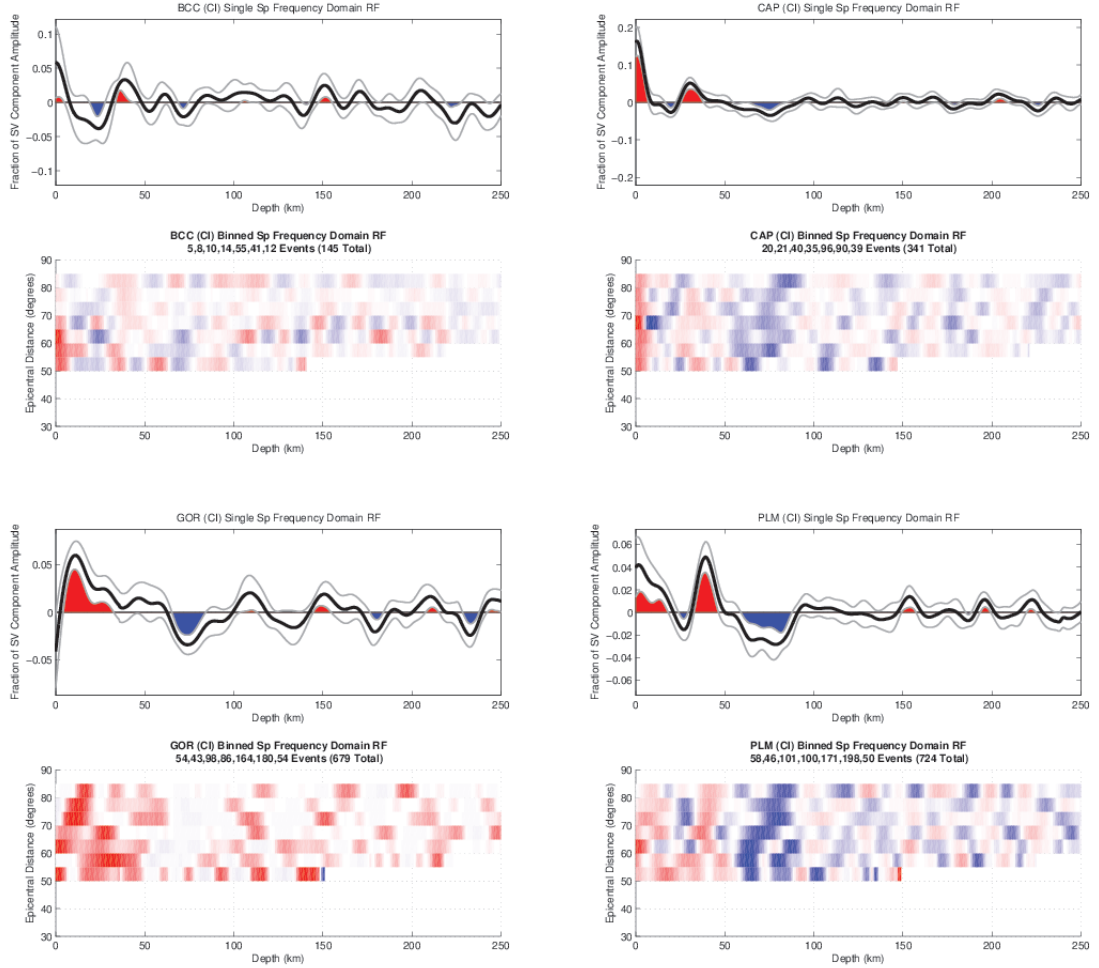


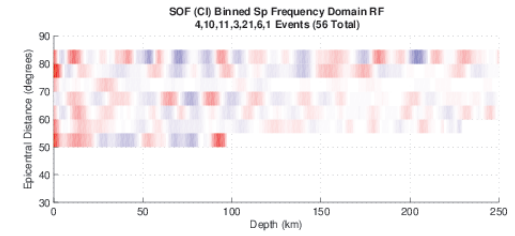
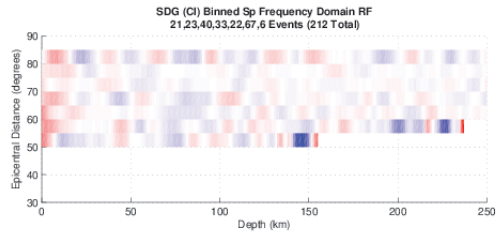
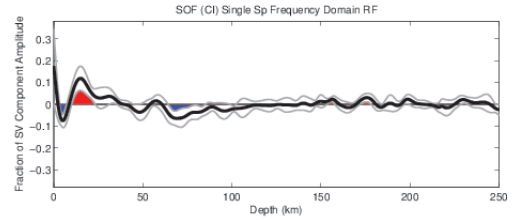
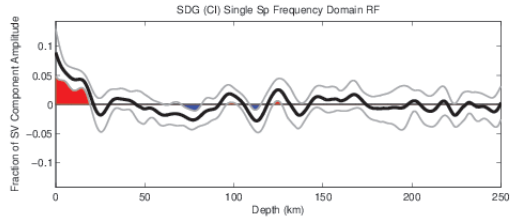
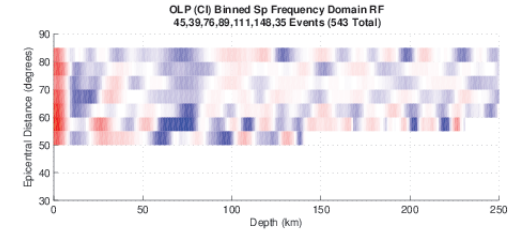
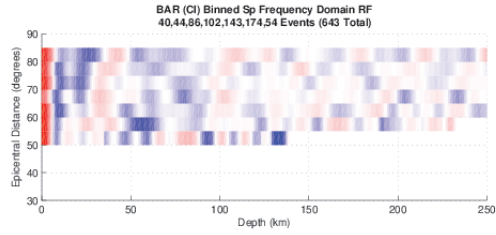
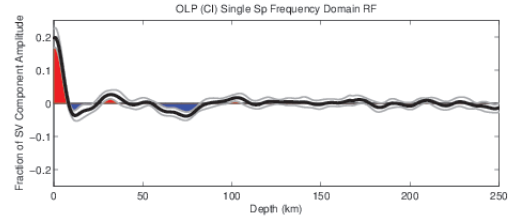
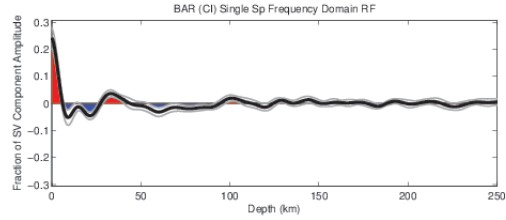
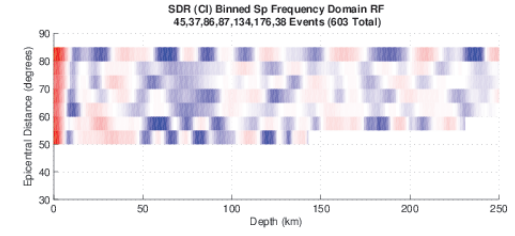
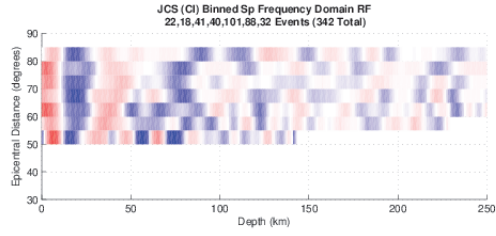
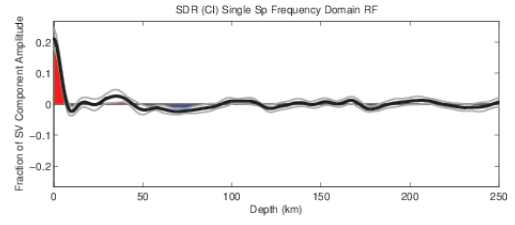
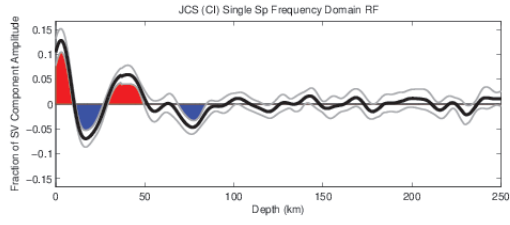


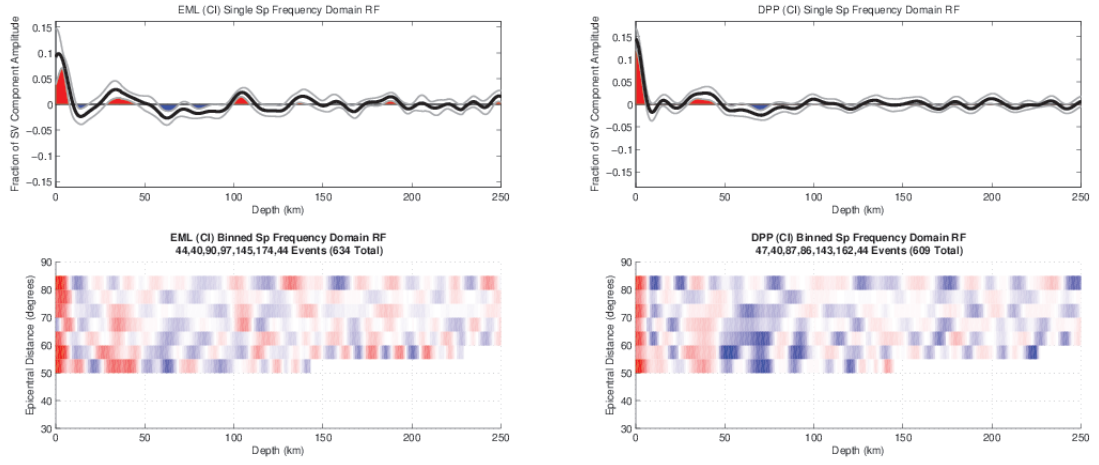


Peninsular Ranges

Figure B.14:

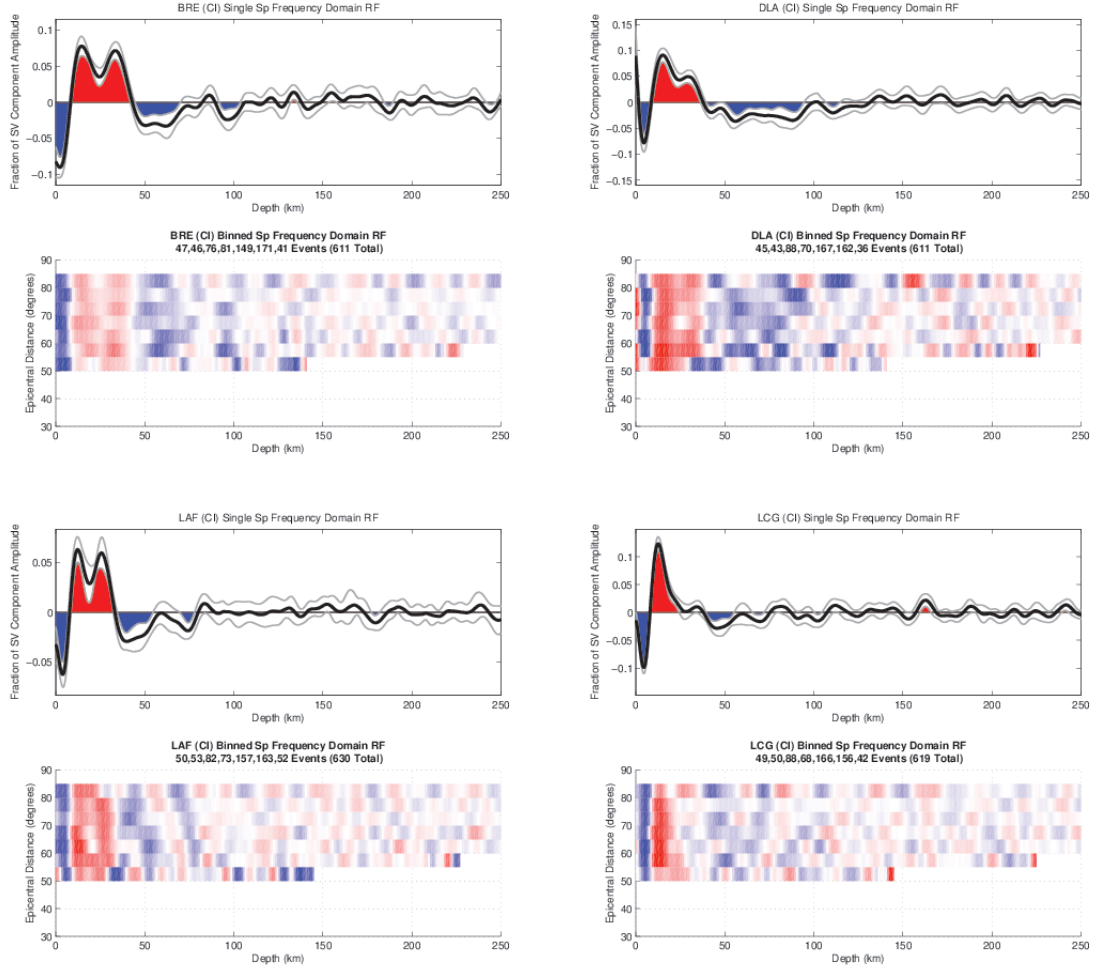


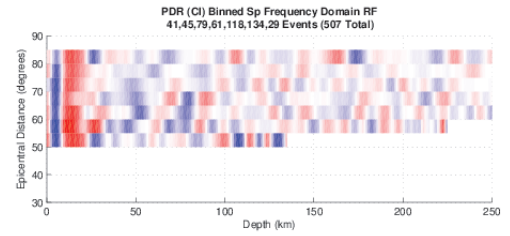
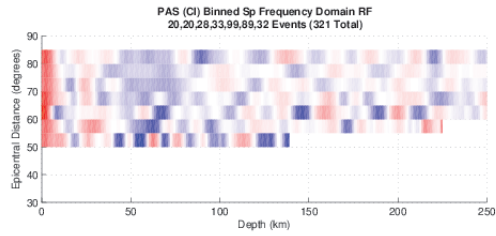
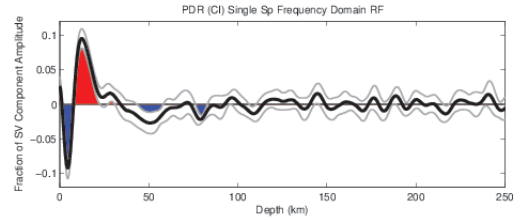
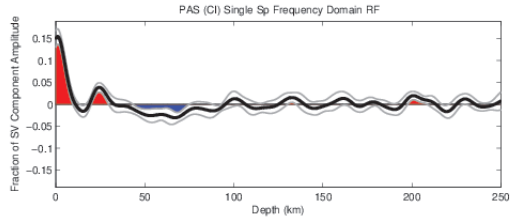
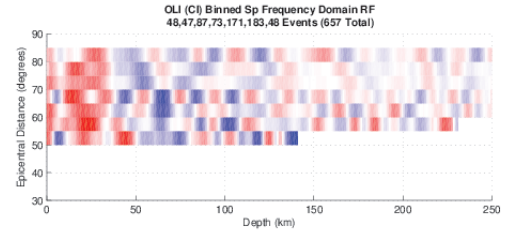
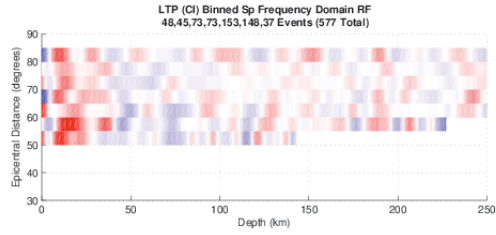
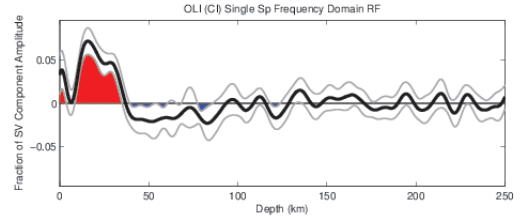
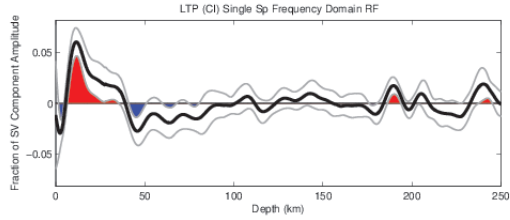
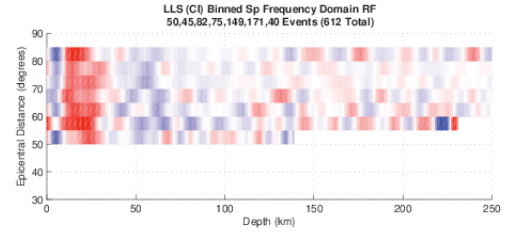
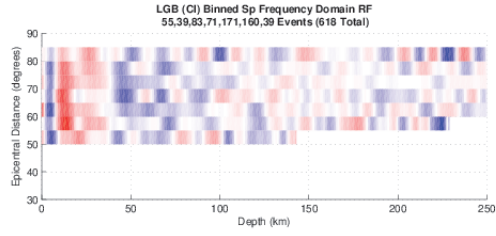
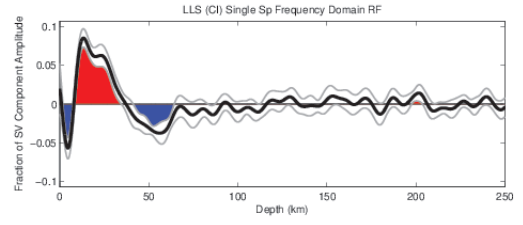
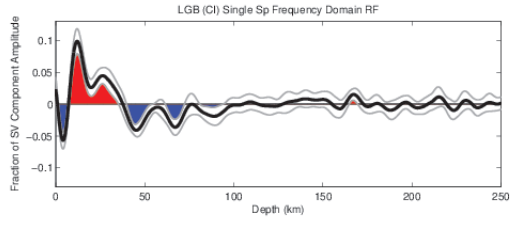


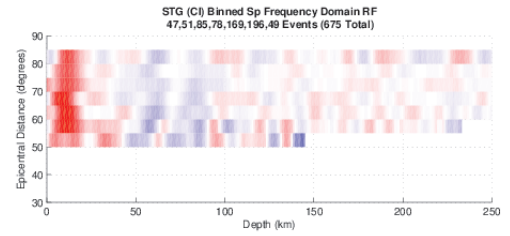
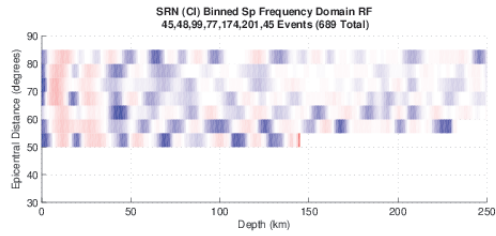
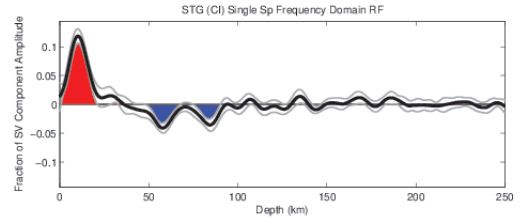
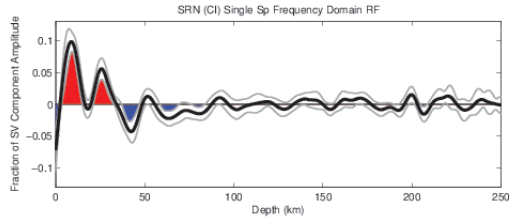
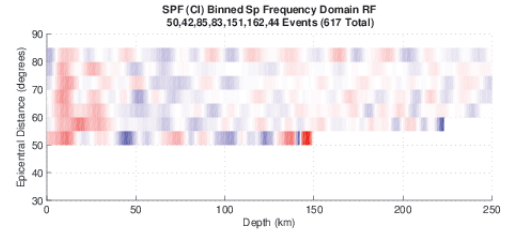
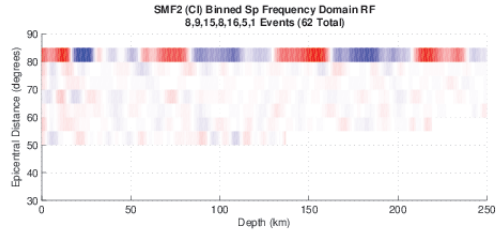
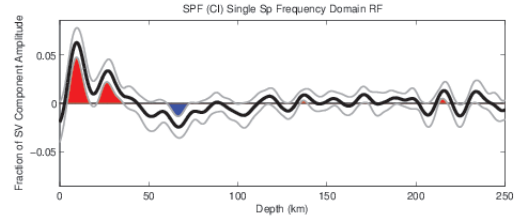
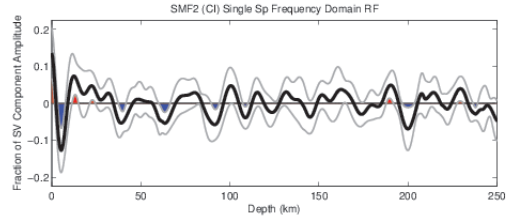
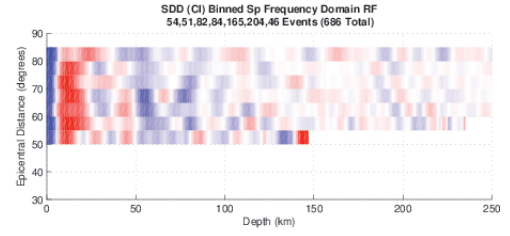
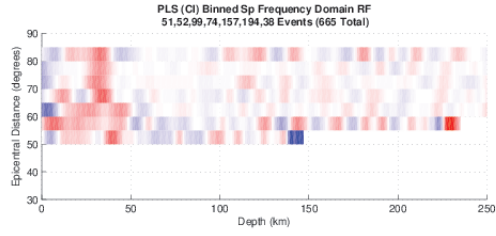
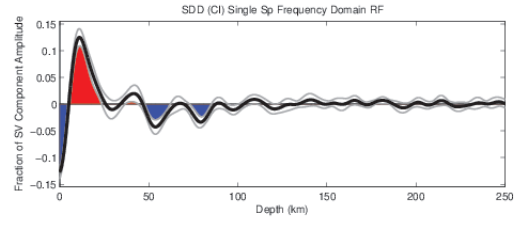
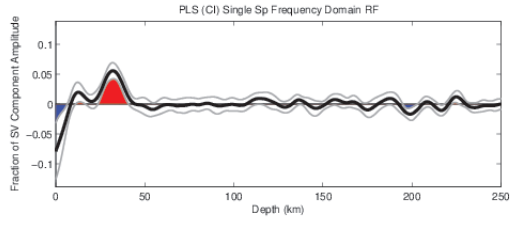


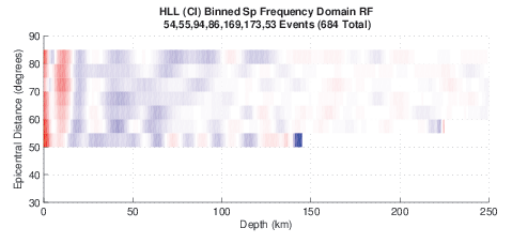
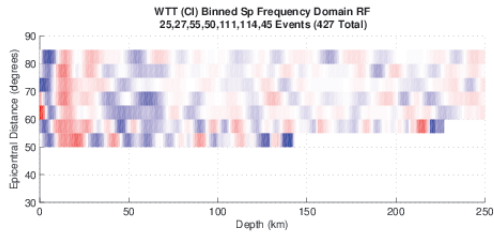
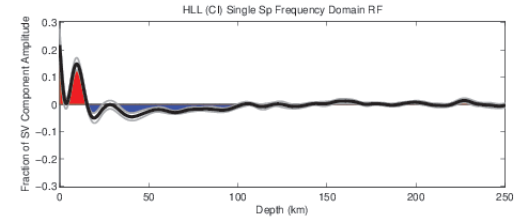
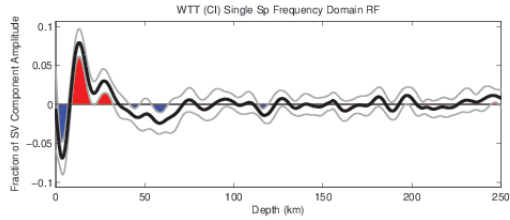
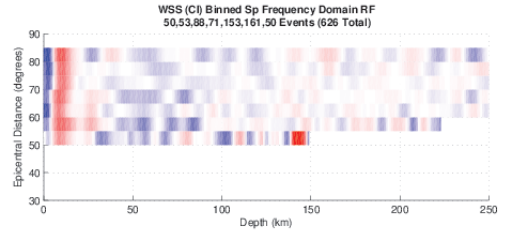
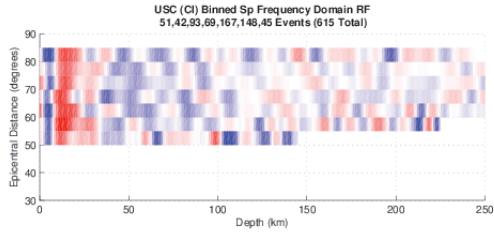
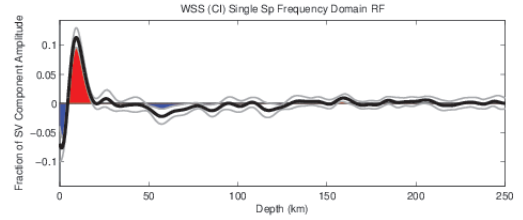
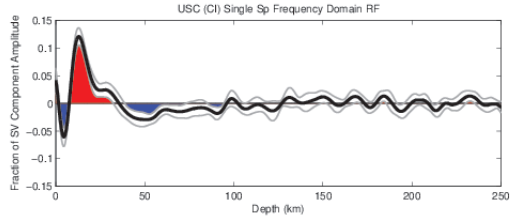
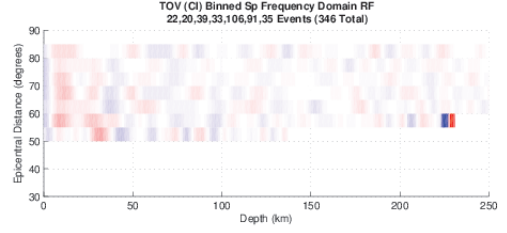
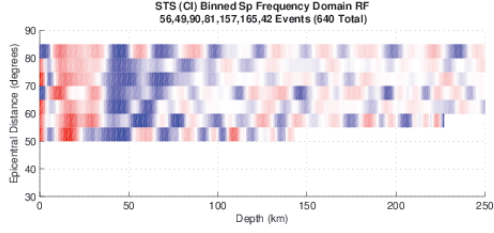
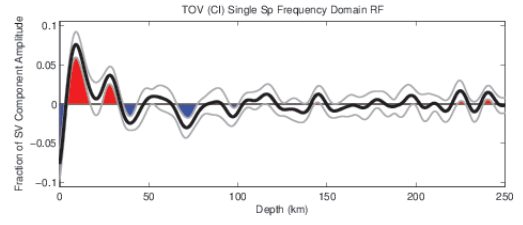
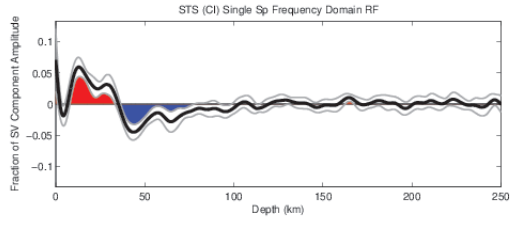
L.A. Basin

Figure B.15:



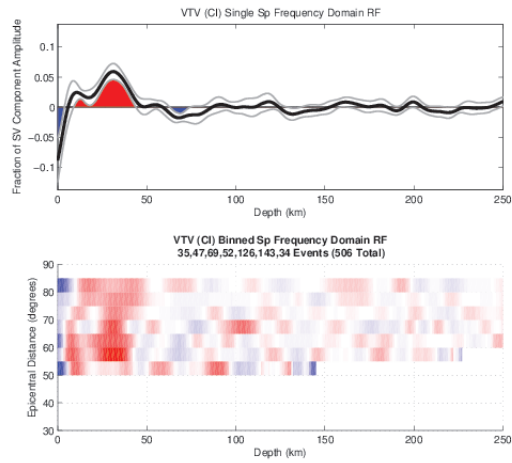






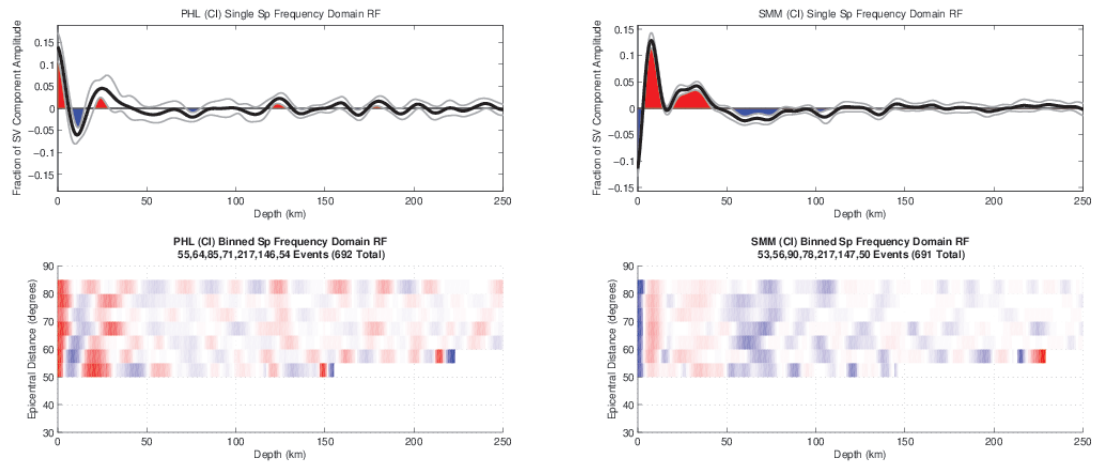
Control Station

Figure B.16:



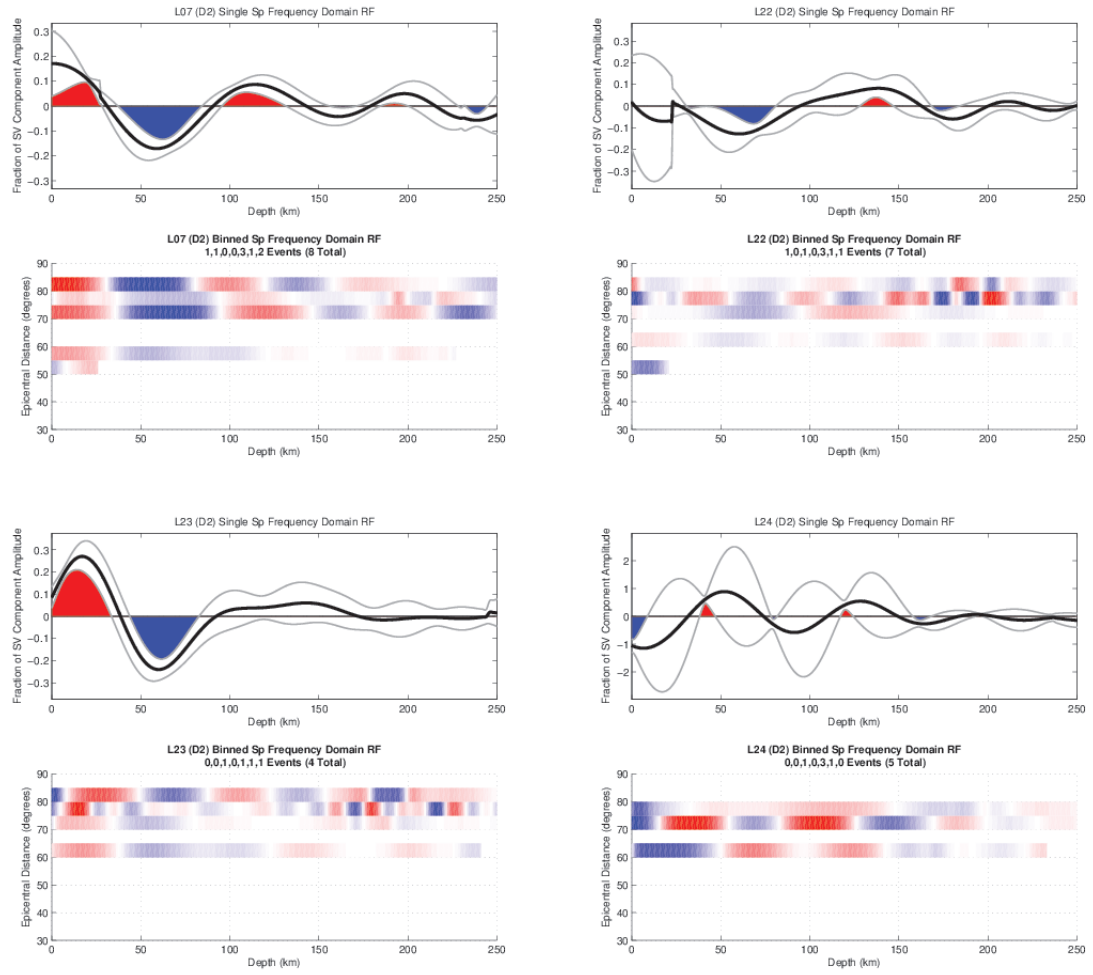
North of Western Transverse Range

Figure B.17:



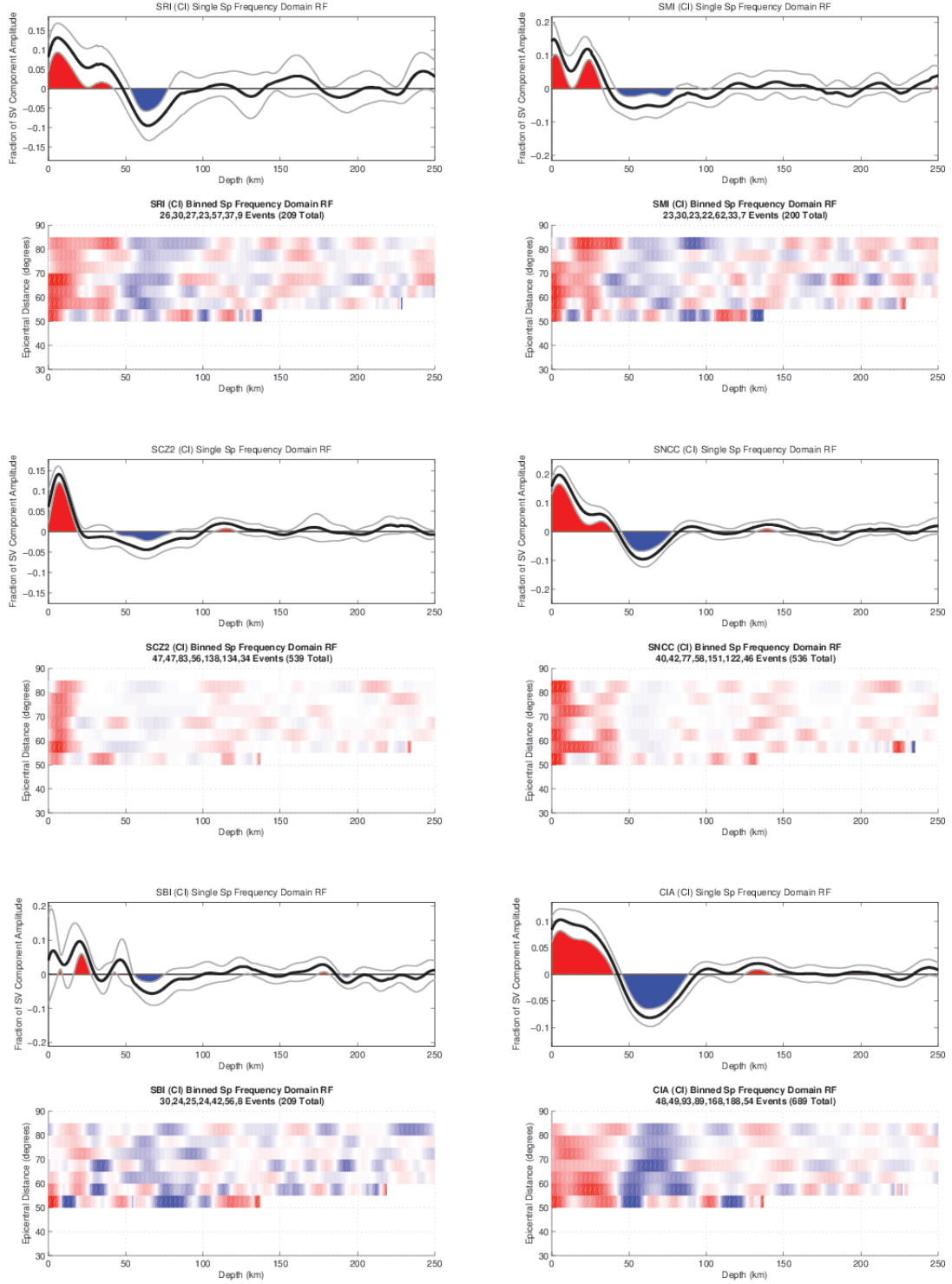
Stations inbetween abyssal plain and Outer Borderland

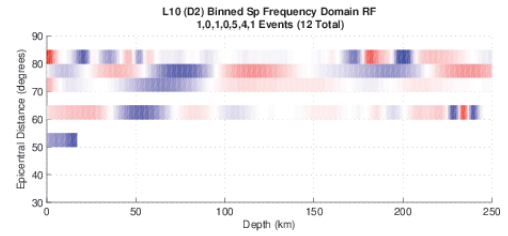
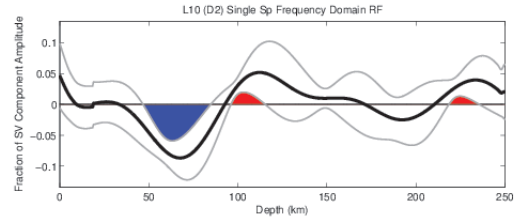
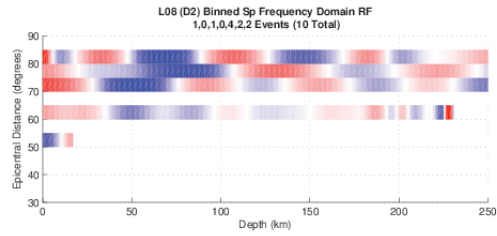
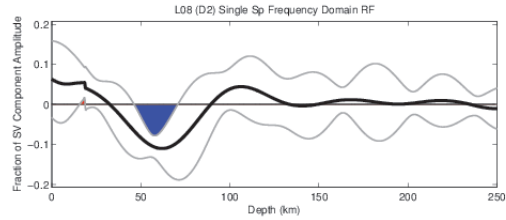
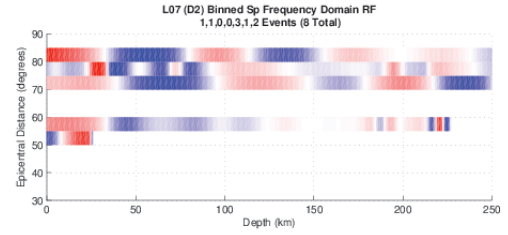
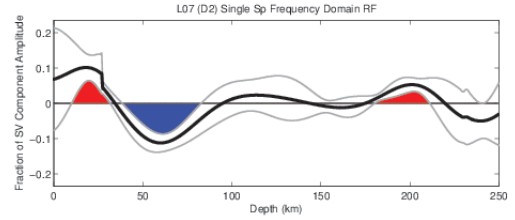
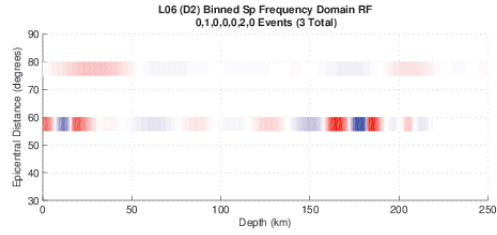
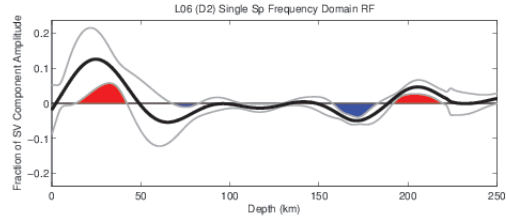
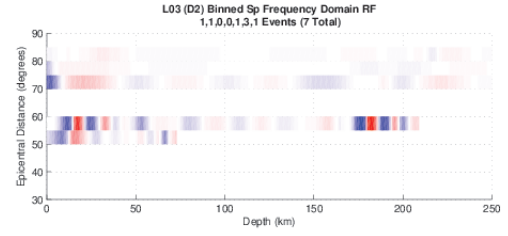
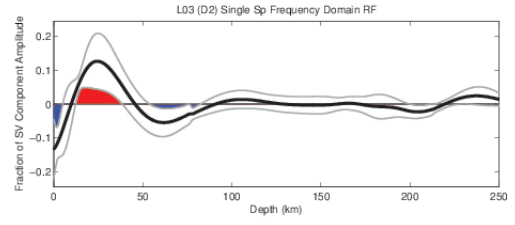
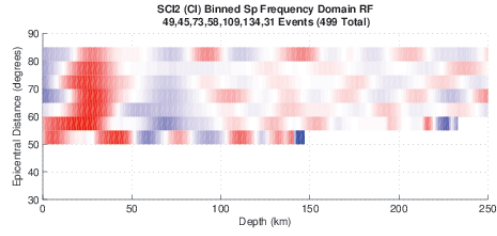
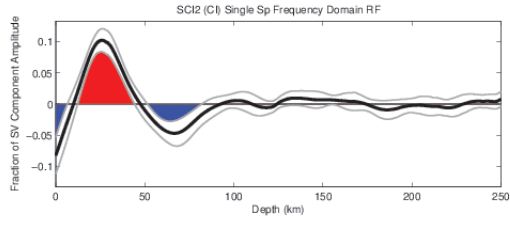
Figure B.18:

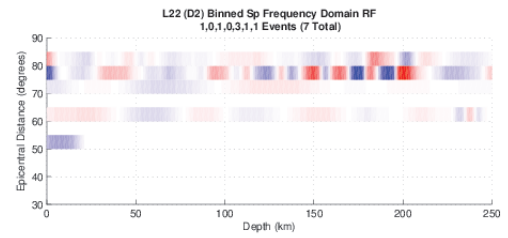
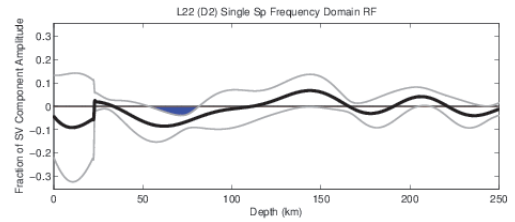
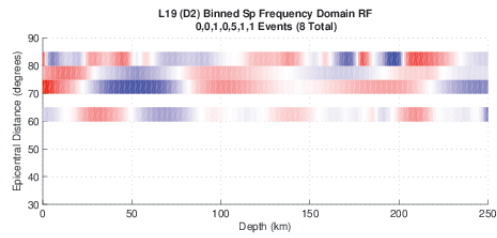
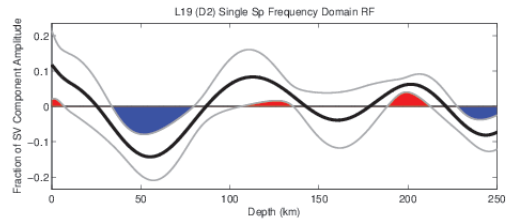
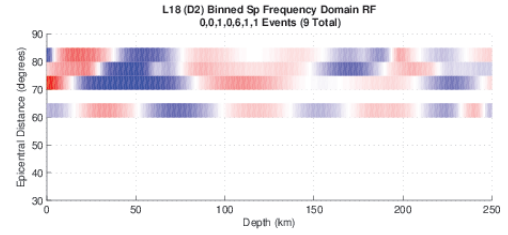
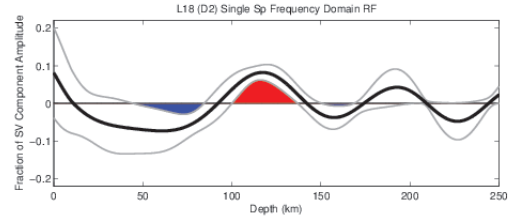
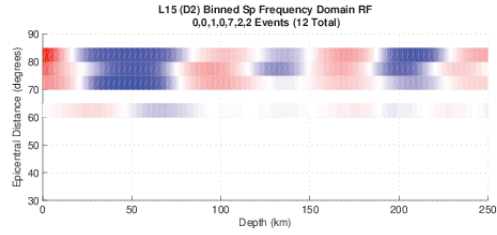
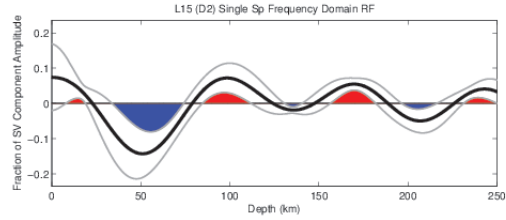
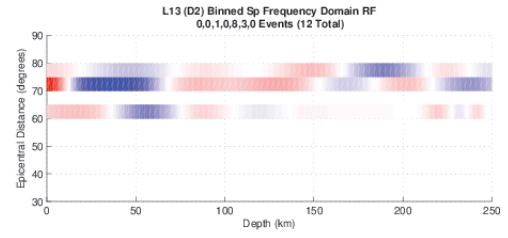
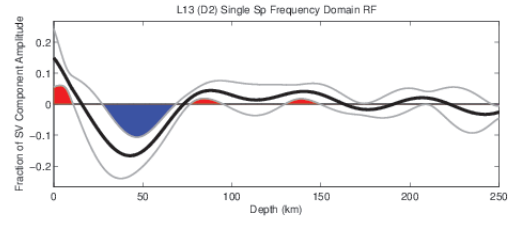
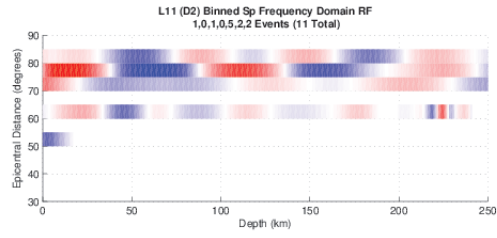
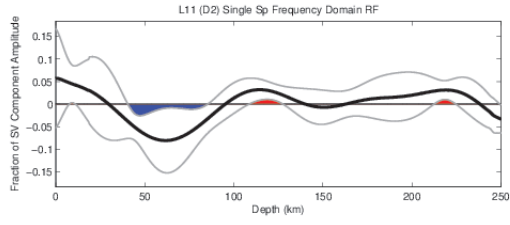


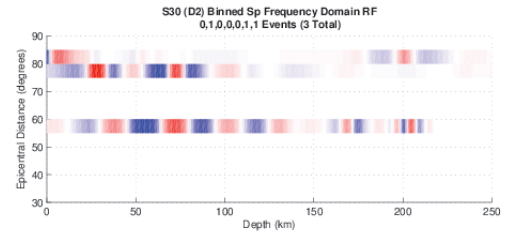
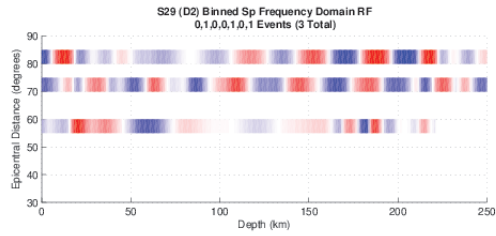
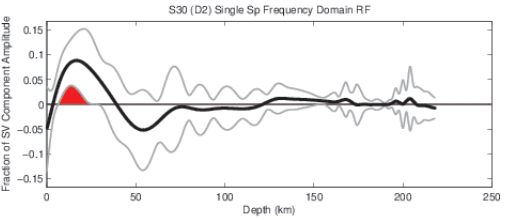
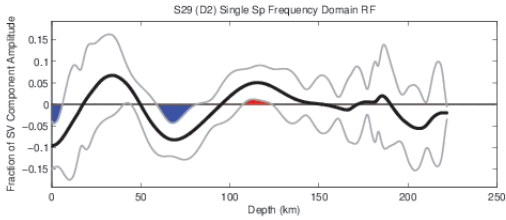
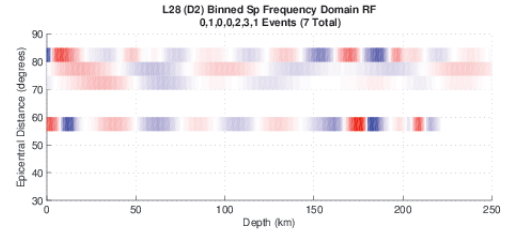
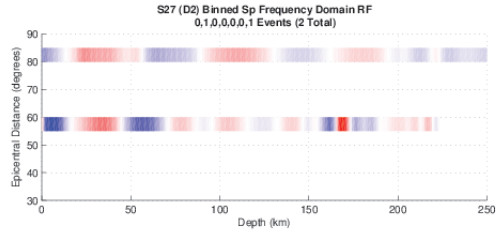
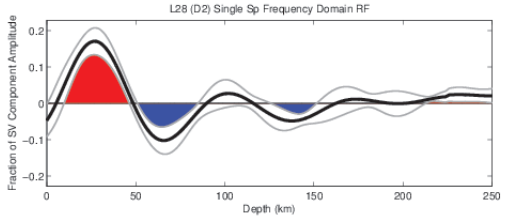
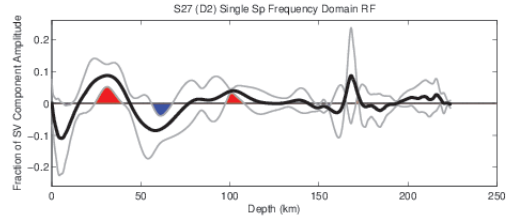
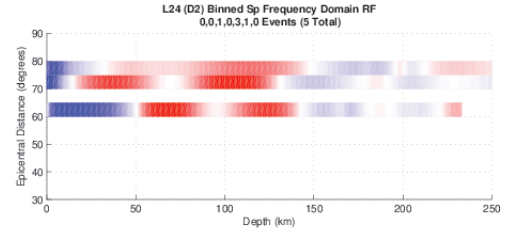
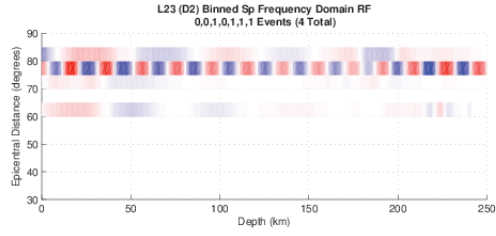
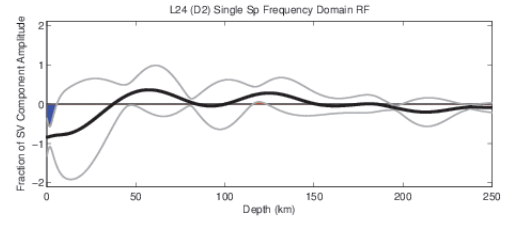
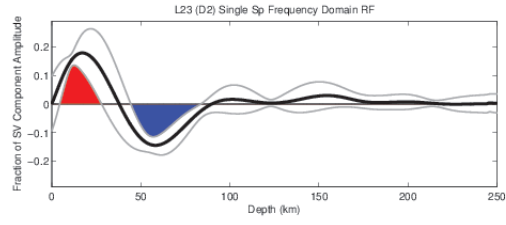
Sp with water level = 1

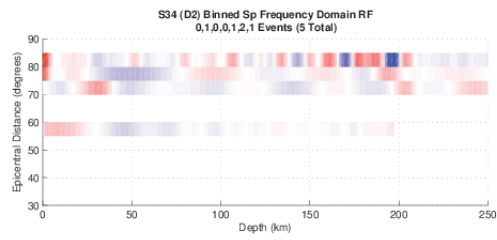
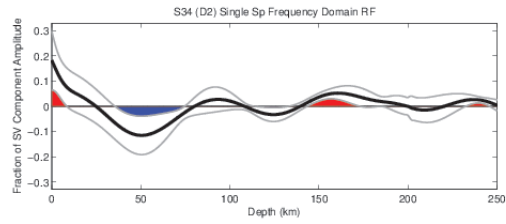
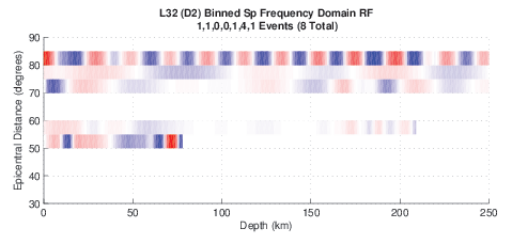
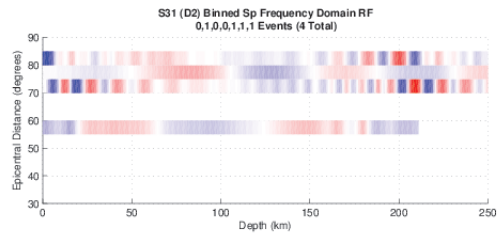
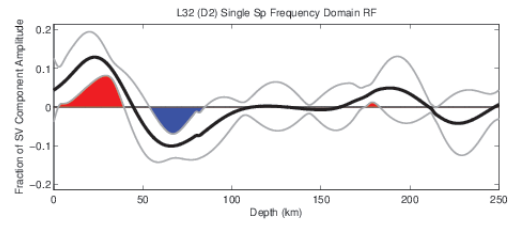
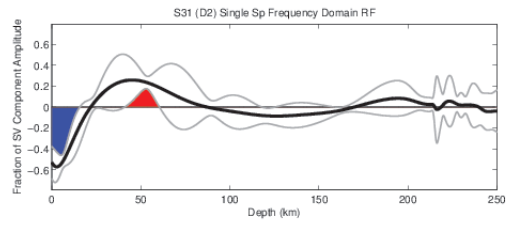
Figure B.19:











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