

Advances in Understanding the Lithosphere and Asthenosphere System: A Case Study of Seismological Structure in Alaska

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*Dedico esta tese de doutorado aos meus pais, Marinez e Amaro,
a quem sou eternamente grata por me mostrarem o valor da educação
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Chapter 1

Introduction

The system of plate tectonics governs the transport of material between Earth's interior, hydrosphere, and atmosphere to create a dynamic planetary surface that sustains life (NASEM, 2020; McGuire, J.J., T. Plank, et al. 2017). A fundamental component of this system is the behavior of the lithosphere and its relationship with the underlying asthenosphere (e.g., Fig. 1.1).

Since the early years of plate tectonic theory, we have known that the lithosphere, a layer formed by the crust and the rigid upper mantle, rests above the weaker mantle of the asthenosphere. This arrangement permits the lithosphere to migrate over the more deformable asthenosphere, and this interaction alone has significant ramifications for the Earth and life as we know it today.

1.1 Scientific Questions

Despite the importance of this arrangement, there are multiple aspects of the lithosphere, asthenosphere, and their interactions that remain elusive, especially at plate boundaries. In the case of Alaska and many other regions globally, subduction and the lithosphere-asthenosphere system are yet to be fully understood. Subduction zones are regions where one lithospheric plate converges with another, and the resulting earthquakes,

tsunamis, and volcanoes can pose significant hazards globally and to local communities. In Alaska, surface signatures range from patterns of deformation observed from GPS and earthquake slip data (Finzel et al., 2015) to features such as terrain uplift, mountain building, and large-scale fault systems. However, these features, together with their relationship to the underlying lithospheric structure are debated and not yet understood in the context of plate tectonics. By studying the lithosphere and asthenosphere in Alaska, this thesis aims to understand the role of plate tectonics in subduction zones and how it influences the deformation and surface processes we observe.

1.2 Approaches to Understanding the Lithosphere and Asthenosphere

Various geophysical approaches have examined the lithosphere and the asthenosphere to understand their roles in plate tectonics (Evans et al., 2005; Naif et al., 2013; Wang et al., 2020). In addition to the (i) crust-mantle boundary (or Moho), which provides insight on crustal buoyancy and isostasy as well as fault processes, (ii) vertically localized gradients in seismic velocity (discontinuities) in the mantle, such as the lithosphere-asthenosphere boundary (LAB) and mid-lithospheric discontinuities (MLDs), have also been studied (Krueger, et al., 2020). The very existence of these discontinuities provides information about the lithosphere and asthenosphere, but there is still debate on their origins. Vertical variations in temperature and pressure play a role in creating a strong (high-velocity) lithosphere and weak (low-velocity) asthenosphere (Hirth and Kolstedt, 2003; Stixrude and Lithgow-Bertelloni et al., 2005). Additionally, vertically-localized velocity gradients may arise from heterogeneity in (or the presence of) chemical composition (Hirth and Kolstedt, 2003; Cline et al., 2018; Rader et al., 2015; Selway et al., 2015), grain size (Faul and Jackson, 2005), elastically accommodated grain boundary sliding (Karato et al., 2015), near melt conditions in the asthenosphere (Yamauchi and Takei, 2016), partial melt in the asthenosphere (Debaille et al., 2020) and anisotropy (Beghein et al., 2014; Rychert et al., 2007). None of these models, however, can explain the vast range of

independent findings that show a sharp LAB, including slow seismic velocity anomalies (Abt et al., 2010; Gama et al., 2020) and low electrical resistivity anomalies (Naif et al., 2013; Wang et al., 2020) in the asthenosphere.

Seismological data allows insight into the characteristics of the lithosphere-asthenosphere boundary (LAB). Surface and body wave tomography provide constraints on seismic velocities throughout the volumes of the lithosphere and asthenosphere but cannot resolve sharp velocity gradients (Shapiro and Ritzwoller, 2002; Schmandt and Humphreys, 2010; Obrebski et al., 2011; Rawlinson and Fishwick, 2012; French et al., 2013; Golos et al., 2018). Active source seismic data provide finer-scale vertical resolution, but the ability to reach LAB depths is relatively rare (Bostock, 1998; Thybo, 2006; Kawakatsu et al., 2009; Rychert et al., 2012; Wang et al., 2021). However, some types of teleseismic body waves have generated key constraints on LAB velocity gradients on a widespread basis, including converted seismic body waves P-to-s and S-to-p receiver functions (Kumar et al., 2005; Li et al., 2004, Rychert and Shearer, 2009; Kawakatsu et al., 2009; O'Driscoll and Miller, 2015), reflected body waves such as SS precursors (Tharimena et al., 2017; Rychert and Shearer, 2011; Schmerr 2012) or topside SH reflections (Shearer and Buehler, 2019; Liu and Shearer, 2021), and high-frequency coda waves (Shito et al., 2015; Kennett et al., 2017).

1.3 The Seismic Approach

Among these different approaches, S-to-P (or Sp) converted waves, which are seismic body waves that propagate as S-waves and convert to P-waves at boundaries or gradients, are suitable for mapping the LAB for various reasons. Converted phases have higher frequencies than surface waves (e.g., Rychert et al., 2007), and they offer information from the region underneath a seismic station. These conversions happen at LAB depths to a radius of 100-200 km for Sp and 20-40 km for Ps around the station, depending on how close to vertical the rays are when they hit the boundary.

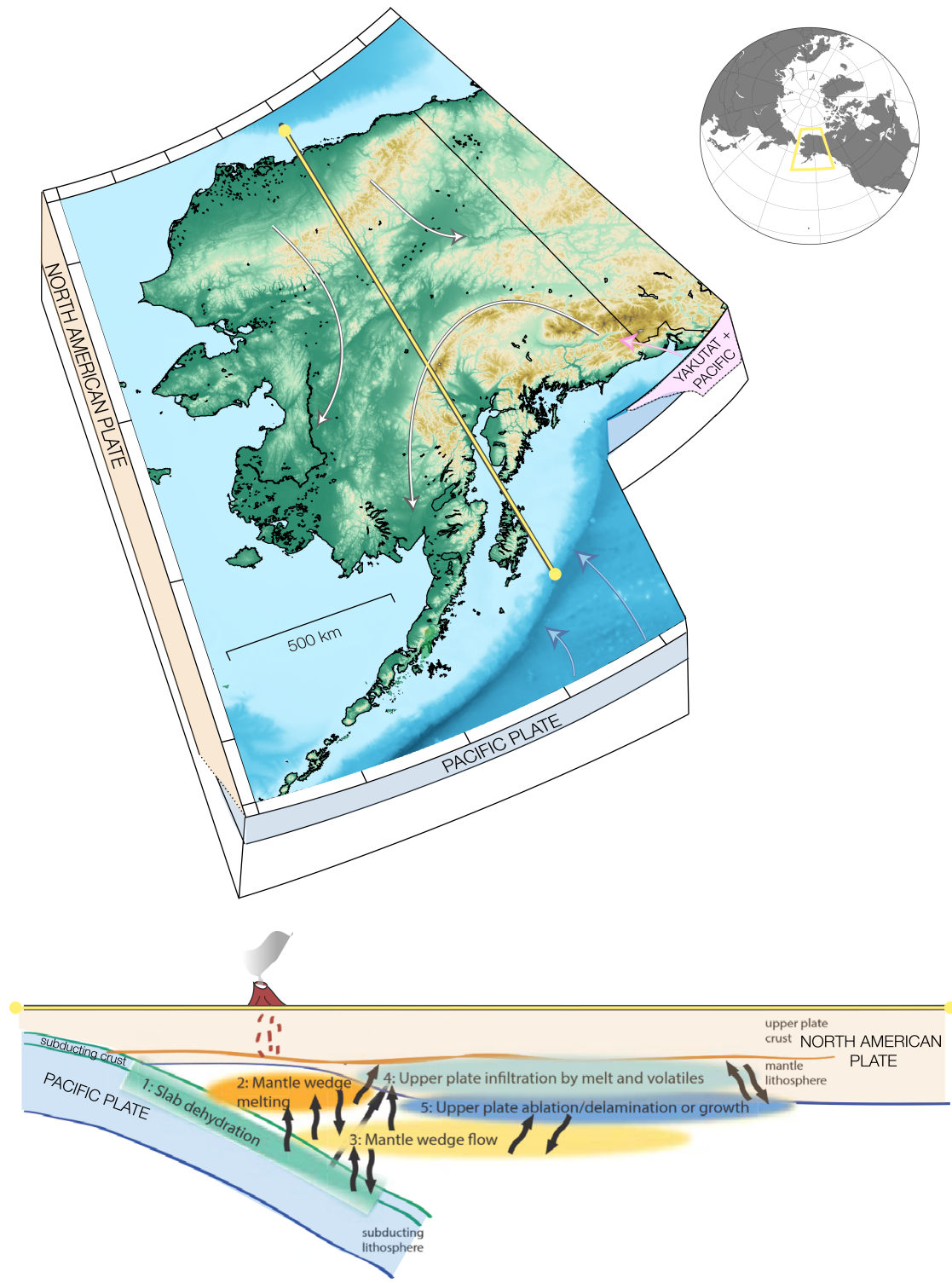


Figure 1.1: Top: Alaska tectonic setting with topography. White arrows are deformation patterns after Finzel et al. (2015). Pink and blue arrows at the bottom indicate subduction direction. Yellow line is marking where an assumed location would be for a cross section shown below. Bottom: Schematic of a subduction zone cross section with the Pacific plate, North American plate, and the interactions that are likely happening between them (Modified from Karen M. Fischer, personal communication).

Despite Ps having good lateral sensitivity, Sp benefits from the absence of reverberations induced by shallow structure in its waveform, which can hide LAB phases in Ps converted phases. Sp phases are key to constrain LAB depth.

The combination of LAB depth constrained by Sp phases and absolute velocity constraints from surface wave phase provides a more robust understanding of the LAB. To do this, I use Bayesian statistics, which is a powerful mathematical approach that has been extensively implemented to study inverse problems with multiple data types. I show in this thesis how I calculated a Bayesian joint inversion of Sp converted body waves and Rayleigh waves phase velocities (based on Eilon et al., 2018) to statistically probe the range of seismic velocity parameters that provide good fits to the data, enabling me to address critical scientific questions. Here, I apply this method to Alaska, but the codes and methodology can be applied anywhere, regionally or globally.

1.4 The Thesis

In this thesis, I used seismic data to shed light on the lithospheric and asthenospheric structure near the subduction zone and the continent in Alaska to depths of 300 km. The contrast between an active convergent margin and a more stable continental interior in Alaska highlighted the variation of lithospheric and asthenospheric properties throughout these regions. I organized the work in each one of the chapters in a complimentary manner.

In Chapter 2, from seismic waveform data recorded by permanent stations in Alaska, I used a joint inversion of converted waves and Rayleigh wave data to probe the questions: (1) How has subduction modified the upper plate lithosphere? (2) What are the accreted terranes' mantle signatures? (3) What differences exist between the subducting plate carrying the Yakutat terrane and the Pacific plate lithosphere to the west? (4) What are the origins of the Denali Volcanic gap?

To answer these questions, I performed a receiver function analysis of Ps and Sp converted body waves and combined the Sp receiver functions with Rayleigh wave phase velocity maps in the Bayesian joint inversion that resulted in velocity profiles for individual stations. I tested the performance of the inversion with synthetic data that simulated

the structure similar to what I found in Alaska. These synthetic tests illustrate that Sp receiver functions enhance the resolution of the velocity gradients at the Moho and the lithosphere-asthenosphere boundary. At the same time, Rayleigh surface waves provide information about absolute velocities. Also, Sp worked well when mapping the Moho, so Ps phases were not included for the final inversion results to avoid the addition of Ps reverberations. To create high-quality Sp profiles, I used data from 31 permanent stations operating for at least six years, ensuring data from thousands of different earthquakes. Although these restrictive criteria guaranteed that I obtain robust velocity model results, this relatively small number of permanent stations in the whole state limited the spatial detail of our analysis and interpretation.

Nonetheless, by interpreting lithospheric and asthenospheric properties along key transects, I addressed the target questions, finding that, for questions (1) and (2), the continental lithosphere is thinnest (60km), and the asthenosphere has the slowest velocities in central Alaska, above the shallow slab. Under the Brooks Range and the northern Arctic Alaska terrane, the continental lithosphere thickens to the north, reaching 110 to 130 km and lithospheric velocities comparable to Archean cratons. This trend is compatible with a northward decrease in upper plate modification by melt and volatiles from the slab and variances in lithospheric terranes, mantle velocity, and viscosity. This suggests that, while subduction-related flow may reach these latitudes, possibly weakening the lithosphere, subduction-related processes have not clearly changed Alaska's interior mantle lithosphere velocity structure. For question (3), the subducting plate carrying the Yakutat terrane and the Pacific plate lithosphere to the west show different thicknesses that are due to the $\sim 20\text{-}30$ km of additional crust from the Yakutat plate (Christeson et al., 2013) and its ~ 10 km thinner thermal mantle expected from the age difference (20 My old versus 40 My old Pacific plate), according to plate cooling model, predicting a total thickness that is about ~ 20 km greater than that of the Pacific plate. The Denali Volcanic Gap is located in the thin lithosphere zone in central Alaska, but I could not definitively resolve its origins for question (4). Instead, I discussed that Denali Volcanic Gap models, including feedback between subduction zone processes and the upper plate, agree with LAB depths. I saw that asthenospheric velocities above the subducting plate

within the Denali Volcanic Gap are not significantly different from those beneath adjacent magmatically active regions, implying that lower slab fluid fluxes and reduced wedge partial melt content were not the dominant factors in the formation of the Denali Volcanic Gap unless these effects have unobserved effects on Vs.

In Chapters 3 and 4, waveforms from the National Science Foundation (NSF) EarthScope Transportable Array (TA) provided the opportunity for more continuous 3D seismic velocity models for much of the state of Alaska. The transportable seismic array recorded 2-5 years of data at temporary stations across Alaska, concluding data collection in May 2021. In Chapter 2 I stacked Sp phase receiver functions with earthquake data that arrived at a single station, which allowed me to map the structure mainly beneath the station. In Chapter 3, I performed common-conversion-point (CCP) stacking to map Sp conversions in a continuous 3D volume beneath Alaska.

Chapter 3 questions guided my work: (1) What are the Moho and LAB patterns observed in Alaska at a higher resolution? (2) How do upper plate structure and tectonics relate at and north of the subduction zone? (3) What mechanisms of melt generation could be taking place in Alaska?

To answer these questions, I used the CCP method to create a dense distribution of ray paths beneath Alaska by taking advantage of many stations that record multiple rays and that leave a "common point" with a high density of data where seismic rays pass through the boundaries. This chapter used 20210 robust Sp receiver functions to construct the 3D CCP volume. This result is, to my knowledge, the first and most complete map of crust and upper mantle discontinuities from Sp phases recorded by the EarthScope TA in Alaska, in contrast to prior Sp studies (e.g., O'Driscoll and Miller, 2015) that did not use the TA stations. The results provide constraints on the depths of the Moho, the lithosphere-asthenosphere boundary, and a positive velocity gradient (PVGm) in the mantle. To address question (1), Moho depth observations outline the edge of the thick crust of the underthrust Yakutat terrane, a significant drop in crustal thickness to the north over the Denali fault, and the thicker crustal root beneath the Brooks Range in the north. I continued answering question (1) and addressed question (2) by comparing new constraints on the negative velocity gradient at the LAB to define

a zone of particularly thin lithosphere just beyond the position of the deepest seismicity in the central Alaska subduction zone, which is consistent with upper plate thinning caused by subduction zone flow and melt rising from the mantle wedge. LAB amplitude diminishes just above the deepest seismicity, most likely due to increased infiltration of melt and heat into the upper plate, reducing the velocity contrast between the lithosphere and the asthenosphere. The thin lithosphere extends far northwest and north of the subduction zone, passing through the Seward Peninsula and reaching the southern border of the Brooks Range, where it meets thicker, colder, and higher viscosity lithosphere under the northern Arctic Alaska terrane. While historical extensional tectonics, and maybe lithospheric foundering may have contributed to the thinning of the lithosphere under the Seward Peninsula and interior Alaska, continuous ablation of the top plate by asthenospheric flow most certainly contributes to the lithosphere remaining thin. A notably new feature observed in this investigation is a positive velocity gradient that indicates the base of a low-velocity layer whose upper border corresponds to the LAB. This finding led to question (3), which I answered by dividing the regions into five areas. I discussed mechanisms of melt generation due to partial melting that shows the signature of subduction processes (i) in the primary subduction zone and (ii) around the Wrangell Volcanic Field (WVF) in the east; partial melting from mantle decompression due to (iii) mantle flow around the edge of the slab or a slab tear, with (iv) upwelling flow beneath the Seward Peninsula related to the extrusion of southwest Alaska; and partial melting from mantle decompression (Elliott and Freymueller, 2020). In the regions where slab-derived fluids appear to be lacking, carbon-assisted melting may still be possible (Hammouda et al., 2021). Besides the four previously stated locations, (v) at depths of up to 200 km, weak and intermittent PVGm phases have been found beneath the subducting lithosphere and parts of the thick lithosphere in northern Alaska. If mantle melting is involved in forming these velocity gradients, the base of a layer containing carbon-assisted partial melt is one possible explanation (e.g., Hammouda et al., 2021).

In Chapter 4, I returned to the Bayesian joint inversion approach I employed in Chapter 2. However, I applied it to the entire 3D volume from the CCP stacking, combined with surface wave phase velocities, to result in a 3D velocity model of the crust and mantle

beneath Alaska from Chapter 3. These new 3D models enabled me to explore the scientific questions: (1) Where do offsets in upper plate properties occur? (2) What is the depth extent of shear zones? (3) How do major faults and terrane boundaries play a role in terrane reconstruction and deformation?

In this chapter, I calculated seismological constraints based on decades (1986-2021) of broadband recordings of earthquakes from permanent, temporary, and the Earth Scope Transportable seismic arrays in Alaska. The dataset, which includes the results of ~ 300 inversions, provides a higher resolution of the 3D seismic velocity of the Alaskan crust and upper mantle than previously calculated. The results I analysed are V_s profiles with depth that were interpolated to create maps and cross-sections for interpretation.

For the first posed question, (1) I showed new seismic evidence for deep shear zones beneath many of Alaska's major faults and a sharp boundary of the lithosphere of the craton-like Arctic Alaska terrane in the north. I inferred that the laterally abrupt lithospheric boundary in the north at the Kobuk fault and southern margin of the northern Arctic Alaska terrane might limit the northern boundary of the present-day subduction zone impacts and the extent of deformation from subduction. I highlighted that Alaska's strike-slip faults and terrane boundaries offset the Moho and create gaps in the LAB signal, answering question (2) that shear zones extend to the mantle. This is important because, at the surface, fault geometries and plate movements are relatively well understood, but they are contested deeper in the crust (Lamb et al., 2015; Lay et al., 2016), and the shape of the plate boundary in the deep mantle lithosphere is still debated. To answer question (3), I called attention to the differences in lithospheric thickness and mantle shear wave velocity under terrane borders and major fault zones that indicate that distinct lithospheric blocks are juxtaposed where I see gaps in the LAB. These findings agree with the idea that the lithosphere deforms mainly or exclusively in concentrated deformation zones (Bercovici and Ricard, 2012), which contradicts theories that deformation is distributed and not localized at faults (Wilson et al., 2004; Zietlow et al., 2014). While speculative, we observe a high velocity block south of the Brooks Range that might represent lithospheric foundering, which might help favor one of the Alaska reconstruction models that are highly debated.

The work in Chapter 4 also lays the foundation for the integration of seismological and geodynamical constraints with my collaborators that is planned for the future. This future collaboration will use shear velocity and attenuation to place bounds on temperature, bulk composition, grain size, water content, and partial melt. These parameters will allow us to calculate viscosity and density for geodynamical modeling of surface deformation. In addition, these geodynamic models will be compared with observations of azimuthal anisotropy in Rayleigh wave phase velocities and SKS splitting from published studies (e.g., Venereau et al., 2019; McPherson et al., 2020) to shed light on how all our findings relate to mantle flow.

To conclude, I studied the lithosphere and asthenosphere in Alaska, aiming at understanding the role of plate tectonics in subduction zones and how it influences the deformation and surface processes we observe. Among my many conclusions, these stand out as the most influential new insights.

- It is not surprising that subduction processes thin the upper plate, but I showed that this interaction extends to higher latitudes in Alaska than previously thought, including thin lithosphere and a low velocity asthenosphere that extends all the way to the Seward Peninsula and the southern margin of the northern Arctic Alaska Terrane. In these far-flung regions, surface magmatism is likely due to decompression melting in upwelling asthenosphere, as opposed to the mantle melting creating by slab-derived fluids that occurs closer to the subduction zone.

- While many studies have shown that the lithosphere beneath the northern Arctic Alaska terrane is relatively thick, my work shows that the transition from thin to thick lithosphere occurs over a surprisingly small horizontal distance, resulting in a steep southern boundary to the thick lithosphere in the north. The high velocity and depth extent of this lithosphere allow its characterization as a “mini-craton.”

- I also showed that the structure of the upper plate lithosphere is perturbed beneath many of the major fault zones and terrane boundaries in Alaska, and that these signatures persist all the way to the base of the lithosphere. These results support the view that mantle lithosphere can deform in thin concentrated zones of shear.

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Chapter 2

Shear Wave Velocity Structure beneath Alaska

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Summary

- The present-day lithospheric structure of Alaska is the result of a unique tectonic history of subduction and terrane accretion that controls upper plate thickness and rheology.
- We jointly inverted Sp receiver functions and Rayleigh wave phase velocities to calculate 31 shear velocity profiles across Alaska.
- Tests of the Bayesian joint inversion with synthetic data illustrate that Sp receiver functions enhance the resolution of the velocity gradients at the Moho and the lithosphere-asthenosphere boundary, while Rayleigh surface waves provide information about absolute velocities.

- Our results show that in central Alaska, above the shallow slab, the continental lithosphere is thinnest (~ 60 km) and the asthenosphere has its lowest velocities.
- The continental lithosphere thickens to the north, reaching values of 110 to 130 km, with high lithospheric velocities that are comparable to Archean cratons.
- This pattern is consistent with a northward decrease in upper plate modification by melt and volatiles derived from the slab, in addition to intrinsic mantle velocity and viscosity differences between inherited lithospheric terranes.

2.1 Introduction

Subduction of the Pacific plate and Yakutat microplate beneath the North American and Bering plates deforms southern Alaskan lithosphere, uplifting mountain ranges such as the Alaska Range (e.g. Jadamec et al., 2013; Finzel et al., 2015) (Fig. 2.1). In contrast, the lithosphere in northernmost Alaska has been hypothesized to be thicker, stronger and less deformable (e.g. O'Driscoll and Miller, 2015; Finzel et al., 2015). However, uncertainty remains about how the transition between these domains is reflected in continental lithospheric structure, and the role of subduction in modifying the upper plate (e.g. Rondenay et al., 2010). In this study, through joint inversion of surface wave and converted body wave (S_p) data, we obtained new constraints on the seismic velocity structure in the crust and mantle of Alaska in order to better understand the relationship of subduction and the upper plate mantle.

2.1.1 Alaskan Subduction and Terrane Accretion

The North American plate in Alaska is a complex mosaic of accreted terranes (e.g. Colpron et al., 2007; Fuis et al., 2008; Colpron and Nelson., 2011) (Fig. 2.1b). At the southern margin of Alaska, the Yakutat terrane is a subducting oceanic plateau with anomalous thick crust (e.g., Fuis et al., 2008; Christeson et al., 2013), while the subducting plate to its west is normal Pacific oceanic lithosphere. Following Colpron et al. (2007) and Colpron and Nelson (2011), the terranes in Alaska and neighboring Canada

can be categorized into four major groups (Fig. 2.1b): i) the lithosphere of ancestral North America, including both the craton and continental margin of Laurentia; ii) peri-Laurentian terranes that largely represent Devonian to Jurassic arcs formed at the margin of ancestral North America; iii) the Neoproterozoic to Jurassic Northern Alaska and Insular Terranes, which did not derive from the western margin of Laurentia, but instead from other continents, or possibly from further east in Laurentia in the case of the northern portion of the Arctic Alaska terrane that contains much of the Brooks Range; and iv) the Late Accreted terranes that docked in the late Mesozoic to Cenozoic and include the Yakutat. The various tectonic origins of these terranes likely produced lithosphere with different thicknesses and internal properties, and these properties in turn influence the evolution of the lithospheric blocks.

2.1.2 Prior Seismic Studies in Alaska

Numerous previous studies have imaged the crust and mantle structure of Alaska using receiver functions (RFs) (e.g., O'Driscoll and Miller, 2015; Rondenay et al., 2010; Kim et al., 2014; Miller et al., 2018; Zhang et al., 2019), surface wave tomography (e.g., Wang and Tape, 2014; Jiang et al., 2018; Feng and Ritzwoller, 2019), body wave tomography (e.g., Eberhart-Phillips et al., 2006), and S(X)S splitting studies (e.g., Venereau et al., 2019; McPherson et al., 2020). Some have combined multiple datatypes that include Ps receiver function data to provide additional constraints on crustal properties within the tomographic inversions (e.g., Ward and Lin, 2018; Martin-Short et al., 2018; Berg et al., 2020). However, to date, none have utilized data capable of capturing mantle velocity discontinuities at high resolution.

In this study we incorporate Sp phases (in the form of Sp receiver functions), ambient noise, and earthquake-based surface waves in a Bayesian approach (Eilon et al., 2018) to solve for crust and mantle structure in Alaska. To our knowledge, this study is the first to combine these types of data in Alaska. With the resulting shear-velocity models, we addressed four questions: (1) How do upper plate properties relate to the position of the subducting slab and terrane boundaries? (2) How far north does the influence

of subduction extend? (3) How do mantle wedge properties relate to volcanism? (4) What differences are observed between the subducting Pacific plate and thick Yakutat microplate lithospheres?

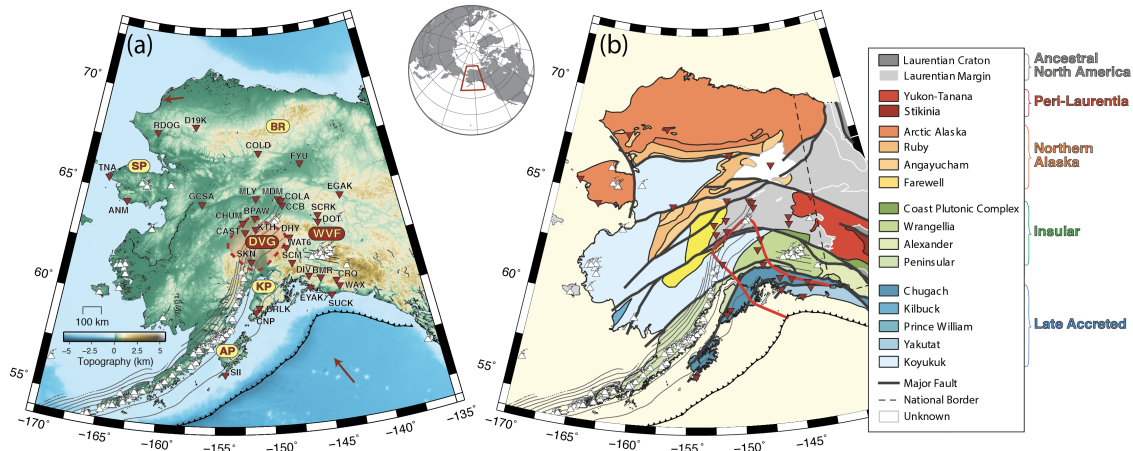


Figure 2.1: (a) Regional setting of Alaska plotted on ETOPO1 topography (Amante and Eakins, 2009). Regions with yellow background are: BR = Brooks Range; SP = Seward Peninsula; AP = Alaska Peninsula; and KP = Kenai Peninsula; and features with red background are: Wrangell Volcanic Field (WVF) and Denali Volcanic Gap (DVG), also marked with the red dashed line. In both figures, red inverted triangles represent the stations employed in this paper with their names (black text) and white triangles are volcanoes (avo.alaska.edu). Grey continuous lines mark seismicity contours from (Hayes et al., 2018) except near the WVF where they are from Kiara Daly, personal communication, 2020. The trench is shown as the black lines with ticks. Small red arrows in the northwest and southeast of the map indicate the direction of plate motion in the no-net rotation reference frame (Argus et al., 2011); rates are 45 mm/yr for the Pacific plate, and 23 mm/yr for the North American plate. (b) Tectonic terranes (Colpron et al., 2007; Colpron and Nelson, 2011). Black thick lines are major faults. Red line highlights the bounds of subducted Yakutat crust based on Eberhart-Philips et al. (2006) as well as the unsubducted Yakutat terrane.

2.2 Methodology

We incorporated seismic data types that offer complementary information in the joint inversion. To constrain absolute velocities, we used fundamental mode Rayleigh wave phase velocities measured from ambient noise and teleseismic earthquakes between periods of 5 and 140 s. To enhance resolution of velocity gradients, in the mantle in particular, but also at the Moho, we used Sp converted phases. Analysis of Sp conversion sensitivity kernels demonstrates these phases are particularly effective at resolving velocity gradients

with near-horizontal dips, such as the LAB (Hua et al., 2020). Because of the respective strengths of the Sp and surface wave data, the resulting shear velocity profiles contain new information on the depth of the LAB, as well as the average velocity of the lithosphere and the minimum velocities in the asthenosphere.

2.2.1 Body Wave Receiver Functions

We employed Sp RFs (SV deconvolved from P) in the joint inversions with the primary goal of enhancing the vertical resolution of mantle velocity gradients. Clear Sp RFs are easier to interpret at mantle depths compared to Ps because converted Sp phases arrive before the direct S wave, thus avoiding the complications due to crustal reverberations that can occur with Ps RFs (e.g., Eilon et al., 2018). However, we also calculated Ps RFs (P deconvolved from S) to verify that Moho depths from Sp and Ps receiver function stacks are consistent (Fig. A1). One challenge with Sp RFs is that they often have lower signal-to-noise ratios compared to Ps. To improve signal-to-noise ratios in receiver function stacks, we objectively selected RFs using K-means cluster analysis. To avoid artifacts such as side lobes when calculating RFs, we employed both a wide-band filter (2-100s) that minimizes the introduction of spectral ringing and time-domain deconvolution. The supplemental material contains more detailed discussion of the steps involved in the receiver function calculation.

We analyzed body wave data recorded at 31 permanent stations from the Alaska Regional Seismic Network (AK, doi:10.7914/SN/AK), the IRIS Global Seismograph Network (IU, doi:10.7914/SN/IU), and the USGS Advanced National Seismic System (US, US doi:10.7914/SN/US). We selected stations that had been recording for over six years to ensure large numbers of body wave arrivals and therefore high-quality receiver function stacks. For additional sampling in northern Alaska, we also analyzed data from station D19K from the NSF EarthScope Transportable Array (TA, doi:10.7914/SN/TA). The data from D19K indicate the promise of TA data for this type of joint inversion. However, because initial Sp receiver functions at a number of other TA stations had lower signal-to-noise ratios, joint inversions involving data from all Alaska TA stations is planned for

the future when more events will have been recorded. The waveforms in this study represent events with $M_w > 5.7$, depths < 100 km, and epicentral distances of $55^\circ - 80^\circ$ for Sp (and $35^\circ - 80^\circ$ for the Ps examples shown in the supplement). We grouped Sp and Ps phases into geographic clusters over 30° of back azimuth. The resulting stacks of receiver functions for these back-azimuthal slices represent average velocity structure over more limited geographical regions than would a stack for all back-azimuths, resulting in better lateral resolution of shear velocity structure when the receiver function stacks are incorporated in the joint inversion. The distribution of back azimuths is dominated by two groups, $220 - 250^\circ$ (southeast Pacific events) and $250 - 280^\circ$ (north and northwest Pacific events).

Examples of Sp RFs (Fig. 2.2 and Fig. A1) reveal a clear phase from a velocity increase in the time range expected for the Moho conversion (3-6 s), and another phase from a negative velocity gradient in the time range predicted for a conversion at the transition between lithosphere and asthenosphere (6-12 s). The example RF stacks in Fig. 2 are migrated to depth with the SEMum2 model (French et al., 2013) to illustrate observed phase depths, and RFs are shown both in time and depth in Fig. A1. When used in the inversion, the RFs are treated as functions of time. The times/depths of the potential LAB velocity gradients vary geographically, with phase peaks (which approximate gradient midpoint depths) ranging from 60 km to 120 km assuming the SEMum2 model for migration (French and Romanowicz, 2013).

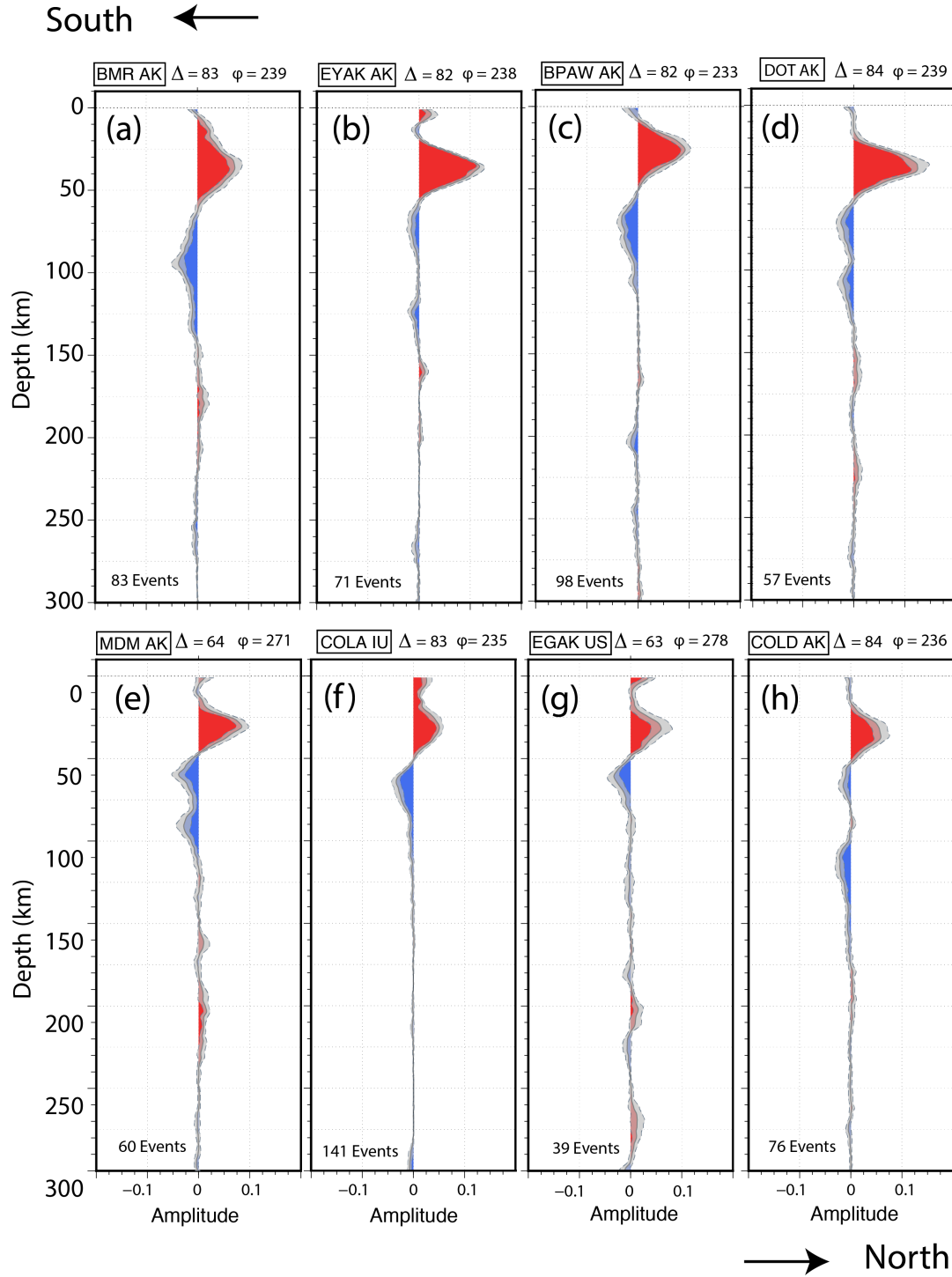


Figure 2.2: Examples of S_p receiver function stacks. Colored regions show the median of weighted individual receiver functions; grey shading shows one standard deviation uncertainties. S_p receiver functions were flipped to have the same polarity as P_s . Mean distance (Δ) of the events, number of events, and mean back azimuth (ϕ) above each panel. Stations are shown from southern (upper left) to northern Alaska (lower right). Positive (red) phases in the 30-40 km depth range correspond to the Moho; negative (blue) phases in the shallow upper mantle are potential LAB phases. Receiver functions in this figure were migrated to depth with the SEMum2 velocity model (French and Romanowicz, 2013).

2.2.2 Surface Wave Phase Velocities

We used fundamental mode Rayleigh wave phase velocities for Alaska obtained from ambient noise and earthquake sources. For Rayleigh waves generated by earthquakes, we followed the approach used by Babikoff and Dalton (2019). Travel times in the period range of 25-140 s were measured with interstation cross-correlation using the Automated Shear Wave Measurement System (Jin and Gaherty, 2015), and phase-velocity maps were determined with the Eikonal wavefront-tracking approach (Lin et al., 2009). For the joint inversions we also incorporate ambient noise phase velocities for periods from 5 to 25 s, following the approach of Ekström et al. (2009).

Rayleigh wave phase velocity maps (Fig. 2.3) reveal distinct regional variations. At periods of 25 s, where sensitivity is primarily within the crust, the lowest velocities occur beneath volcanically active regions in southern Alaska and the Brooks Range in northern Alaska. At longer periods, higher velocities occur in the vicinity of the northwest-dipping subducting lithosphere subducting to the northwest as well as in northern Alaska.

2.2.3 Bayesian Joint Inversion

We employed a transdimensional hierarchical Bayesian approach to determine robust and smooth 1D shear velocity models beneath each station using a modified version of the framework developed by Eilon et al. (2018). In each iteration, we created a synthetic Earth velocity profile, predicted the corresponding Sp converted phases and Rayleigh wave dispersion curves, and calculated the misfits of these predictions relative to the observed Sp RFs and Rayleigh wave phase velocities. A Monte Carlo Markov Chain (MCMC) search probed the model parameter space by randomly perturbing model parameters, and using a Metropolis Hastings criterion (e.g. Hastings, 1970), models were accepted if their misfits were lower than in the previous iteration, but models with higher misfits were also accepted with a probability that declined with increasing misfit. This approach created an ensemble of models within acceptable misfit levels. These steps are explained in the following sections.

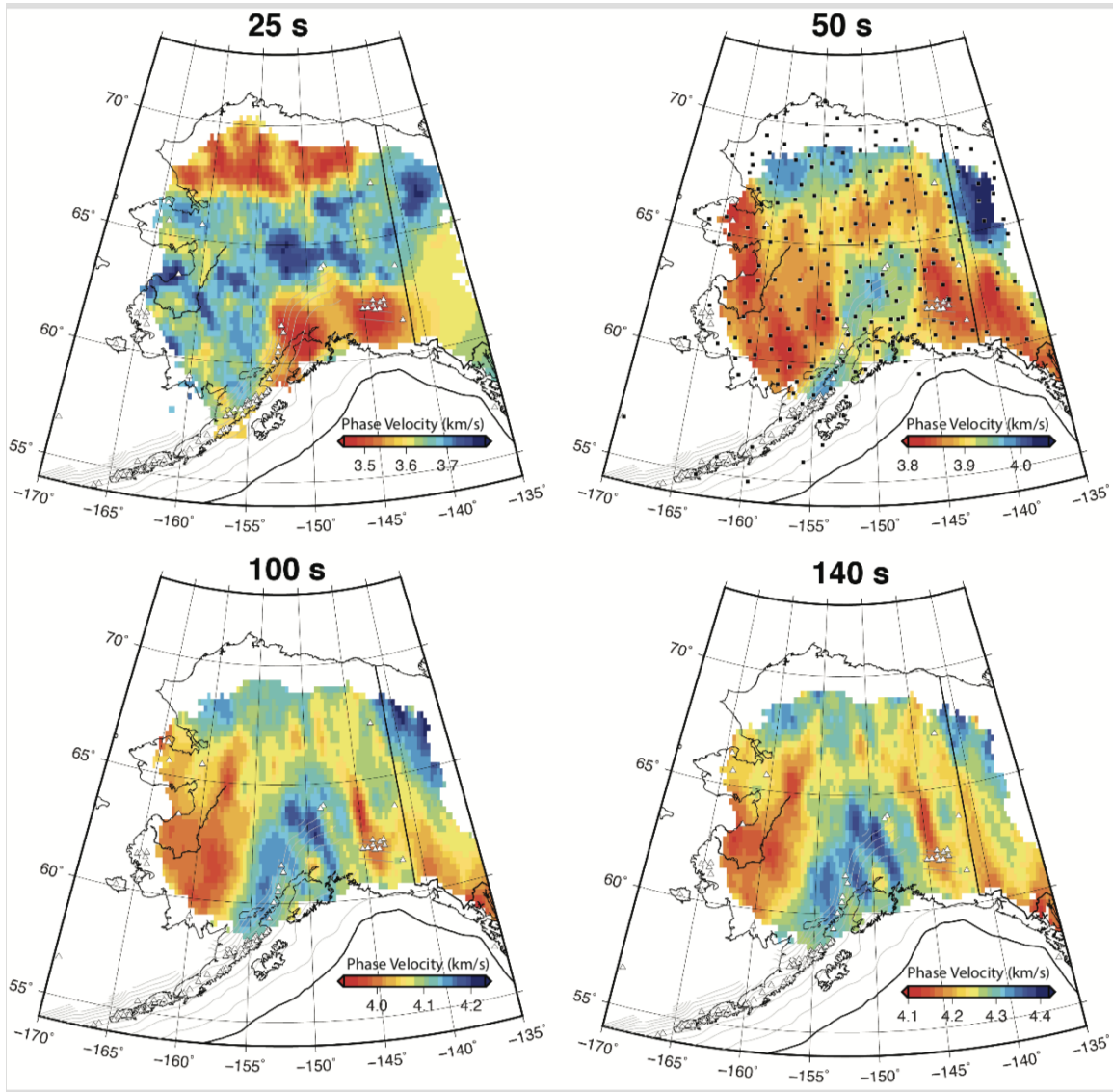


Figure 2.3: Rayleigh wave phase velocity maps with absolute velocities (km/s) for periods of 25 s, 50 s, 100 s, and 140 s. Pixels are shown only if more than 120 measurements contributed to the median phase velocity at that location for periods of 50 s, 100 s, and 140 s. Stations used in calculating phase velocities are shown by small black squares in the 50-s map. Seismicity contours and volcanoes are plotted as in Fig. 2.1.

Model Parameterization

We parameterized our models with adaptive discontinuous cubic B-spline functions, one set for the crust and one set for the mantle, to capture both smooth and sharp velocity variations. We solved for the isotropic velocity coefficients for each spline, the number and depth of spline knots, the depth and velocity contrast at the Moho, and hyperparameters that describe the relative uncertainties of the different data types. In the

mantle, both negative and positive shear velocity gradients were allowed, but in the crust only models with positive shear velocity gradients were accepted. This latter constraint does not significantly alter the shear velocity profiles in most cases, in part because the S_p receiver functions typically do not contain crustal phases that correspond to negative velocity gradients. Nodes were allowed to increase in number for depths up to 200 km to enable the resolution of sharper velocity gradients, but adjacent spline knots could not be closer than 3 km. A single value of V_p/V_s in the crust was a free parameter, but we fixed this value in the mantle at each depth following the AK135 reference model (Kennett et al. 1995). The ratio of density to V_s in the crust was defined after Brocher (2005) and in the mantle after Abers and Hacker (2016).

Forward Modeling

To calculate synthetic S_p phases for each model, we generated synthetic waveforms using the propagator matrix algorithm (Keith and Crampin, 1977). A dominant source period of 2 s and a ray parameter equal to the mean ray parameter of the stacked S_p phases were assumed. The synthetic data were processed in a manner that was identical to that applied to individual traces in the real data, resulting in filtered P and SV components. In practice, the final step of deconvolving the SV component from the P component was omitted for the synthetics to improved computational efficiency. This choice does not bias the inversions (Figs. A2 and A3). The timing and relative amplitudes of the phases in the non-deconvolved synthetics are identical to those after deconvolution. Thus, omitting deconvolution from the synthetics only eliminates what amounts to a filtering operation. The effects of including deconvolution in the observed data processing are discussed further in Section 2.4.

The synthetic P and SV components typically have greater short period content than the observed S_p RFs, but this is accounted for by the form of the body wave misfit function employed in the inversion. This misfit function, $\phi_{Sp}(m)$, is defined as:

$$\phi_{Sp}(m) = ||[P_p(t, m) * SV_{auto}(t)] - [SV_p(t, m) * SpRF(t)]||^2 \quad (2.1)$$

where P_p and SV_p are the predicted P and SV components, Sp_{RF} is the stack of observed Sp receiver functions, and SV_{auto} is the stack of the auto-deconvolved observed SV time series. Both terms in the misfit function contain cross-convolutions of observed and synthetic data, thus removing the need for observed and synthetic data to have the same period content. When calculating misfit, we used a -36 to -1 s range (note reversed time window with 0 s being the S arrival) that contained the Moho and LAB phases for Sp. We normalized the amplitudes of the Moho phases in the observed Sp receiver function and the predicted SV waveforms to 1. The contribution of each point in time to the misfit was weighted by empirically determined bootstrap errors for the observed RFs.

We calculated synthetic fundamental-mode Rayleigh wave phase velocities using the Mineos algorithm from Computational Infrastructure for Geodynamics (Masters et al., 2011). Because Mineos requires 1-D Earth models over the full radius of the Earth, the velocity models used for the forward calculation of Rayleigh wave phase velocities included the modeling being inverted for in the 0-300 km depth range, and the PREM model (Dziewonski and Anderson, 1981) from 400 km to the center of the Earth, with a gradient between the two models in the 300-400 km depth range. To optimize running time, we avoided directly calculating phase velocities from Mineos at each iteration. Instead, at each period in the data, we calculated perturbation kernels for V_p , V_s and density, all varying with depth. Perturbations in phase velocity were then determined by integrating the kernels with structural velocity and density perturbations over depth. Kernels were updated when changes in model parameters were sufficiently large. Eilon et al. (2018) describe thresholds and tests of this process in detail. The surface wave misfit function is defined as:

$$\phi_{sw} = ||C(f) - c_p(f, m)||^2 \quad (2.2)$$

where C and c_p are observed and predicted phase velocities for each frequency, respectively.

Inversion

We evaluated each candidate model using the posterior probability $p(m|d_{obs})$:

$$p(m|d_{obs}) \propto p(m) \times p(d_{obs}|m) \quad (2.3)$$

where $p(m)$ is the prior probability for a given model parameter and $p(d_i|m)$ is the likelihood function. The prior distributions for all parameters were determined empirically (Eilon et al., 2018), approximately uniform, and circumscribed by bounds given in Table A1. The likelihood function is defined as:

$$p(d_i|m) = \frac{1}{\sqrt{(2\pi\sigma_i)^n}} \times \exp\left(\frac{-\phi_i}{2\sigma_i^2}\right) \quad (2.4)$$

where ϕ_i is the misfit for each data type described in Section 2.2, and σ_i is the data error hyperparameter that controls the relative weights of the misfits for each data type. The degrees of freedom, n , for surface waves is the number of frequencies in the dispersion curves. For body waves, we estimated the degrees of freedom by using the first zero-crossing of the data auto-correlation function. We measured the total likelihood function for a given model as the product of likelihoods for both of the two data types.

$$p(d_{obs}|m) = p(d_{Sp}|m)p(d_{SW}|m) \quad (2.5)$$

We searched the parameter space using a Monte Carlo Markov Chain (MCMC) approach (e.g., Mosegaard and Tarantola, 1995) run on six independent parallel chains. Each chain was initiated with a model whose parameters were randomly selected from the prior distributions $p(m)$ (Table A1).

The models tested in subsequent iterations were obtained by perturbing a single model parameter by a random value drawn from a Gaussian distribution. Probabilities for perturbing a given parameter type appear in Table A1. Each chain contained twenty thousand iterations. The first eight thousand iterations represented a burn-in period with a rapid increase in likelihood and decrease in misfit. These iterations included a decaying "ther-

mal parameter” that multiplied the standard deviation of the Gaussian distribution from which parameter perturbations were drawn, increasing parameter variations early in the chain and avoiding confinement of the search within local minima (Eilon et al., 2018). The thermal parameter also multiplied the likelihood that a model would be accepted, enhancing model acceptance early in the chain. In each iteration, we accepted or rejected models using a Metropolis-Hastings acceptance criterion (Hastings, 1970). In the burn-in period, models were not saved to the final ensemble. After the burn-in period, likelihood values were high and steady, and we saved the accepted models every 30 iterations. We summed the posterior distribution from all accepted chains to form a final probability distribution. The final velocity model was defined as the median shear velocity at each depth, and crustal V_p/V_s and Moho depth were defined by their median values. Median values were chosen to reduce the influence of outlier accepted models. Uncertainties were represented by 95th and 68th percentile values.

2.2.4 Tests with Synthetic Data

In earlier applications of a similar inversion procedure (Eilon et al., 2018), stacks of P and SV waveforms (without deconvolution) were employed, in contrast to the Sp RFs used here. To evaluate the impact of adding deconvolution with the observed waveforms, we conducted inversions with synthetic data both with and without deconvolution. For an input velocity model that resembles some of the inversion results we obtained for stations in northern Alaska (e.g. the black line in Fig. 2.4) synthetic waveforms were calculated with the same propagator matrix method used in the forward calculation steps of the inversion (Keith and Crampin, 1977). Synthetic waveforms were converted to P and SV components, and then to Sp RFs and auto-deconvolved SV components, using the same procedures that were applied to the real data (Section 2.1). Before deconvolving we added white Gaussian noise with constant standard deviation of $\sigma = 0.02$ to the synthetics. Fundamental mode Rayleigh wave phase velocities were calculated for the input velocity model using Mineos (Masters et al., 2011) with Gaussian noise standard deviation of $\sigma = 0.01$. The synthetic data were inverted using the process described in

Section 2.3. The inversion with only synthetic surface wave phase velocities produces a V_s model probability distribution (colors in Fig. 2.4b) where the most likely models have absolute velocities that are similar to the input model, but the velocity increase at the Moho is not captured accurately, resulting in over-estimation of V_s in the shallow mantle lithosphere. In addition, the LAB velocity gradient that lies at depths of 130 km to 160 km in the input model is not captured, and instead V_s gradually decreases from approximately 75 km to 175 km.

In contrast, when S_p phase information is added to the inversions (Fig. 2.4c and 4d), the depths and velocity gradients at the Moho and LAB are well-resolved, highlighting the importance of the S_p data for constraining these boundaries. In the case where S_p RFs represent observed P phases, and the auto-deconvolved SV components represent observed SV phases, the resulting distribution of velocities at mantle depths close to the LAB gradient match the input model better (Fig. 2.4d) than in the case without deconvolution (Fig. 2.4c). This difference is likely due to the additional filtering introduced by the deconvolution operation. While the Gaussian noise in the non-deconvolved P and SV components creates greater possibilities for spurious phases to be fit by models that diverge from the input structure, this noise was largely removed in the deconvolution operation. Despite the differences in the frequency content of the deconvolved “data” and the non-deconvolved predicted waveforms used in the inversion in Fig. 2.4d, the cross-convolution misfit calculation enables good resolution of the input model. Diagnostics from the inversions with the deconvolved synthetic data and the other cases indicate the stability and accuracy of the inversion process (Figs. A2-A4). Inversions with S_p data without surface waves, which are shown in Eilon et al. (2018) but not in this paper, resolve velocity gradients at depths similar to those in the input model, but fail to accurately constrain absolute V_s .

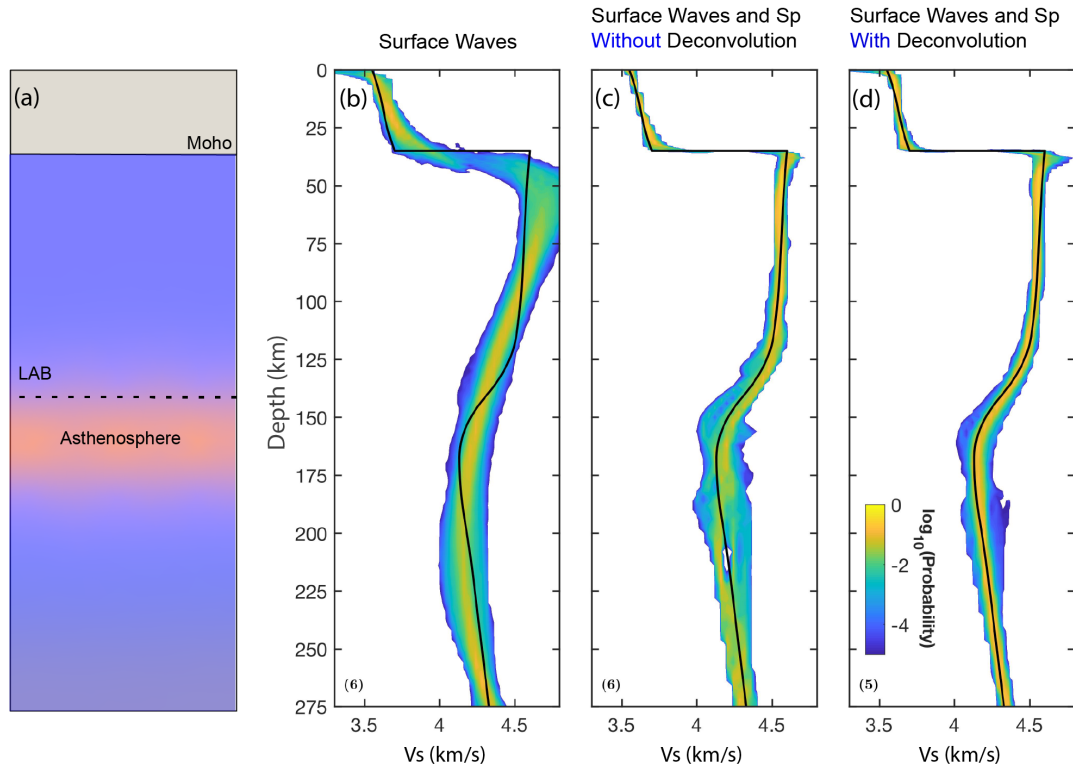


Figure 2.4: (a) Schematic of velocity structure used in synthetic test of inversion with a sharp change in velocity at Moho and a gradual decrease in velocity in the LAB. (b-d) The posterior probability of tests performed (colors) and the V_s model used to generate synthetic data (black). (b) Inversion of surface waves only; (c) surface waves and Sp without deconvolution; and (d) surface waves and Sp with deconvolution.

2.3 Results

2.3.1 Measuring Properties of the Lithosphere and Asthenosphere

In the shear-velocity models obtained by taking the mean of the models accepted in the Bayesian joint inversion, the Moho discontinuity is marked by a sharp increase in velocity. At mantle depths, we typically observe a layer with higher velocities, which we interpret as the mantle lithosphere, over a lower velocity layer which is consistent with the asthenosphere. In many cases, a second local maximum in velocity occurs at depths of 150 km to 250 km (Fig. 2.6).

We measured several properties that characterize the lithosphere and asthenosphere from the V_s profiles (Fig. 2.6): (i) the thickness of the lithosphere, (ii) the depth of

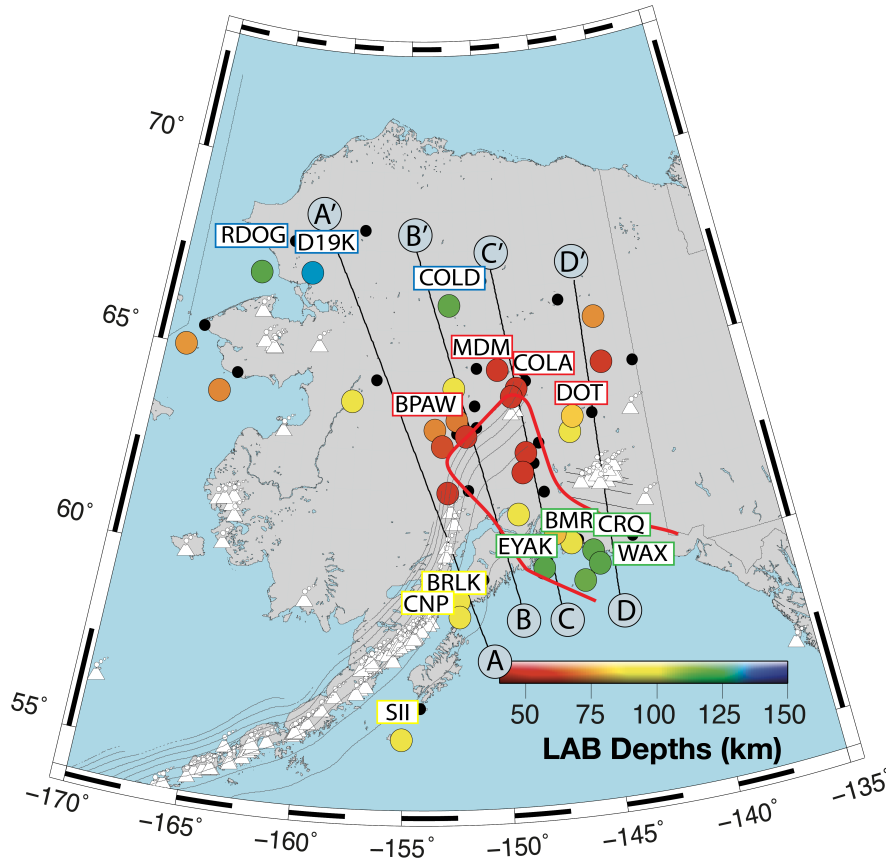


Figure 2.5: Map of LAB depths shown by colored points plotted at phase conversion points. Black dots are station locations. Seismicity contours and volcanoes are plotted as in Fig. 2.1. Yakutat block is outlined in red. Thick black lines show cross-sections in Fig. 2.8 and highlighted are the stations with V_s profiles plotted in Fig. 2.6.

the crust-mantle boundary or Moho, (iii) the average velocity in the lithosphere and (iv) the minimum velocity in the asthenosphere. To capture variations in the depth of the negative velocity gradient from lithosphere to asthenosphere, we defined the thickness of the lithosphere (i.e. LAB depth) as the mid-point between the maximum velocity in the mantle lithosphere (measured at depths starting 10 km below the Moho) to the next velocity minimum, as marked in Fig. 2.6 by the grey dashed line and red squares, respectively. LAB depth uncertainties correspond to 25% of this measurement depth range. LAB depths measured from the joint inversion shear velocity profiles in many cases differ from the depths of LAB phase peaks on the Sp RFs shown in Fig. 2.2 and Fig. A1, although their uncertainty ranges overlap. These discrepancies are expected, because the

Sp RFs in these figures were migrated with the SEMum2 model (French and Romanowicz, 2013) which differs from the results of the joint inversion, and because predicted Sp phase depths are influenced by the shape of a velocity gradient and will not necessarily lie exactly at its mid-point depth. Moho depths are directly calculated in the inversion as described in Section 2.3.

We defined the minimum velocity in the asthenosphere as the average within 5 km of the velocity minimum. Moho depths were determined as a parameter in the inversion as explained in Section 2.3.1, with 2σ error bounds, and are characterized by a large positive velocity jump in the Vs profiles, as seen in Figs. 2.4 and 2.6. We calculated the average velocity in the lithosphere by taking the mean of the velocities in the profile between the Moho and the determined LAB depths. Moho, lithosphere, and asthenosphere parameter values for each station appear in Table A2.

2.3.2 Regional Patterns in Shear Velocity

Our shear-wave velocity profiles and the parameters measured from them show distinct regional patterns. In this section, we describe observed patterns in all parameters and compare Moho and LAB depths to previous converted wave and active source studies. Further discussion of the parameter patterns and their origins appears in Section 2.4.

Moho depths vary significantly across the region (Fig. 2.7a). In central Alaska, at the northern edge of the subducting slab, the crust is relatively thin (<30 km) with the exception of one station, and its thickness increases modestly to the north and east, for example reaching values of 30 to 35 km beneath the Brooks Range. This northward thickening is consistent with active source studies (e.g. Fuis et al., 2008). Apparent Moho depths are 35 to 40 km where the Yakutat block is thought to have subducted (red outline, Figure 2.7a; Fuis et al., 2008), with the exception of the two southeastern stations (AK.WAX, AK.SUCK) which have very shallow Moho depths. Beneath the southern Kenai peninsula, stations AK.BRLK and AK.CNP indicate relatively thick crust (35 km to 42 km), although they lie outside the limits of thick Yakutat crust as constrained by offshore active source data (Fuis et al., 2008). Finally, the crust is thin (~ 25 km) beneath the

Alaska Peninsula in the southwest where the Pacific plate is subducting.

The overall pattern of crustal thickness obtained in the inversions is similar to the large-scale variations in crustal thickness obtained with Ps phases in Zhang et al. (2019) and Miller et al. (2018). For 18 of the 26 stations with crustal thickness values in this study and in Zhang et al. (2019), or 20 of the 28 stations in this study and Miller et al. (2018), our Moho depths are within 5 km of the values in the other studies, taking into account our uncertainties. At the other stations, the crustal thicknesses in this study are smaller, with one exception. The average absolute difference between our Moho depths (including their uncertainties) and those from Zhang et al. (2019) is $3.5 \text{ km} \pm 3.3 \text{ km}$ (one standard deviation), and between our results and Miller et al. (2018) the average absolute difference is $4.2 \text{ km} \pm 5.1 \text{ km}$. These Sp versus Ps discrepancies are similar to the average intra-Ps difference between Zhang et al. (2019) and Miller et al. (2018) Moho depths.

LAB depths vary coherently over broad regions (Fig. 2.5). For instance, the northernmost group of stations (TA.D19K, AK.COLD, and AK.RDOG), which lie in the margins of the Brooks Range (Fig. 2.1a), in and near the northern Arctic Alaska terrane (Fig. 2.1b), manifest the thickest lithosphere (118 km to 132 km), while stations in central Alaska have the thinnest lithosphere (45 km to 91 km), including stations on the west coast of the Seward Peninsula whose conversion points sample the offshore continental plate (AK.TNA and AK.ANM). In southern Alaska, where LAB velocity gradients mark the base of the subducting lithosphere, the LAB depths for the Yakutat plate (whose outline is shown in Fig. 2.1b) range from 79 km to 122 km. The values at the upper end of this range, which correspond to the Sp conversion points that sample farthest offshore, are significantly larger than those observed at the base of the subducting Pacific plate to the west (91 km to 98 km). Both the thickening of lithosphere from central Alaska to northern Alaska, and the thicker lithosphere associated with the subducting Yakutat terrane relative to the Pacific plate, were observed in the Sp receiver function stacking of O'Driscoll and Miller (2015). Minimum asthenospheric velocities follow broad regional patterns (Fig. 2.7c). The lowest asthenospheric velocities ($< 4.3 \text{ km/s}$) are found across central Alaska where the lithosphere is thinnest, including the Seward Peninsula. The av-

erage of the minimum asthenospheric velocities for this region ($4.30 \text{ km/s} \pm 0.07 \text{ km/s}$) is lower than the average for the remaining regions of Alaska ($4.42 \text{ km/s} \pm 0.07 \text{ km/s}$). The highest asthenospheric velocities occur beneath the Brooks Range in the northern portion of the Arctic Alaska terrane (TA.D19K) and beneath the lithosphere subducting under the Kenai Peninsula (AK.BRLK and AK.CNP) (Fig. 2.6d and Fig. 2.7c).

Values for average velocity in the mantle lithosphere in some cases vary significantly at short spatial scales, but two aspects of their distribution stand out. We observe the lowest mantle lithosphere velocities at the eastern edge of the subducting lithosphere ($\sim 4.25 \text{ km/s}$). In addition, the three northern stations with the thickest lithosphere have relatively high average lithospheric velocities ($> 4.55 \text{ km/s}$).

A final observation from the Vs profiles is that some of them contain local maxima at $\sim 175 \text{ km}$. These features are particularly pronounced under the slab offshore (Fig. 2.6c-d) but are also seen further north (Fig. 2.6a).

2.4 Discussion

The first-order features of the shear velocity models obtained by the Bayesian joint inversions are that the lithosphere in northern Alaska near the Brooks Range is significantly thicker than in central Alaska, and that asthenospheric velocities are low in a broad band beneath the thin lithosphere of central Alaska, including the Seward Peninsula. Similar patterns are echoed in numerous studies (e.g., Wang and Tape, 2014; O'Driscoll and Miller, 2015; Martin-Short et al. 2018; Jiang et al., 2018; Ward and Lin, 2018; Feng and Ritzwoller, 2019; Berg et al., 2020). In addition, the shear velocity models indicate a greater maximum thickness for the subducting Yakutat plate than for the subducting Pacific plate, as also observed in O'Driscoll and Miller (2015) and Martin-Short et al. (2018). However, while some of the major features of the shear-velocity models obtained in this study are consistent with prior work, our use of Sp phases provides narrower constraints on LAB depths than would be possible with surface waves alone, and the estimates of absolute shear velocities in the lithosphere and asthenosphere that honor the LAB depths would not be possible with converted waves alone. In this section,

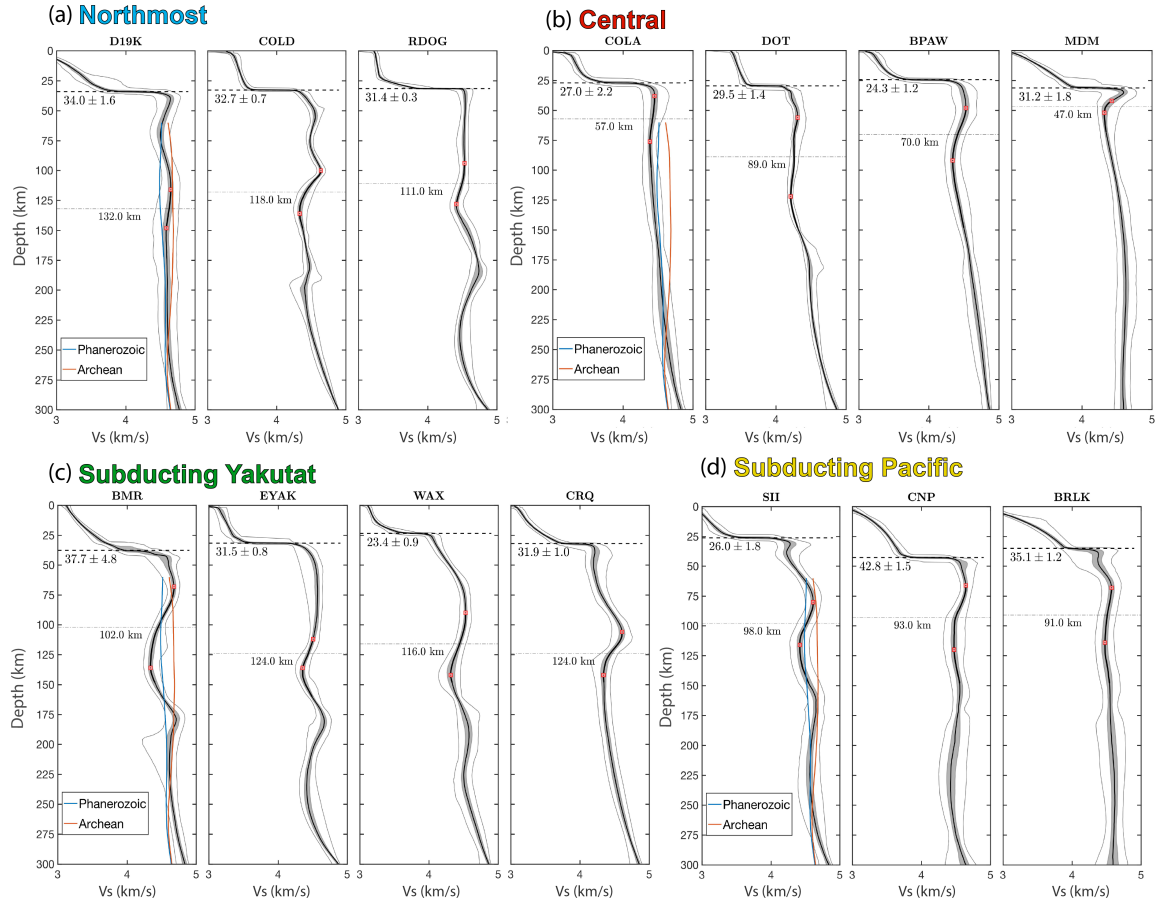


Figure 2.6: Isotropic shear wave velocity profiles. Black lines are the average models and grey lines are 2σ bounds. (a-d) group of figures sample different Alaskan geographic regions. Overlaid lines are V_s from global model SEMum2 averaged by the age of the continental crust, Phanerozoic (blue) and Archean (red) on the first profiles of each region. Red squares mark the maximum velocity in the mantle lithosphere (starting 10 km below the Moho) and the minimum velocity in the asthenosphere; grey dashed line is LAB depth defined as the mid-point of this depth range. Black dashed line is Moho depth.

we discuss the implications of these results for upper plate modification by subduction-related processes in the mantle wedge, and the characteristics of the subducting Pacific and Yakutat plates.

2.4.1 Interactions between Subduction-related Processes and the Upper Plate

The broad band of thin lithosphere and low velocity asthenosphere that stretches across central Alaska is easily interpreted as a signature of subduction. Low asthenospheric

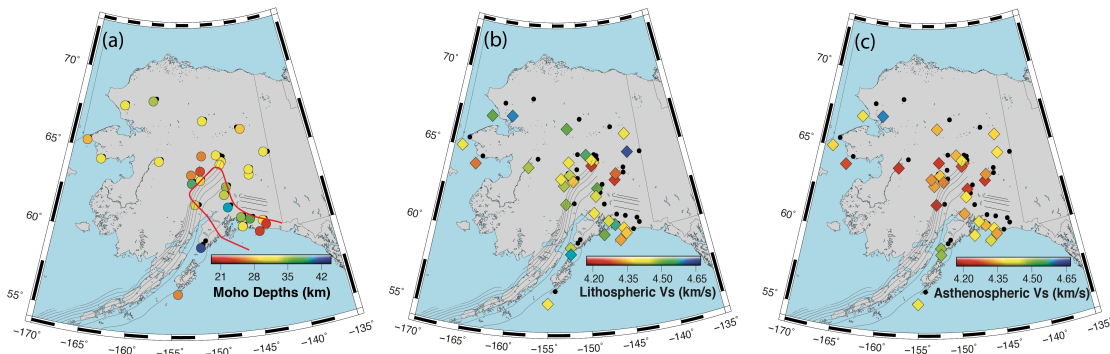


Figure 2.7: (a) Map of Moho conversion points from the shear velocity profiles. Yakutat block is outlined in red. Maps of velocities in the (b) lithosphere and (c) asthenosphere. Black dots are the station locations. The velocities are plotted at the lithosphere-asthenosphere boundary conversion points to facilitate the comparison with the calculated LAB depths. Therefore, the average lithospheric velocities correspond to a depth range for which conversion points would plot closer to the station and the minimum asthenospheric velocities are from a depth range for which conversion points are farther from the station. Seismicity contours are plotted as in Fig. 2.1.

velocities likely reflect high temperature mantle entrained into the mantle wedge by the subducting slab, slab fluid flux into the mantle wedge (although the competing roles of water and grain size on mantle wedge velocities are debated, e.g. Abers et al., 2014), and possibly the presence of partial melt (e.g. Rondenay et al., 2010; Rodriguez-Gonzalez et al., 2012). Rising slab fluids and/or partial melt, potentially in combination with shallowing asthenospheric flow lines, in turn provide a mechanism to thin the upper plate lithosphere. This interpretation is supported by the correlation of thin upper plate lithosphere and surface magmatism, as seen in cross-sections A and C in Fig. 2.8 and beneath the Seward Peninsula (Fig. 2.1a). The zone of thin lithosphere and low asthenospheric velocities lies landward of the low-attenuation, and presumably cold, wedge corner imaged by Stachnik et al. (2004). The thin lithosphere correlates with a region of particularly high horizontal surface strain rate (Finzel et al., 2011), and is consistent with the idea that the thinner lithosphere is weaker and more easily deformable. Comparison of apparent LAB depths with slab seismicity indicates that the velocity reductions at two stations (AK.DHY and AK.WAT6, Fig. 2.8c) likely represent a slab-related interface that corresponds to a decrease in velocity with depth.

The lithosphere of northernmost Alaska poses a stark contrast. Beneath the Brooks

Range and the northern Arctic-Alaska terrane (Fig. 2.1) the lithosphere not only reaches thicknesses of ~ 130 km (Fig. 2.8a-b), it also contains shear velocities (stations TA.D19K and AK.COLD, Fig. 2.6a) that are comparable to the global average for Archean cratons from the SEMum2 model (French and Romanowicz, 2013). These seismic properties match other geophysical data in the northern Arctic-Alaska terrane that are typical of cratonic lithosphere, including low heat flow (except near the northern coast (Saltus and Hudson, 2007; Batir et al., 2016)) and a zone of high magnetization in the deep crust that has been interpreted as evidence for cold lithosphere (Saltus and Hudson, 2007). These properties have led to the conclusion that the thick northern Alaska lithosphere represents high viscosity mantle (e.g. Finzel et al., 2015). The apparent contrasts in lithospheric thickness and viscosity between northern and central Alaska are in turn important for focusing tractions from asthenospheric flow and predicting surface deformation that matches southward directed surface deformation (Finzel et al., 2015).

Within the subduction modified upper plate lithosphere of central Alaska, the origin of the “Denali Volcanic Gap” has been a topic of debate. The Denali Volcanic Gap (DVG) is an area that lies above subducting slab seismicity, yet has lacked significant volcanism since the Miocene (Fig. 2.1a) and is spatially correlated with the outline of the subducting Yakutat terrane (Fig. 2.1b). To explore implications for the DVG, we compared shear velocity model properties for a group of stations whose Sp stack mean conversion points lie within the DVG (AK.CAST, AK.BPAW, AK.CHUM, AK.KTH, AK.MLY) to stations whose conversion points lie outside the DVG but at a similar location with respect to slab seismicity contours (AK.MDM, IU.COLA, AK.CCB, AK.SKN). Minimum asthenospheric velocities and average lithospheric velocities do not differ significantly between the two groups: comparing average values within the DVG to those outside the DVG, asthenospheric velocities are 4.30 ± 0.06 km/s versus 4.30 ± 0.09 km/s, respectively, and lithospheric velocities are 4.43 ± 0.05 km/s versus 4.45 ± 0.09 km/s, respectively.

In contrast, variations in LAB depth do show DVG-related contrasts. Inside the DVG, the LAB lies at depths of 50-60 km directly above slab seismicity located at 140 km depth, in agreement with a sharp reduction in shear velocity imaged and interpreted as pooled partial melt by Rondenay et al. (2010) (Fig. 2.8b). However, 70 km north from

the deepest seismicity along the cross-section, the LAB deepens to approximately 70 km in depth (although this is not technically resolved given the generous error bars on LAB depth) and by 150 km from the deepest seismicity, the LAB has reached a depth of 90 km (AK.MLY). In contrast, on the cross-section on the eastern side of the DVG (Fig. 2.8c) the LAB appears to remain at a depth of approximately 50 km to a distance of 70 km from the deepest seismicity (AK.MDM). This difference in LAB topography can also be seen on the LAB depth map in Fig. 2.5.

In some models for the DVG, low angle subduction of the Yakutat terrane beneath the DVG cools the top of the mantle wedge (Rondenay et al., 2010; Martin-Short et al., 2018). For example, Rondenay et al. (2010) propose that this cooling reduces upward flow of warm asthenosphere where the arc would otherwise be located, and instead causes melt to pond below the upper plate at a depth of approximately 60 km somewhat farther landward. Our results are consistent with this model, in the sense that we do observe a shallow LAB where Rondenay et al. (2010) predict ponded melt. One caveat is that volcanoes are present to the northeast of the DVG (Fig. 2.8c) where the angle of subduction is comparably low to that in the DVG. However, these volcanoes may be explained by their proximity to the edge of the slab, assuming the interpretation of seismicity we employ in this study (e.g. Kiara Daly, personal communication, 2020). Alternatively, upper plate thickness, for example the significantly thicker lithosphere in northern Alaska, may contribute to the shallow Yakutat slab dip, as opposed to attributing the shallow dip to the buoyancy of thick Yakutat crust alone. For example, in the geodynamical models of Rodriguez-Gonzalez et al. (2012) thicker upper plate lithosphere alters wedge dynamics and reduces slab dip. Following this logic, the relatively thin lithosphere beneath the Seward Peninsula would be associated with the more steeply dipping Pacific plate.

In a different type of model for the DVG, Chuang et al. (2017) propose that low angle subduction produces slab dehydration at anomalously shallow depths, reducing the presence of fluids and the production of partial melt in the mantle wedge. The observed lack of variation in asthenospheric velocities between the DVG and surrounding regions is not consistent with this model, unless the proposed variation in fluid and melt content produces no discernible effect on asthenospheric shear velocity.

Our results also shed partial light on the degree to which the upper plate in Alaska is modified by subduction zone processes versus intrinsic lithospheric variations imparted by the original accreted terranes. Our results suggest that both are important. On the one hand, the thick lithosphere in northern Alaska (Fig. 2.8, Fig. A.5), largely made up of the NeoProterozoic-Cambrian northern Arctic-Alaska terrane, has high shear velocities comparable to Archean cratons globally (as discussed above, e.g. Fig. 2.7b). This cratonic character suggests that the bulk mantle lithospheric properties of the northern Arctic-Alaska terrane differed from those of the terranes to its south at the time that they were accreted, including the nearly adjacent Koyukuk terrane which has been interpreted as oceanic arc lithosphere accreted in the Late Jurassic to early Cretaceous (Colpron et al., 2007; Colpron and Nelson, 2011). On the other hand, subduction processes clearly influence the upper plate at latitudes that at least border the thick northern Arctic Alaska lithosphere. The low asthenospheric shear velocities, thin lithosphere and magmatism beneath the Seward Peninsula are evidence suggestive of a subduction influence substantially to the north of the present-day convergent boundaries. In addition, studies of mantle azimuthal anisotropy and regional flow models indicate that mantle flow driven by subduction extends to the latitudes of the northern Arctic Alaska terrane (e.g. Jadamec et al., 2013; Venereau et al., 2019; McPherson et al., 2020).

2.4.2 Subducting Plate Properties

Another feature of the shear velocity models is the variation between the thicker subducting lithosphere that lies within the outline of the subducting Yakutat terrane (Fig. 2.8c-d) and thinner subducting lithosphere with Pacific oceanic crust (Fig. 2.5). Conversion points sampling the furthest offshore subducting plate with Yakutat crust (Fig. 2.8c-d) indicate lithosphere that is ~ 120 km thick, with maximum values of 122 ± 12 km. In contrast, the conversion point for station SII (Fig. 2.5), which lies beneath subducting lithosphere with Pacific plate crust at a comparable offshore position relative to subducting slab contours, occurs at 98 ± 16 km. Although only one station in our study captures this offshore distance on the Pacific plate, the LAB depth at this station is comparable to

values for Pacific lithosphere near this location from Kumar and Kawakatsu (2011). An apparent difference in incoming plate thickness has also been observed by O'Driscoll and Miller (2015) and Martin-Short et al. (2018).

Variations in crustal thickness explain at least some of the greater total observed thickness of the Yakutat lithosphere. Based on its 20 My age (Müller et al., 2008), the Yakutat lithosphere would actually have a thermally-defined plate thickness that is ~ 10 km less than the 40 My-old Pacific lithosphere, assuming a cooling plate model with ~ 90 km thick infinite-age lithosphere that matches heat flow data (Hasterok, 2013) and that the base of oceanic lithosphere lies within 50°C of asthenospheric temperature. However, the thickness of the offshore Yakutat crust is approximately 30 km (Christeson et al., 2013), as opposed to the ~ 9 km thick crust of the Pacific plate (Fuis et al., 2008). Thus the ~ 21 km thicker crust of the Yakutat plate offsets greater its ~ 10 km thinner thermal mantle, predicting a total thickness that is greater than that of the Pacific plate.

Another facet of subducting plate thickness is that on cross-section C (Fig. 2.8), the subducting plate carrying Yakutat crust, while ~ 120 km offshore, appears to thin as it bends into the trench. Pooling of partial melt below a solidus-related interface near the base of the lithosphere just as it subducts has been invoked beneath Nicaragua from seafloor magnetotelluric data (Naif et al., 2013), and beneath Cascadia in the U.S. from velocity tomography (Hawley et al., 2016). The presence of accumulated melt is one candidate process that could produce the apparent erosion of the base of the Yakutat lithosphere.

As previously mentioned, the negative velocity gradients at depths of 66 km and 58 km beneath stations AK.DHY and AK.WAT6 appear to be related to the top of the subducting plate, based on their location relative to slab seismicity (Fig. 2.8c). Using Ps receiver function migration, Kim et al. (2014) found a thin layer with reduced velocity that lies close to the AK.DHY and AK.WAT6 conversion points. They interpreted this feature as layer related to the plate interface that represents thick sediments the trench in addition to high pore fluid pressure between the subducted slab and upper plate (Kim et al., 2014).

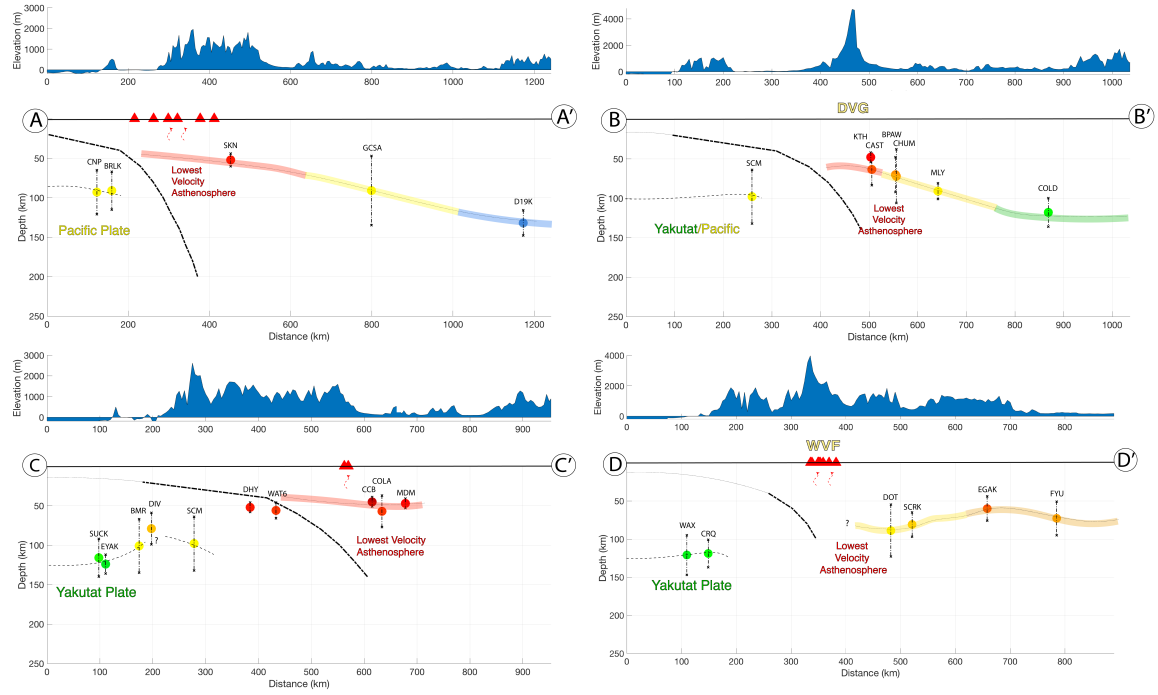


Figure 2.8: Cross-sections that represent shear velocity structure beneath Alaska. Cross-section locations are shown in map view in Fig. 2.5. Colored circles are LAB depths shown at conversion point locations, except for conversions at stations AK.DHY and AK.WAT6 in cross-section C-C' which likely represent the top of the crust of the subducting plate. Light dashed lines (at the base of the subducting plate) and colored lines (at the base of the upper plate) are our interpretations of LAB shape. Thick dashed lines are the top of the subducting plate from Slab2.0 (Hayes, et al. 2018) and Wrangell field contours from Kiara Daly, personal communication (2020) for profile D-D'. Additional discussion of the WVF appears in the supplement. The Denali Volcanic Gap (DVG) location is shown in B-B'. On top of the cross-sections, we show topography from ETOPO1 (Amante and Eakins, 2009). Red triangles represent volcanoes. Both volcanoes and LAB conversion points lie within 0.8 geocentric degrees from the cross-sections.

2.5 Conclusions

We employed Bayesian joint inversion of Sp RFs and Rayleigh wave phase velocities to produce shear velocity profiles across Alaska. These new constraints on lithospheric thickness, and lithospheric and asthenospheric velocities shed light on several questions.

How has subduction modified upper plate structure? What are the mantle signatures of accreted terranes? A wide band of thin upper plate lithosphere and low velocity asthenosphere extends across central Alaska and the Seward Peninsula. Its spatial correlation with subducting plate seismicity and upper plate magmatic activity indicates that subduction-related fluids and melt have played a key role in thinning the upper plate. In

contrast, not only is the lithosphere in and near the northern Arctic Alaska terrane significantly thicker, its mantle has shear velocities that reach values typical of Archean cratons. The latter property suggests that while subduction-related flow may reach these latitudes, possibly thinning the lithosphere, subduction-related processes have not obviously altered the mantle lithosphere velocity structure of the Alaskan interior.

What are the origins of the Denali Volcanic Gap? LAB depths are consistent with Denali Volcanic Gap models involving feedbacks between subduction zone processes and the upper plate. Asthenospheric velocities above the subducting plate within the Denali Volcanic Gap are not significantly different from those beneath adjacent magmatically active regions, suggesting that lower fluxes of slab fluids and reduced wedge partial melt content are not the dominant factor in creating the Denali Volcanic Gap, unless these effects have unobservable impacts on shear wave velocities.

What differences exist between the subducting plate carrying the Yakutat terrane and the normal Pacific plate lithosphere to the west? The total thickness of the Yakutat lithosphere offshore is about 20 km greater than the Pacific lithosphere, a result which can be explained by the greater thickness of Yakutat crust (Christeson et al., 2013). Thinning of Yakutat lithosphere as it bends into the trench is consistent with erosion of the base of the lithosphere by accumulated partial melt, as has been proposed in Nicaragua and Cascadia (e.g., Naif et al., 2013; Hawley et al., 2016).

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2.7 References

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Chapter 3

Seismic Velocity Discontinuities beneath Alaska from Sp Phases

This chapter is ready to be submitted as: *Isabella Gama, Karen M. Fischer, Junlin Hua. Mapping the Lithosphere and Asthenosphere beneath Alaska with Sp Converted Waves, Geochemistry, Geophysics, Geosystems.*

Summary

- We obtained a 3D image of crust and mantle seismic velocity gradients beneath the state of Alaska using common-conversion point (CCP) stacking of Sp converted body waves recorded by hundreds of stations from the NSF EarthScope Transportable Array and other portable arrays and permanent networks.
- Moho depths delineate the thick crust of the underthrust Yakutat terrane and the crustal root beneath the Brooks Range. The North American lithosphere is particularly thin close to the subducting lithosphere in the main Alaska subduction zone, consistent with thinning of the upper plate by subduction zone flow and melt rising from the mantle wedge.
- The lithosphere remains relatively thin far to the northwest and north, including the Seward Peninsula, as well as regions just to the south of the Brooks Range where

lithospheric foundering may have occurred. The lithosphere dramatically thickens beneath the Brooks Range and northern Arctic Alaska terrane where it appears to be both cold and highly viscous.

- The CCP stack also revealed a pronounced positive velocity gradient at depths of 130-230 km that represents the base of a layer whose low velocities are best explained by the presence of partial melt. Although this gradient is present close to the subducting lithosphere, where partial melting is enabled by slab-derived fluids, it is strongest beneath the Seward Peninsula and to the northeast of the Wrangell Volcanic Field (WVF), suggesting the presence of partial melt in the asthenosphere hundreds of kilometers away from the slab, likely due to decompression melting in upwelling asthenosphere.

3.1 Introduction

The global distribution of partial melt in the asthenosphere is still widely debated. In subduction zones, seismic velocity and attenuation models have been used to trace the generation and transport of partial melt in the mantle wedge from just above the subducting slab to the arc (e.g., Tamura, 2003; Syracuse et al., 2008; Abers et al., 2014), including in Alaska (McNamara and Pasyanos, 2002; Eberhart-Phillips et al., 2006; Soto Casteneda et al., 2021), and beneath back-arc spreading centers (e.g., Wei and Wiens, 2018; Wei and Wiens, 2020). However, the distribution of partial melt on a more widespread basis in subduction zone mantle wedges, and its extent beneath the adjacent lithosphere, remains uncertain.

Mantle seismic velocity discontinuities - or vertically localized seismic velocity gradients - have the potential to help resolve the distribution of melt in the mantle. A negative velocity gradient (decreasing velocity with depth) at the lithosphere-asthenosphere boundary (or LAB) has been explained by various models. While temperature clearly plays a role in creating high seismic velocities in the mantle lithosphere (Hirth and Kohlstedt, 2003; Stixrude and Lithgow-Bertelloni, 2005), additional factors have been invoked to

produce negative velocity gradients that are vertically localized, including the presence of volatiles and partial melt in the asthenosphere, elastically accommodated grain boundary sliding, near-solidus weakening, and seismic anisotropy (e.g., Yamauchi and Takei, 2020; Hansen et al., 2016; also see reviews in Rychert et al., 2018; Karato and Park, 2019; Fischer et al., 2020). In some regions, the seismic LAB has been associated with solidii, for example, a relatively dry peridotite solidus in the western U.S. (Plank and Forsyth, 2016; Golos and Fischer, 2022), a hydrated peridotite solidus in subduction zones (e.g., Wang et al., 2020), or a carbonated peridotite solidus beneath the oceans and in subduction zones (e.g., Hirschmann, 2010; Dasgupta, 2018; Hammouda et al., 2020). In Alaska, Rondenay et al. (2010) imaged a negative velocity gradient at depths of ~ 60 km, which they attributed to partial melt pooled beneath the base of the North American lithosphere. Mantle discontinuities may also be created at the lower boundaries of the layers where asthenospheric temperatures exceed the temperature of a particular solidus (e.g., Hirschmann, 2010; Dasgupta, 2018; Wang et al., 2020; Hammouda et al., 2020), creating a positive velocity gradient with depth, as has been observed beneath Anatolia (Hua et al., 2020b). A small amount of partial melt has also been invoked to explain zones of high electrical conductivity in some regions of the asthenosphere (e.g., Naif et al., 2013; Chesley et al., 2019; Selway et al., 2019).

Converted body waves (Sp and Ps) and reflected phases effectively image vertically-localized velocity gradients and provide an essential complement to body and surface wave tomography, which reflects velocities or velocity anomalies throughout a given volume. For example, the addition of Sp (e.g., Gama et al., 2021), Ps (e.g., Julia et al., 2000; Shen et al., 2013; Schmandt et al., 2015; Shen and Ritzwoller, 2016; Shen et al., 2018; Calo et al., 2016; Bodin et al., 2013; Porritt et al., 2015; Roy and Romanowicz, 2017; Ward et al., 2017; Martin-Short et al., 2018; Berg et al., 2020) or Sp and Ps phases (Eilon et al., 2018) to inversions of surface-wave phase velocities enhances vertical resolution of both mantle and Moho velocity gradients. Sp phases have an advantage over Ps. Sp mantle arrivals are not over-printed by Moho reverberations, and Moho phases are not over-printed by reverberations from shallow crustal layers. Due to their Fresnel zones (e.g., Hansen and Schmandt, 2017; Mancinelli and Fischer, 2017), Sp phases are typically

not able to accurately capture velocity gradients with dips of more than 10° - 15° (e.g., Hansen and Schmandt, 2017; Lekić and Fischer, 2017; Hua et al., 2020a). However, Sp Fresnel zones are especially well-suited to imaging near-horizontal velocity gradients and stacking approaches that rely on horizontal averaging, such as common conversion point (CCP) stacking (e.g., Hua et al., 2020a; Hua et al., 2020b).

In this paper, we employed CCP stacking of Sp converted phases to study lithospheric and asthenospheric properties near the subduction zone and across the continental lithosphere beneath the state of Alaska. Prior studies have imaged mantle discontinuities beneath Alaska using Sp (Bauer et al., 2014; O'Driscoll and Miller, 2015) and Ps (Rondenay et al., 2008; Rondenay et al., 2010; Bauer et al., 2014; Mann et al., 2022) phases; explored upper and subducting plate crustal structure with Ps data (Ferris et al., 2003; Veenstra et al., 2006; Rondenay et al., 2008; Rondenay et al., 2010; Brennan et al., 2011; Kim et al., 2014; Allam et al., 2017; Miller et al., 2018; Miller and Moresi, 2018; Zhang et al., 2019); and incorporated Sp (Gama et al., 2021) or Ps (Ward et al., 2017; Martin-Short et al., 2018; Berg et al., 2020; Feng, 2021) data in joint inversions with surface waves. However, to our knowledge, this is the first study to comprehensively map crust and mantle discontinuities in Alaska with Sp phases recorded by the NSF EarthScope Transportable Array. In Alaska, the contrast between the active convergent margin in the south and the more stable continental region in the north (e.g., Finzel et al., 2015) provides insight into how asthenospheric signatures vary as a function of tectonic setting. We focused on three main boundaries: the crust-mantle boundary (or Moho), which is critical for understanding the interaction of Alaskan tectonic terranes, the LAB, and the base of a low-velocity asthenosphere layer which appears as a positive velocity gradient (PVG). We argue for the presence of partial melt where the LAB and PVG are strong in some areas of Alaska and explore different mechanisms for melt generation and their variation beneath Alaska.

3.2 Data and Method

To image shear-wave velocity gradients, we applied CCP stacking to Sp phases from teleseismic earthquakes recorded by permanent and temporary broadband stations in Alaska, in particular the stations of the NSF EarthScope Transportable array (Fig. 3.1A). These data create a dense distribution of ray paths beneath Alaska; 20,210 Sp receiver functions were used to construct the 3D CCP volume. Each step of the data analysis process is described in the following sections.

3.2.1 Data

We selected earthquakes between the years 1986 and 2021 and with Mw larger than 5.8. Epicentral distances of 50° to 90° were used for Sp receiver functions and CCP stacking, and these paths are dominated by back-azimuthal groups, 215° - 255° (southeast Pacific events) and 255° - 285° (north and northwest Pacific events) (Fig. 3.1B). P waves used in calculating free-surface velocities were collected from events at 35° to 90° .

We employed data from 449 seismic stations in Alaska (Fig. 3.1A). Of these, 153 are EarthScope Transportable Array stations, and the remaining are other temporary and permanent stations. References for the relevant network codes (AK, AT, AV, DW, II, IM, IU, PN, PP, TA, US, XE, XR, XZ, YE, YG, YM, YV, Z5) appear in the Data Availability Statement.

3.2.2 Calculating Receiver Functions

To enable robust phase identification, we calculated the arrival time of the P and S phases using an array-based method (Lekić and Fischer, 2014). We filtered the seismograms using a 2-100 s band-pass filter. We used a free-surface transform to better separate waveforms into P and SV components (Kennett, 1991). We obtained the free-surface velocities needed for the free-surface transform using the particle motion modeling approach from Hua et al. (2020b) applied to direct P and S arrivals.

We deconvolved the SV component of the direct S arrival from the P component

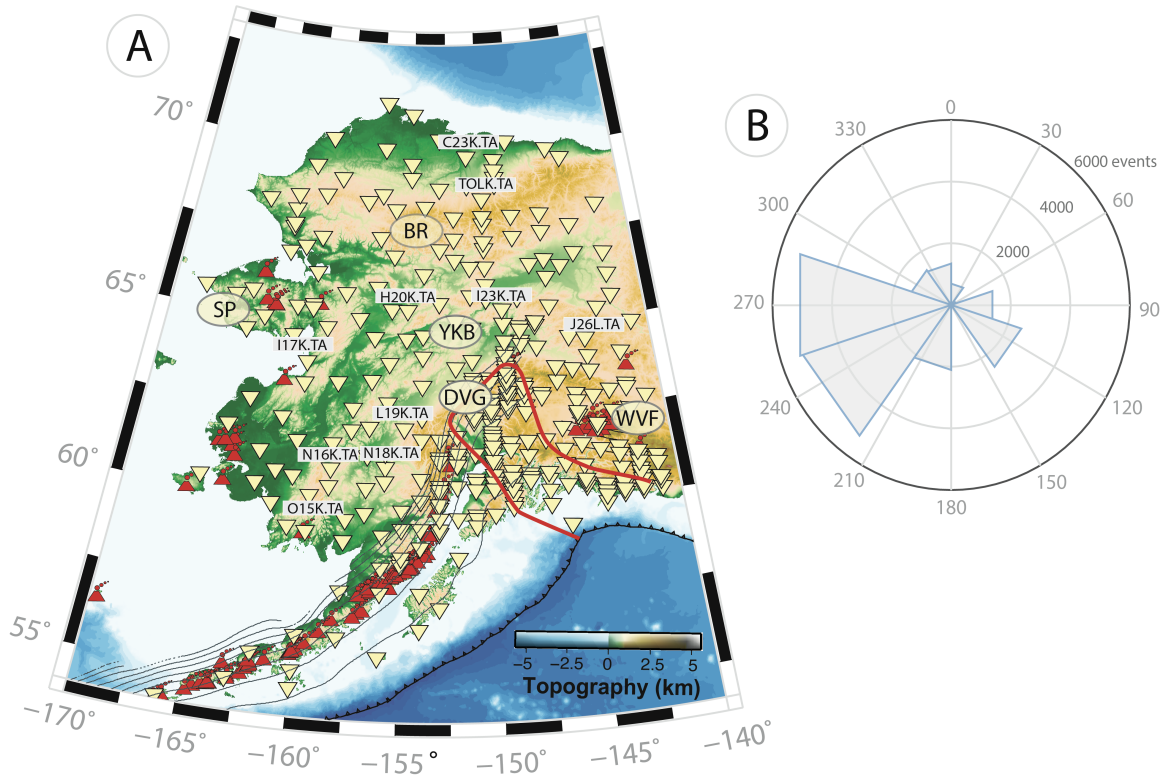


Figure 3.1: Topographic map of Alaska (ETOPO1, Amante and Eakins, 2009) with 449 stations used in the Sp CCP stacks in yellow upside-down triangles. In red triangles we show the volcanoes (avo.alaska.edu) and slab contours in grey lines (Hayes et al., 2018). The red outline marks the extent of the Yakutat block (Eberhart-Philips et al., 2006). Labeled stations indicate the location of the single stack stations from Fig. 3.3. SP - Seward Peninsula; BR - Brooks Range; WVF - Wrangell Volcanic Field; DVG - Denali Volcanic Gap; YKB - Yukon-Koyukuk basin B) Distribution of earthquakes back azimuth that went to the CCP results.

using a time-domain deconvolution method (Ligorria and Ammon, 1999). We used the resulting impulse responses to convolve with a Gaussian whose half-width is 0.9 s. We applied the 3D velocity model from Nayak et al (2020) to migrate the receiver functions to depth.

3.2.3 Quality Control

To ensure that very noisy or physically implausible receiver functions were not included in the CCP stack, we removed individual receiver functions that did not pass quality control tests for overall amplitude levels and the presence of a Moho phase.

Two criteria were based on comparing the observed receiver function amplitudes relative to synthetic receiver function amplitudes predicted by the Nayak et al. (2020) velocity

model for Alaska. However, we stress that the acceptable ranges of receiver function amplitudes are broad enough that the choice of reference velocity model has almost no impact on the selected receiver functions. Following the approach in Hua et al. (2020a), we generated synthetic receiver functions for 1D velocity models based on Nayak et al. (2020) at 1° horizontal increments, using the propagator matrix method (Keith and Crampin, 1977) and S wave ray parameters that are consistent with the observed receiver functions. From the predicted receiver functions for the entire region, for each depth, we found the mean synthetic amplitude value, forming the vector RF_{mean} , and the half-width (half the range between minimum and maximum amplitudes, RF_{hw}). We discarded observed receiver functions (RF_o) with abnormally large amplitudes, identified by $\|RF_o - RF_{mean}\|_2^2$ greater than five times the median $\|RF_o - RF_{mean}\|_2^2$ from all receiver functions. We also counted depth layers greater than 60 km where the receiver function amplitude RF_o is either smaller than $RF_{mean} - 0.8 \times RF_{hw}$ or larger than $RF_{mean} + 0.8 \times RF_{hw}$. To remove sustained large amplitudes in the mantle, we removed observed receiver functions whose depth layer count was greater than the median from all observed receiver functions. This second criterion is the most stringent of our tests, as it eliminates half of the data and significantly decreases noise in the mantle depth range.

Given that clear Moho phases are observed across Alaska (e.g., Miller et al., 2018; Zhang et al., 2019; Ward and Lin, 2018; Mann et al., 2022), we also required that observed receiver functions have a positive amplitude in the potential Moho depth range of 15 to 60 km depth. Among these, we discarded receiver functions with positive amplitudes smaller than 20% of the median amplitude in this depth range from all receiver functions. We discarded receiver functions with negative phases five times larger than the median amplitude from all receiver functions in the 15 to 60 km depth range.

20,210 Sp receiver functions (28% of the initial potential paths) remained after all quality control criteria were applied. Examples of observed receiver functions stacked at single stations are shown in Fig. 3.3. Clear phases corresponding to positive velocity gradients at the Moho (red/positive amplitudes) are observed at all stations. Stations in central and southern Alaska (H20K, I23K, I17K, J26L, L19K, N18K, N16K, O15K) show phases consistent with negative velocity gradients in the 50-100 km depth range. In

contrast, the two stations in the northern Arctic Alaska terrane (TOLK and C23K) have weaker and deeper negative phases in the potential LAB depth range.

3.2.4 CCP Stacking

To map individual *Sp* receiver functions to 3D space, we employ common conversion point stacking (Fig. 3.2). For the source-receiver pair represented by each receiver function, P wave ray paths are traced backward from the station through the grid of CCP points. For each layer of points in depth, the amplitude at the time in a receiver function that corresponds to that depth is added to the stack for a given grid point using a horizontal spatial weighting function. In contrast to prior studies that employ geographic bins of conversion points or weighting functions that reflect the Fresnel zones of vertically incident waves (e.g., Dueker and Sheehan, 1997; Kind et al., 2012; Lekić and Fischer, 2017), in this study we incorporate the properties of *Sp* sensitivity kernels at the true ray parameter of the incident S wave (Hansen and Schmandt, 2017; Mancinelli and Fischer, 2017; Hua et al., 2020a), using the spatial weighting functions developed by Hua et al. (2020b). The standard deviation of the stacked receiver function amplitude is used to quantify uncertainty in the CCP stacks. We applied a theoretical expression developed by Hua et al. (2020b) for precisely estimating the standard deviation of any weighted average, eliminating bootstrapping requirements.

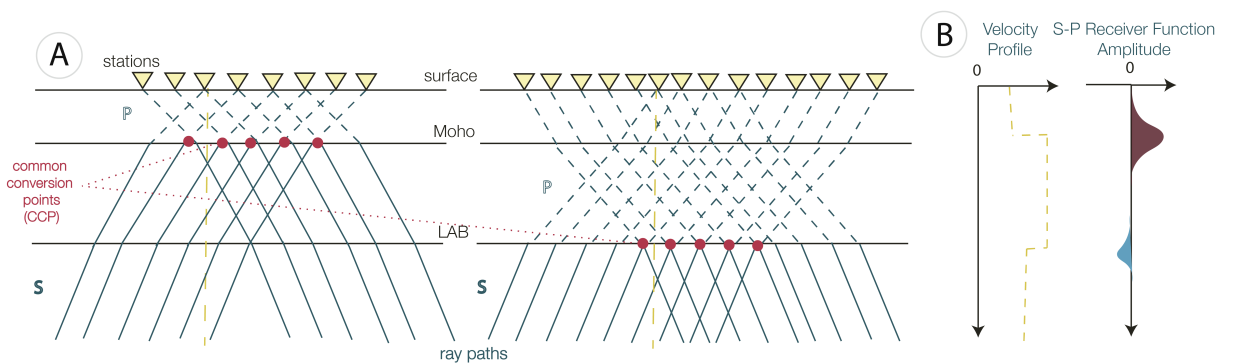


Figure 3.2: A) Schematic of Earth's layers and common conversion points at boundaries. The S -waves convert to P waves at the Moho and lithosphere-asthenosphere boundaries (LAB) and the waveforms are stacked together. We can trace and process the waveforms to get an image of the Earth in terms of positive and negative phases, and each 1D vertical slice (dashed in yellow) is a receiver function. B) The vertical slice from the CCP stack with layers with a particular velocity profile (left) will produce the receiver function on the right, schematically. As in the results figures below, the red pulse represents an increase in velocity, and the blue represents a decrease in velocity.

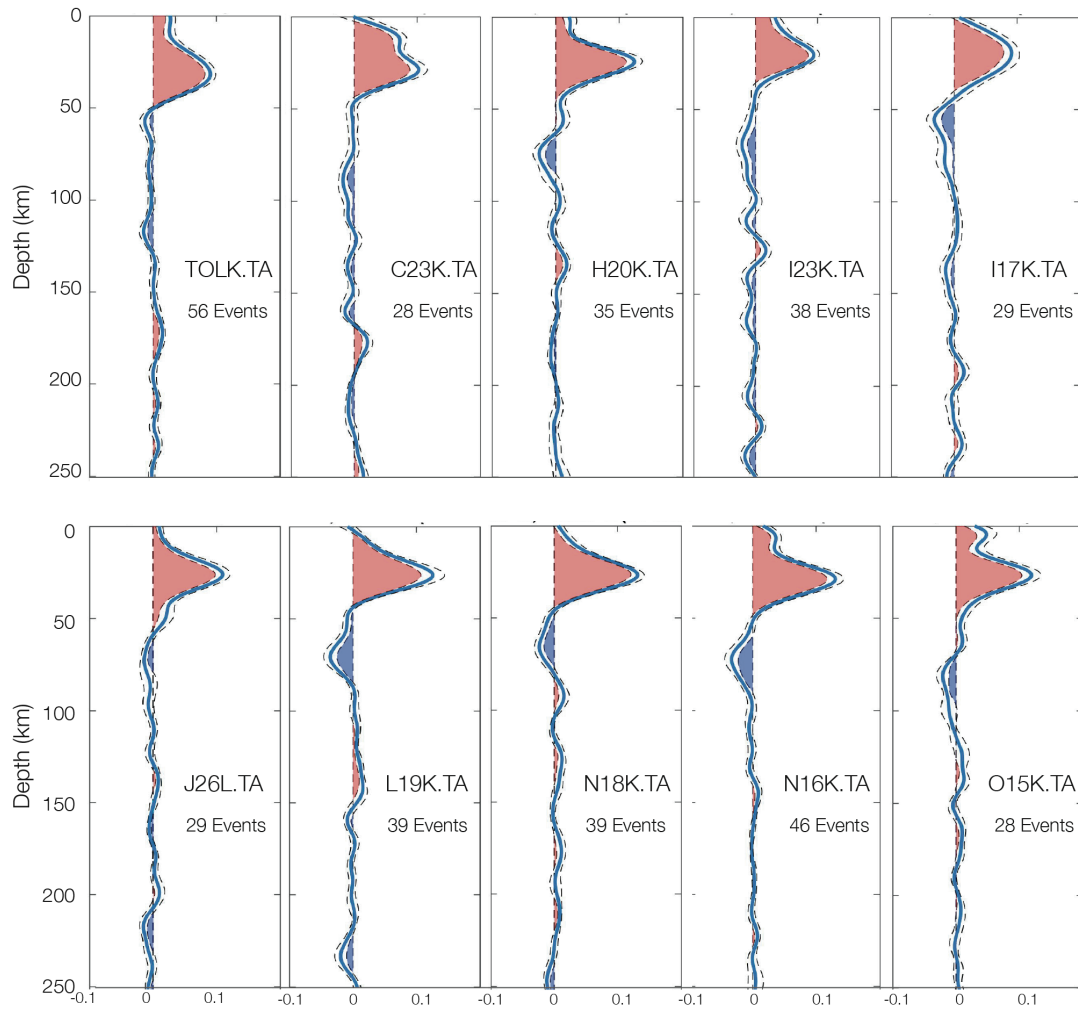


Figure 3.3: Single station stacks of S_p receiver functions with Moho filters on Transportable Array stations. Stacks were calculated using earthquakes from all back azimuths. Station locations are shown in Fig. 3.1.

3.3 Results

From the CCP stacks, we resolved three primary velocity gradients: 1. The Moho in the 15-66 depth range; 2. The lithosphere-asthenosphere boundary is mainly in the 50-170 km depth range; 3. deeper positive velocity gradient in the mantle in the 130-230 km depth range (hereafter called the PVGm). The results of the CCP stacks are shown in: a map of the summed CCP weights for all receiver functions in the stack at a depth of 200 km, also with profile locations (Fig. 3.4), profiles with depth (Figs. 3.5 - 3.7) and in maps (Figs. 3.8 - 3.10). In the stacks, positive velocity gradients at the Moho or in the mantle are red, and the stronger phases in blue represent a significant decrease in velocity, such as the lithosphere-asthenosphere boundary.

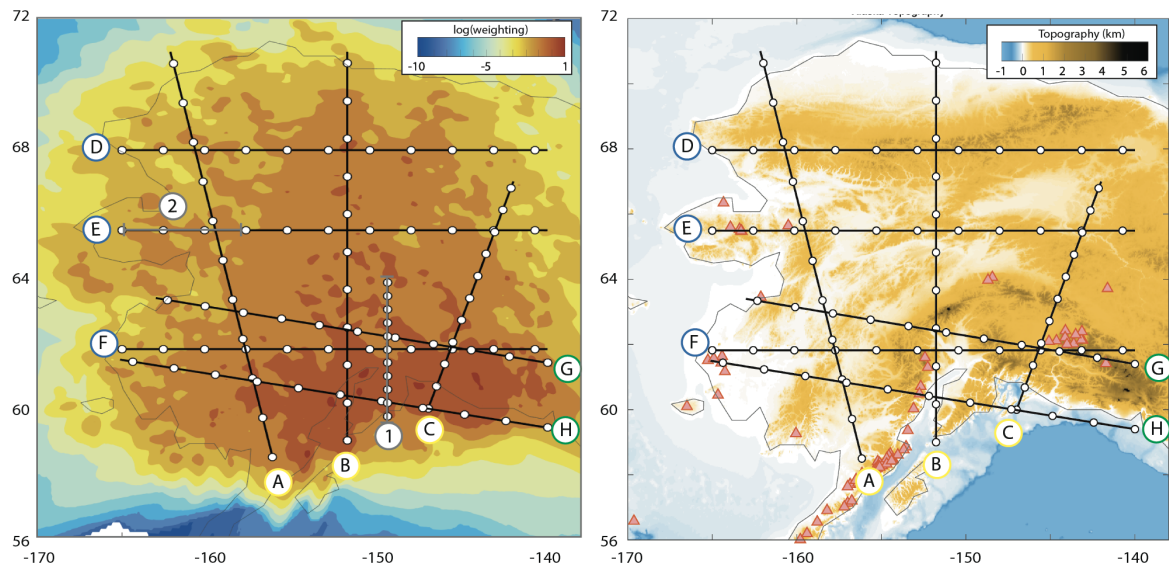


Figure 3.4: Maps show the locations of the stacks and in color the (left) weighting of the CCP volume in all regions of Alaska at 200km and (right) the topography based on ETOPO1 (Amante and Eakins, 2009). The isochron's slope, the geometrical distance between the scatterer and the station, and the depth offset between the scattering depth and the isochron all go into the weighting function. The isochron slope angle is equal to the angle formed by the phase delay time gradient and the vertical axis, as calculated by Hua et al (2020a). Darker brown on weighting map shows areas of higher weighting/ higher density of stations and data. Light blue and light yellow show areas of lower weighting.

To ensure that we interpreted the CCP stack only where the observed phases are robust, we defined cutoff criteria for interpretation based on sampling and stack standard

deviation. We ignored regions of the stack with low sampling based on their relative weight values (values of less than 0.1 and 0.05 for the PVGm and the LAB, respectively) or where the amplitude value did not exceed two times its standard deviation. We also ignored minor amplitude variations below an amplitude cutoff of 0.003 to ensure we interpret only significant velocity variations. To measure the depths of the Moho, LAB, and PVGm, we picked the largest negative or positive amplitudes within the depth window of each feature (10- 70 km for the Moho, 40-200 km for the LAB, and 100-250 km for the PVGm) using the algorithm of Hua et al. (2018). We retained only those picked phase depths that are continuous within a 0.5-degree radius, and we defined the depth uncertainty by the width of the selected peaks, as seen in Fig. 3.7a for Moho picks and Fig. 3.7b for LAB and PVGm phases.

3.3.1 Crust-mantle Boundary

Profiles through the CCP stack show the crust is thickest beneath the Brooks Range in northern Alaska (Figs. 3.5A, B) and where Yakutat crust is underthrust beneath North American crust in the south (Figs. 3.5B, F; Figs. 3.6C, G, H). In the latter case, the underthrust Yakutat terrane is connected to the seismicity of the subducting lithosphere (e.g., Fig. 3.5c and Figs. 3.6G, H). To pick a Moho depth where the Yakutat crust is underthrust beneath the North American crust, we selected the larger amplitude of the two Mohos (e.g., Fig. 3.7(1)) in the 10-70 km depth range, unless the peaks were so close in depth that they could not be distinguished (e.g., Hua et al., 2018). In some cases, a relatively clear Moho phase continues below depths of 70 km (e.g., Fig. 3.6G, H; Fig. 3.7(1)), but at these depths the subducting lithosphere is no longer in contact with the upper plate, and we do not use these deeper portions of the subducting Moho in the Moho picks shown in Fig. 3.8. Moho picks were retained only where the depth of the pick lies more than 5 km beneath the first non-zero amplitude in the stack at that latitude and longitude. Some of the underthrust Yakutat crust profiles also show negative energy at the interface between the North American and Yakutat Mohos (e.g., Fig. 3.5F and Fig. 3.6G, H). This interface has also been observed in other studies (e.g., Fuis et al., 2008;

Mann et al., 2022) with higher resolution than is possible in this CCP stack, where the data have been filtered with a 2-100 s bandpass.

Moho depths picked from the CCP stack range from 15 km to 66 km (Fig. 3.8). Their patterns are broadly consistent with previous studies (e.g., Allam et al., 2017; Miller et al., 2018; Ward and Lin, 2018; Zhang et al., 2019; Mann et al., 2022) where they overlap, although the maximum Moho depths from the CCP stack exceed those in Miller et al. (2018) and Zhang et al., (2019). In addition, Moho estimates from the CCP stack extend farther and/or more continuously across Alaska than in most prior studies. As seen in the profiles, the thickest crust in the maps lies beneath the Brooks Range (in the north of Fig. 3.8) and where the Yakutat crust is most deeply underthrust (in the south). Variations in crustal thickness are associated with some major faults, for example a northward decrease in crustal thickness across the eastern Denali fault (also resolved by Allam et al., 2017, Miller et al., 2018, Haney et al. (2020) and Mann et al., 2022).

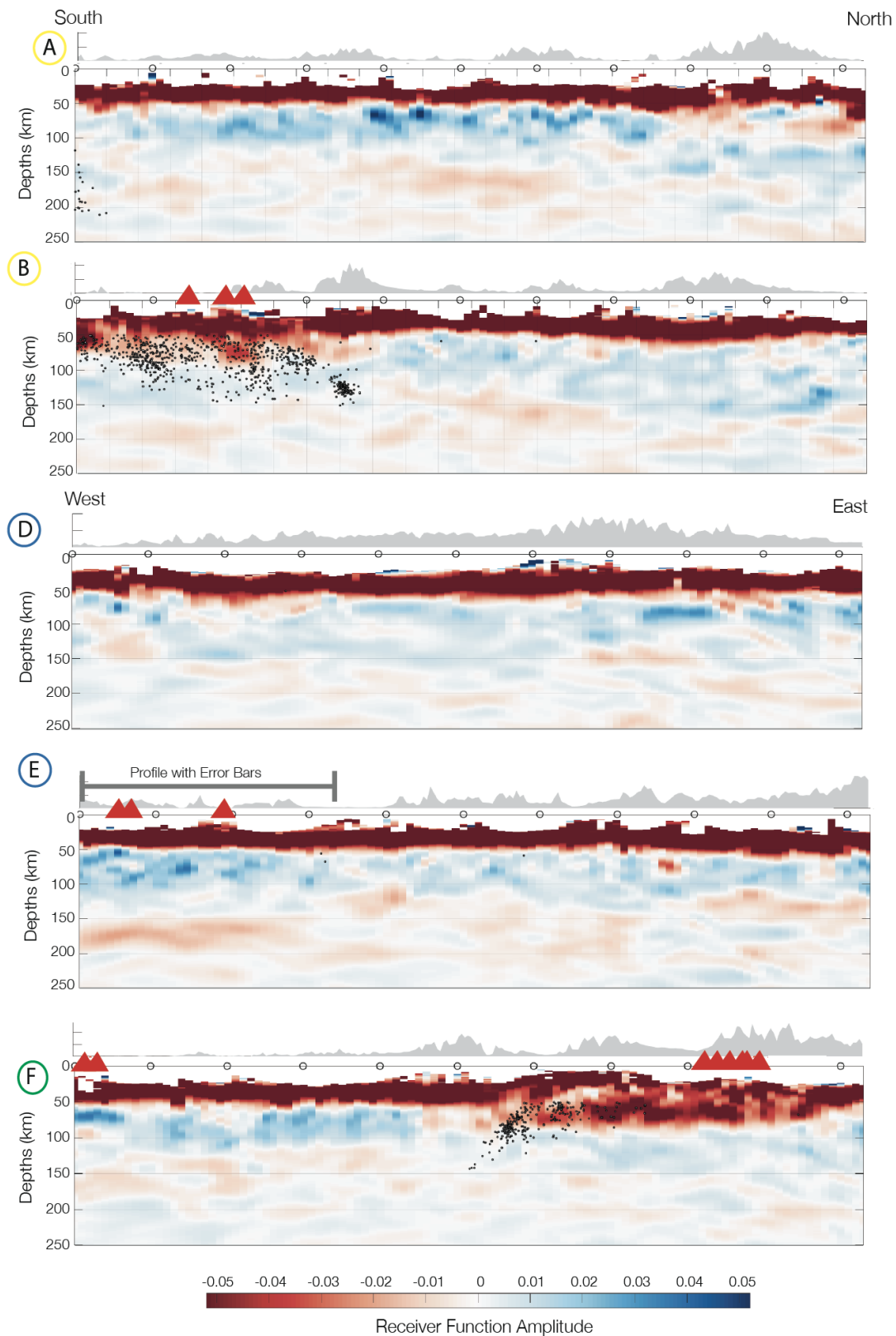


Figure 3.5: CCP stacks showing amplitude in the receiver functions (red-blue) with depth. The location of these profiles are shown in Fig. 3.4. Grey shading at the top shows topography from ETOPO1, Amante and Eakins (2009). Black small dots are the seismicity within 50 km from the profile. Volcanos at the top are also from 50 km away from the profiles. White circles at the top marks the location of each part of the stacks relative to the map.

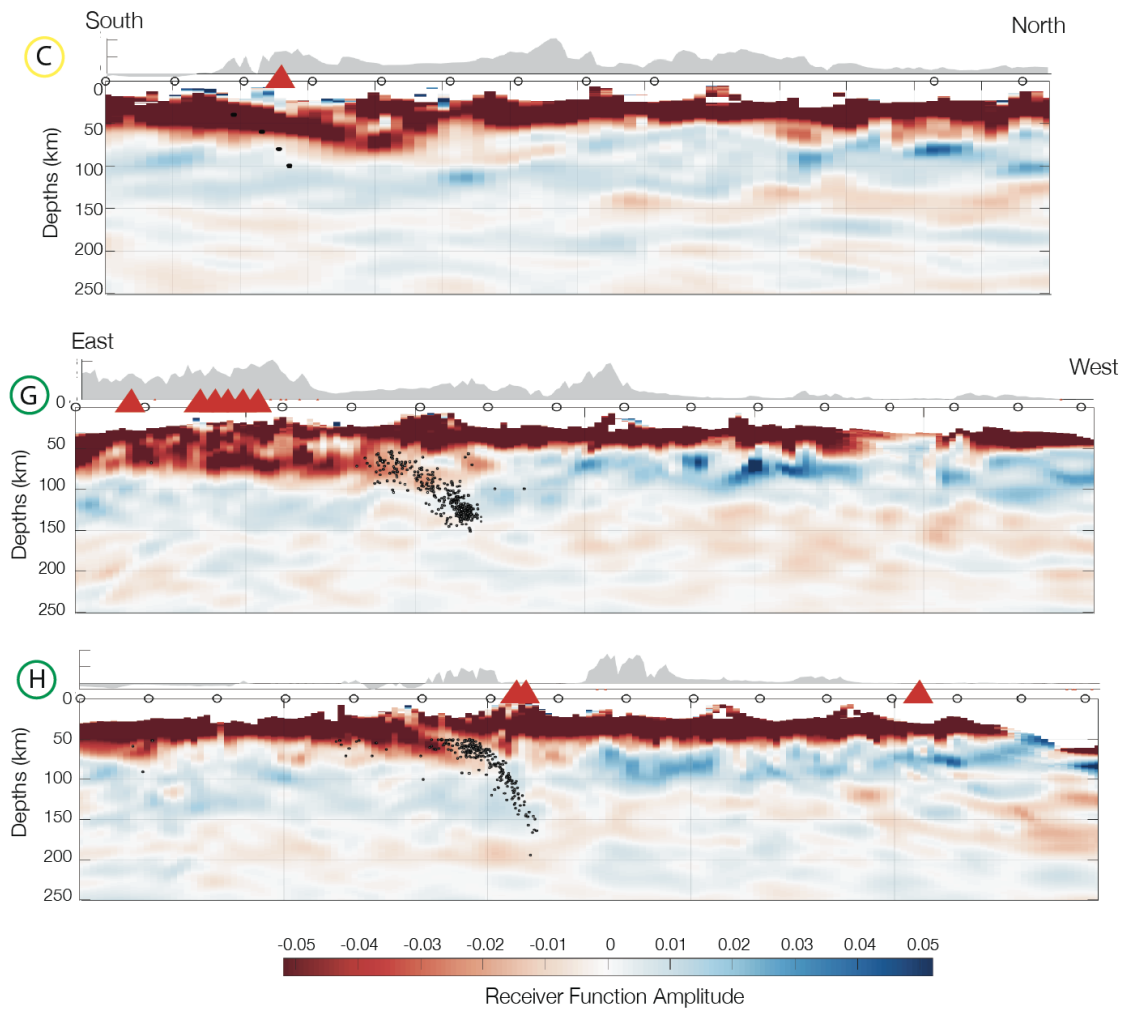


Figure 3.6: cont. CCP stacks showing amplitude in the receiver functions (red-blue) with depth. The slab contours in profile C are from seismicity values from Daly et al. 2020.

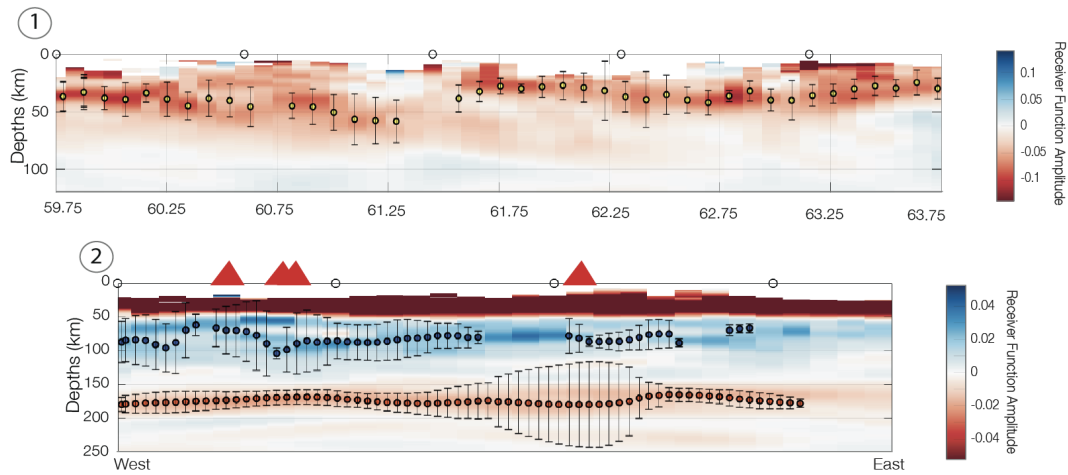


Figure 3.7: (1) Examples of Moho picks using the approach of Hua et al., 2018. (2) Examples of stacks with the LAB (blue dots) and PVG (red dots) error bars. The locations of the profiles are shown in grey Fig. 3.4.

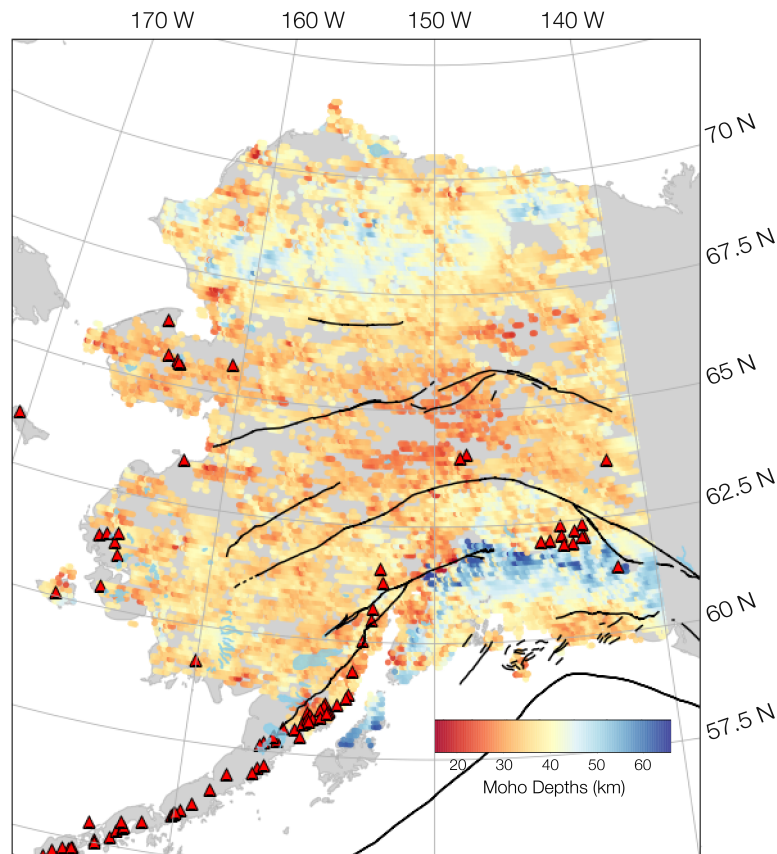


Figure 3.8: Maps of Moho depth measured from the CCP stack. Volcanoes are indicated in red triangles and black lines are major faults (Elliott and Freymueller, 2020)

3.3.2 Lithosphere-Asthenosphere Boundary

At depths of 50 to 170 km, a negative phase in the CCP stack signifies a velocity decrease in the upper mantle. Based on our prior joint inversions of Sp and surface-wave phase velocities (Gama et al., 2021) and other tomography studies (e.g., Ward and Lin, 2018; Feng and Ritzwoller, 2019), we interpret this feature as the lithosphere-asthenosphere boundary.

The lithosphere is relatively thin in a broad swath of central Alaska (50-80 km) and the Seward Peninsula (~ 70 km), as seen in the map in Fig. 3.9; the middle of the profiles in Figs. 3.5A, B, E; the western end of Fig. 3.5F; the northern side of Fig. 3.6C, and the western sides of Figs. 3.6G and H. In contrast, the lithosphere is thicker in northernmost Alaska (120-170 km) beneath the northern Arctic-Alaska terrane (northern end of Figs. 3.5A, B; middle of Fig. 3.5D). and also beneath the subducting plate in the east (~ 130 km) (eastern end of Fig. 3.6H). These broad patterns in relative lithospheric thickness are consistent with prior studies (e.g., Wang and Tape, 2014; O'Driscoll and Miller, 2015; Martin-Short et al., 2018; Jiang et al., 2018; Ward and Lin, 2018; Feng and Ritzwoller, 2019; Berg et al., 2020; Gama et al., 2021).

In terms of amplitudes, Sp phases from the negative velocity gradient at the base of the upper plate are strongest in west-central Alaska, where LAB depths are shallower depths, as seen in darker blue in regions in Figs. 3.5 and red in Fig 3.9b. Joint inversions of Sp data at single stations with Rayleigh phase velocities show comparable LAB depths and low asthenospheric velocities (Gama et al., 2021). Given that partial melt in the asthenosphere has been invoked to explain particularly strong LAB velocity gradients (e.g., Rondenay et al., 2010; Hopper et al., 2014; Rychert et al., 2013, 2018), we tested whether the LAB phase has a systematically larger amplitude near volcanoes. It does not. The mean amplitude of the LAB phase within a 50 km radius of all volcanoes and the mean amplitude of all points are not significantly different.

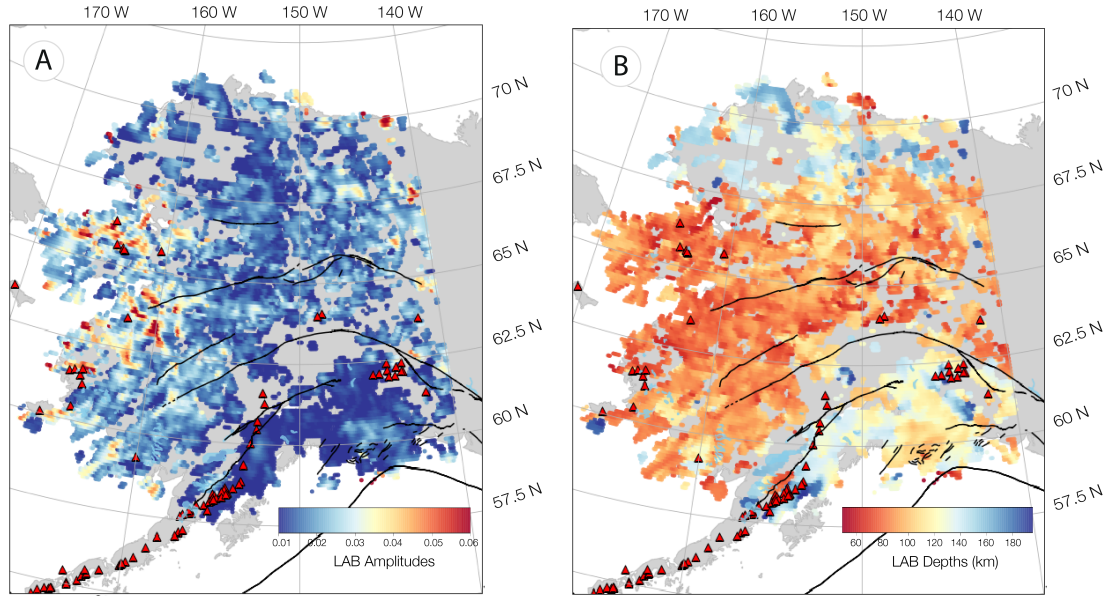


Figure 3.9: Maps of negative mantle velocity gradient (A) amplitudes and (B) depths calculated from the CCP stacks that we interpret as lithosphere-asthenosphere boundary (LAB). Volcanoes are indicated in red and black lines are major faults (Elliott and Freymueller, 2020).

3.3.3 Mantle Positive Velocity Gradient

Below the lithosphere, we see a *Sp* phase corresponding to a positive mantle velocity gradient (the PVGm) (maps of depth and amplitude in Fig. 3.10). This velocity gradient and the LAB above it are strongest and most coherent in western Alaska, including magmatically active regions such as the Seward Peninsula (e.g., left/west of Fig. 3.5E). However, higher amplitudes are also found in eastern Alaska at its central latitudes (e.g., right of Figs. 3.5E and 3.6C). Similar to the LAB calculation, magmatically active regions (within 20 km of volcanoes) do not show average phase amplitudes that are significantly different from the mean value of all PVGm picks. Based on comparisons with shear wave velocity structure obtained by inverting surface wave data at periods that resolve structure in the PVGm depth range (e.g., Martin-Short et al., 2018; Gama et al., 2021), we interpreted this positive mantle phase as the lower boundary of a low-velocity asthenospheric layer.

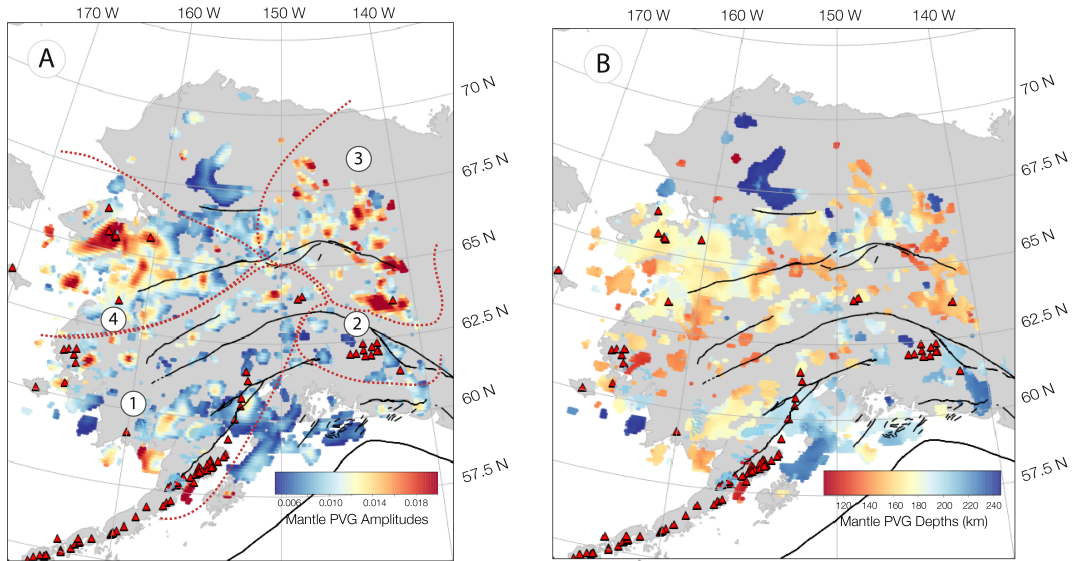


Figure 3.10: Maps of deeper positive mantle velocity gradient (PVG) (A) amplitudes and (B) depths calculated from the CCP stacks. Volcanoes are indicated in red and black lines are major faults (Elliott and Freymueller, 2020). The red dashed lines in the amplitude map on the right shows regions of different PVGm origins that are further explored in the discussion.

3.4 Implications for Subduction, Mantle Flow and Melting beneath Alaska

The overall picture provided by the Moho and mantle discontinuities that appear in the Sp CCP stack (Fig. 3.11) includes: the subduction of the Pacific lithosphere with a well-defined LAB; the thickest crust associated with the underthrust Yakutat terrane (seen more clearly in Figs. 3.5 and 3.6); the thinnest lithosphere just north of subduction zone seismicity; the thickest lithosphere beneath northern Alaska, including thickened crust beneath the Brooks range; and the deeper PVGs marking the base of a low-velocity layer.

3.4.1 Variations in Lithosphere Thickness

The thinnest lithosphere, observed in south-central Alaska, likely represents a modification of the upper plate by slab-derived processes, such as melting in the wedge associated with mantle wedge flow. This interpretation is consistent with the position of this zone to

the north of the deepest seismicity associated with the downgoing plate (Figs. 3.5F and 3.6G and H) and low asthenospheric velocities (Gama et al. 2021). The LAB phase is typically absent, as expected, where the subducting lithosphere intersects the LAB depth range (Fig. 3.9). Profile H (Fig. 3.6) is an exception, likely reflecting the tendency of CCP stacking to horizontally smooth velocity discontinuities. An interesting nuance is that the LAB phase is also weak directly above the deepest seismicity (e.g., Figs. 3.5F, 3.6H, 3.9), suggesting disruption of a pronounced negative velocity gradient at the base of the upper plate by rising melt and slab-derived fluids which lower upper plate velocities. However, such LAB disruption is less apparent within the Denali Volcanic Gap (Fig. 3.6G), perhaps due to cooler temperatures in the mantle wedge (Rondenay et al., 2010; Martin-Short et al., 2018) and lower degrees of partial melt.

However, the thin upper plate lithosphere that extends far northwest of the subducting plate to the Seward Peninsula, where low asthenospheric shear velocities (Gama et al., 2021) and surface magmatism also occur, requiring additional explanation. The thin lithosphere is likely in part due to tectonic history. The Seward Peninsula and its immediate surroundings experienced significant extension at 130-70 Ma (Dumitru et al., 1995; Hannula et al., 1995). Nonetheless, the fact that the lithosphere remains thin today and the ongoing history of magmatism require more recent processes. A variety of origins have been proposed for magmatic activity in this region, including back-arc extension (e.g., Nakamura et al., 1977; vonDrach et al., 1986; Swanson et al., 1987), hotspot activity (e.g., Moll-Stalcup 1994), and continental rifting (Swanson et al. 1981). However, more recent constraints on the age and rates of eruptions support an acceleration of magmatic flux since 6 Ma, consistent with decompression melting (Mukasa et al., 2007). Upwelling flow could be a driver of the decompression melting, given the strongly 3D nature of mantle flow that extends to the Seward Peninsula as well as latitudes near the northern Arctic Alaska terrane, based on studies of mantle azimuthal anisotropy and regional mantle flow modeling (e.g., Jadamec et al., 2013; Finzel et al., 2015; Venereau et al., 2019; McPherson et al., 2020) and the pattern of surface velocities that indicate the westward extrusion of southwestern Alaska (Elliott and Freymueller, 2020). Continued ablation of the upper plate lithosphere by such 3D asthenospheric flow is one potential mechanism

for maintaining the thin lithosphere beneath the Seward Peninsula.

Lithospheric foundering processes such as delamination or viscous downwelling at the edge of thick lithosphere (e.g., Bird, 1979; Conrad and Molnar, 1997; van Wijk et al., 2008; West et al., 2009) may also have contributed to thinning the Alaskan lithosphere. While this idea is more speculative for the Seward Peninsula, further to the east observations are consistent with lithospheric foundering playing a role in the relatively thin lithosphere in interior Alaska that persists north to the southern margin of the Brooks Range and the northern Arctic Alaska terrane (Figs. 3.5A, B, Fig. 3.9a). Consequences of foundering include decompression melting from the asthenosphere that rises in response to downwelling, and crustal extension (West et al., 2009), both of which were widely observed in interior Alaska south of the Brooks Range at 30–40 Ma (Andronikov and Mukasa, 2010). Moreover, the Fort Hamlin Hills basalts that erupted south of the Brooks Range at ~ 31 Ma involved melting of the lithospheric mantle (Andronikov and Mukasa, 2010), another possible indication of lithospheric foundering. If lithospheric foundering did occur in interior Alaska, or beneath the Seward Peninsula, vigorous 3D flow (e.g., Jadamec et al., 2013; Finzel et al., 2015) has likely helped maintain a relatively thin lithosphere at these latitudes since the initial lithospheric thinning occurred.

In contrast, the thick lithosphere indicated by the deep LAB phases beneath the Brooks Range and the northern Arctic-Alaska terrane is cold, as shown by its low velocities (e.g., Gama et al., 2021), which are similar to the global average for Archean cratons (French et al., 2013), and low heat flow (Batir et al., 2016). In addition, this region is characterized by low deformation rates (e.g., Finzel et al., 2015; Elliott and Fremueller, 2020) and high mantle viscosity (Finzel et al., 2015). Beneath much of the thick lithosphere in the north, the LAB phase is too weak to meet the criteria for depth measurements. This is likely because the LAB velocity gradient is more gradual, for example, occurring over depths of 50 km or more (Gama et al., 2021), which makes it harder to identify the LAB with converted waves (e.g., Ford et al., 2010; Rychert et al., 2007). The deep LAB observed beneath the subducting plate in the east (~ 120 km) (also seen in O'Driscoll and Miller, 2015 and Gama et al., 2021) can be modeled as an oceanic plate that has cooled typically (Hasterok, 2013) for its age, plus the additional thickness of the crust of the Yakutat

terrane (e.g., Fuis et al., 2008; Müller et al., 2008; Christeson et al., 2013; Gama et al., 2021).

3.4.2 Patterns of Positive Velocity Gradients

Mantle PVGs in a similar depth range to those we observe beneath Alaska have been imaged in other regions, primarily zones characterized by high mantle temperatures and magmatic activity, including Hawaii (Rychert et al., 2013), the northwestern U.S. (Hopper et al., 2014), Anatolia in Turkey (Hua et al., 2020a), the eastern Iranian plateau (Wu et al., 2021), and high-temperature mantle globally (Hua et al., 2022), leading to their interpretation as the lower boundaries of low-velocity layers containing partial melt. In Alaska, two aspects of the PVGm distribution support its interpretation as the base of a melt-bearing mantle layer. First, the PVGm drops in amplitude and fades away beneath the thicker lithosphere in northern Alaska, which appears to be a lower temperature region based on its seismic velocities (e.g., Martin-Short et al., 2018; Gama et al., 2021) and heat flow (Batir et al., 2016). Second, the spatial correlation of higher amplitude Sp energy from the PVGm with that from the LAB (e.g., the Seward Peninsula and east-central Alaska) are consistent with the idea the clear PVGm arrivals are due to the presence of partial melt since melt could sharpen velocity gradients at both the top and bottom of a melt-bearing layer. The observation that LAB and PVGm amplitudes do not, on average, increase locally beneath volcanoes (sections 3.3.2 and 3.3.3) suggests that partial melt is widely distributed within the apparent low-velocity layer.

To evaluate how the PVGm and the potential melt-bearing layer relate to variations in mantle melting processes across Alaska, we separated PVGm observations into four different regions (Fig. 3.10A): partial melting that shows the signature of subduction processes (1) in the primary subduction zone, and (2) around the Wrangell Volcanic Field (WVF) in the east; and partial melting from mantle decompression due to (3) mantle flow around the edge of the slab or a slab tear, with (4) upwelling flow beneath the Seward Peninsula related to the extrusion of southwest Alaska (Elliott and Freymueller, 2020).

The two southern regions where the LAB and PVGm are prominent are consistent with

the presence of partial melt as a result of subduction zone processes. These processes include slab dehydration, melting in the mantle wedge, and mantle wedge flow (e.g., Preece and Hart, 2004; Grove et al., 2009; Kawakatsu and Watada, 2007; Nye et al., 2018; Wang et al., 2020). For example, the model of Wang et al. (2020) indicates a partial melt bearing zone within the mantle wedge whose lower boundary (the wet solidus) would be comparable in depth to the Alaskan mantle PVGm, and whose upper boundary (also the wet solidus) corresponds to observed LAB depths. Alternatively, the PVGm could represent a melting front related to carbon-assisted melting (Hammouda and Laporte, 2000; Dasgupta and Hirschmann, 2010; Sifré et al., 2014; Duke et al., 2014; Chen et al., 2016; Hammouda et al., 2021). In particular, Hammouda et al. (2021) invoke a redox boundary at depths comparable to the Alaskan PVGm that would allow partial melt to be stable between this boundary and the LAB.

In the primary Alaskan subduction zone (region 1), the PVGm has a higher amplitude close to the subducting plate in the southwest and is low amplitude, sporadic, or absent in the northeast, including in the Denali Volcanic Gap (Fig. 3.9b). This variation may relate to the decrease in slab dip to the northeast. For example, Grove et al. (2009) point out that as slab dip decreases, the lateral extent of the melting region widens, potentially reducing partial melt fractions and decreasing the amplitude of the PVGm associated with the base of the melting layer.

In the vicinity of the WVF (region 2), lavas show signatures of subduction (e.g., Preece and Hart, 2004; Nye et al., 2018). In addition, seismicity (Daly et al., 2021), imaging of the downgoing plate through migration of *Ps* phases (Mann et al., 2021), and attenuation measurements all argue for northeast-directed subduction. Thus, here the PVGm can also be explained as the base of a subduction-related melting zone.

Regions 3 and 4, where PVGm amplitudes are higher, lie farther from the subducting lithosphere. Their magmas do not manifest signatures typical of subduction zone arcs; instead, they are consistent with intraplate decompression melting (Mukasa et al., 2007; Andronikov and Mukasa, 2010).

In region 3, decompression could be created by an upwelling component to the flow that is circulating counter-clockwise around the edge of the subducting lithosphere (e.g.,

Jadamec et al., 2013; Finzel et al., 2015), possibly in combination with flow through a slab tear (e.g., Fuis et al., 2008; Jiang et al., 2018; Gou et al., 2019). In region 4, decompression could be due to local upwelling within the broader pattern of counter-clockwise flow from under the slab into the back arc (Jadamec et al., 2013; Finzel et al., 2015), in combination with the westward extrusion of southwestern Alaska (Elliott and Freymueller, 2020).

Although slab-derived fluids appear absent in these regions, carbon-assisted melted could still enable melting. The redox front proposed by Hammouda et al. (2021) could explain the depths of the PVGm observed in these regions (Fig. 3.10A). Alternatively, a melting front at ~ 150 km depth that produces an observable PVGm can also be created (Hua et al., 2022) with only small water content (< 200 wt ppm) and mantle potential temperatures that overlap the range for mid-ocean ridges (Bao et al., 2022). Outside of the four previously discussed regions, weak and sporadic PVGm phases are observed beneath the subducting lithosphere and beneath a small portion of the thick lithosphere in northern Alaska, sometimes at depths greater than 200 km (Fig. 3.10). If mantle melting plays a role in producing these velocity gradients, the base of a layer containing carbon-assisted partial melt is a candidate explanation (e.g., Hammouda et al., 2021); if oxidized mantle extends to the depth of the PVGm.

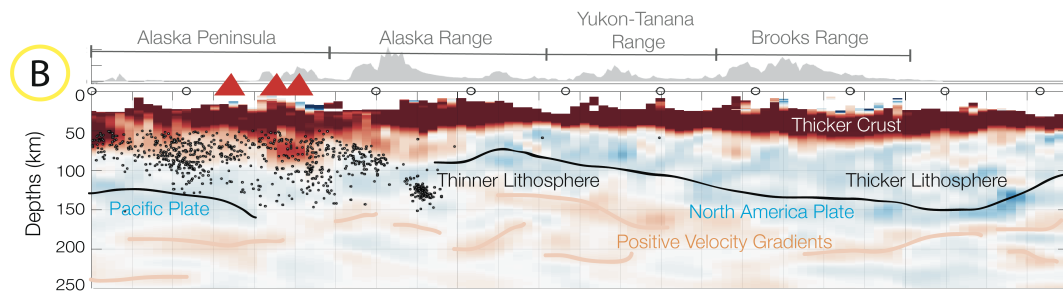


Figure 3.11: Interpretation of the structure beneath a N-S profile (Fig. 3.5B) in Alaska with major trends in lithospheric thickness and mantle discontinuities.

3.5 Conclusions

Common conversion point stacking of Sp phases recorded in Alaska by NSF EarthScope Transportable Array stations and other networks has provided new insight on crust and mantle discontinuities associated with the North American and subducting plates.

Nearly continuous measurements of Moho depth delineate the margin of the thick crust of the underthrust Yakutat terrane, a marked decrease in crustal thickness to the north across the Denali fault, and the thickened crustal root beneath the Brooks Range in the north.

New constraints on the negative velocity gradient at the LAB outline a zone of particularly thin lithosphere just beyond into the back-arc beyond the position of the deepest seismicity in the main Alaska subduction zone, consistent with thinning of the upper plate by subduction zone flow and melt rising from the mantle wedge. Directly above the deepest seismicity, LAB amplitude drops, likely due to greater infiltration of melt and heat into the upper plate that decreases the velocity contrast between lithosphere and asthenosphere.

Thin lithosphere continues far northwest and north of the subduction zone, including the Seward Peninsula and reaching the southern edge of the Brooks Range, where it meets thicker, colder and high viscosity lithosphere beneath the northern Arctic Alaska terrane. While past extensional tectonics - and possibly lithospheric foundering - could have contributed to thinning the lithosphere beneath the Seward Peninsula and interior Alaska, ongoing ablation of the upper plate by asthenospheric flow likely helps to keep the lithosphere thin.

A particularly novel structure imaged in this study is a positive velocity gradient that represents the base of a low velocity layer whose upper boundary corresponds to the LAB. Closer to the subducting lithosphere, this gradient could represent the onset of partial melting enabled by slab-derived fluids, with or without an assist from carbon in the mantle. However, this positive velocity gradient is strongest further away from the subduction zone, beneath the Seward Peninsula and to the northeast of the WVF, suggesting the presence

of partial melt in the asthenosphere hundreds of kilometers away from the slab. Magmas in these locations indicate decompression melting, consistent with components of upwelling asthenosphere.

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Chapter 4

Mantle Signatures of Terrane and Fault Zones in Alaska

This chapter is about to be submitted as: *Isabella Gama, Karen M. Fischer. et al. Mantle Signatures of Terranes and Fault Zones and Implications for Deformation in Alaska, Nature Geoscience.*

Summary

- The dramatic tectonic terrane limits in Alaska, which are both strike-slip and accretionary, offer the opportunity to investigate deformation along terrane borders and faults on a lithospheric scale.
- Fault geometries and plate kinematics are well understood in the shallow crust, but they are contested and considerably less evident deeper in the crust and mantle lithosphere.
- Here we show new seismic evidence for deeper crust and lithospheric mantle shear zones and a sharp lateral margin of the craton-like Arctic Alaska terrane in the north. We show that strike-slip faults in Alaska offset the Moho and are associated with low velocity lithosphere through to the lithosphere-asthenosphere boundary.

- The craton-like lithosphere north of the Kobuk fault appears to delineate the northern limit of present-day subduction zone impacts, acting rigidly and limiting deformation.
- Our findings of shear-wave velocity variations beneath faults suggest that different lithospheric blocks are juxtaposed where we see shear in the mantle. This discovery supports the theory that plate boundaries and deformation internal to the mantle lithosphere can occur in concentrated shear zones.

4.1 Introduction

Alaska's distinctive and active tectonic history creates a complex mosaic of accreted terranes divided by major strike-slip faults (e.g., Colpron and Nelson, 2011). Given its distinct terrane composite, Alaska provides a unique opportunity to test how the signatures of former terrane boundaries appear in lithospheric structure today, including whether deformation associated with terrane boundaries and faults persists into the mantle lithosphere. Differing hypotheses exist for the onset and progression of Alaskan terranes show the diverse source of terranes that created the structure of Alaska today. For plate restoration in Alaska, one hypothesis is a large-scale terrane wreck model (Johnston, 2001) that Alaska formed as an assemblage of four orogenic belts from a ribbon continent, primarily in south-central Alaska. South-central Alaska would have translated Arctic Alaska (AA) to its current position, and the entire reconstruction of Alaska would have happened in a span of less than 50 Ma (e.g., Johnston, 2001; Døssing et al., 2020). A second model hypothesizes that AA rotated counterclockwise around the Laurentian coastline caused by subduction from northern Alaska, and pulling this terrane to central Alaska created a subduction arc (e.g., Plafker and Berg, 1994; Clennett et al., 2020). The continuing collision would have assembled these arcs along the continental margin for over a hundred million years. Deformation has still been occurring for the past 45 Ma from subduction in southern Alaska (e.g., Finzel et al., 2015).

Today, Alaska is divided into four main groups of terranes (Colpron et al., 2007;

Colpron and Nelson, 2011) (Fig. 4.1), and the dip and subduction rate of the subducting plate varies from east to west. The Yakutat terrane is a subducting oceanic plateau that subducts with the Pacific plate, and together they present an anomalous thick crust at Alaska's southeastern margin (e.g., Fuis et al., 2008; Christeson et al., 2013), while the subducting plate to its west is only Pacific oceanic lithosphere. The deformation rates of the Pacific plate are, on average, 45 mm/yr and 23 mm/yr for the North American plate, based on the no-net rotation reference frame (Argus et al., 2011). The group of terranes is generally divided by large strike-slip faults (Fig. 4.1) that control deformation (Finzel et al., 2015; Elliott and Freymueller, 2020).

Observational and theoretical work on mantle rheology supports the possibility of localized lithospheric mantle shear zones beneath major faults that have distributed shear over widths of kilometers (e.g., Titus et al., 2007; Burgmann and Dresen, 2008; Platt and Behr, 2011; Vauchez et al., 2012; Bercovici and Ricard, 2012). Seismic studies have also found evidence for distinct structures beneath major fault zones that suggest that relatively localized deformation persists to mantle depths (Zhu and Kanamori, 2000; Chen et al., 2006; Lekić et al., 2011; Miller and Becker, 2014; Ford et al., 2014; Reeves et al., 2015; Hua et al., 2018). However, the potential for localized mantle deformation beneath faults zones remains vigorously debated (Wilson et al., 2004; Zietlow et al., 2014).

In Alaska, the distribution of deformation in the deep mantle lithosphere under strike-slip plate borders is an essential but enigmatic piece in its tectonic history. Several seismic investigations in Alaska have detected Moho offsets beneath faults (e.g., Allam et al., 2017; Martin-Short et al., 2018; Jiang et al., 2018; Miller et al., 2018) and some found evidence for distinct lithosphere-asthenosphere boundary (LAB) structures associated with terranes or faults (e.g., O'Driscoll and Miller, 2015; Berg et al., 2020; Gama et al., 2021). For example, seismic studies indicate thicker lithosphere beneath the northern Arctic Alaska terrane under the Yakutat terrane (O'Driscoll and Miller, 2015; Feng and Ritzwoller, 2019; Berg et al., 2020; Gama et al., 2021). They see a thinner lithosphere in central Alaska in the Northern Alaska terranes, Late Accreted, and the Seward Peninsula (see Fig. 4.1 for locations) (Wang and Tape, 2014; O'Driscoll and Miller, 2015). The crustal patterns across terranes are analogous to the variance in lithospheric thickness (Miller et al., 2018;

Gama et al., 2021).

Due to limited geographical sampling in past geophysical surveys, the precise positions of the terrane borders and faults with depth remain unclear. Here we provide seismological constraints based on decades (1986-2021) of broadband recordings of permanent, temporary, and the Earth Scope Transportable seismic arrays in Alaska. The dataset, which includes the results of ~ 300 inversions, is the first study to use Sp phases, which add constraints to mantle velocity gradients, in inversions with surface wave phase velocities for shear velocity beneath Alaska. Our research gives unique insights into Alaska's mantle structures and dynamics by combining these complementary seismic data and placing the inversion results in the context of surface geology, geodynamical models, and geochemistry.

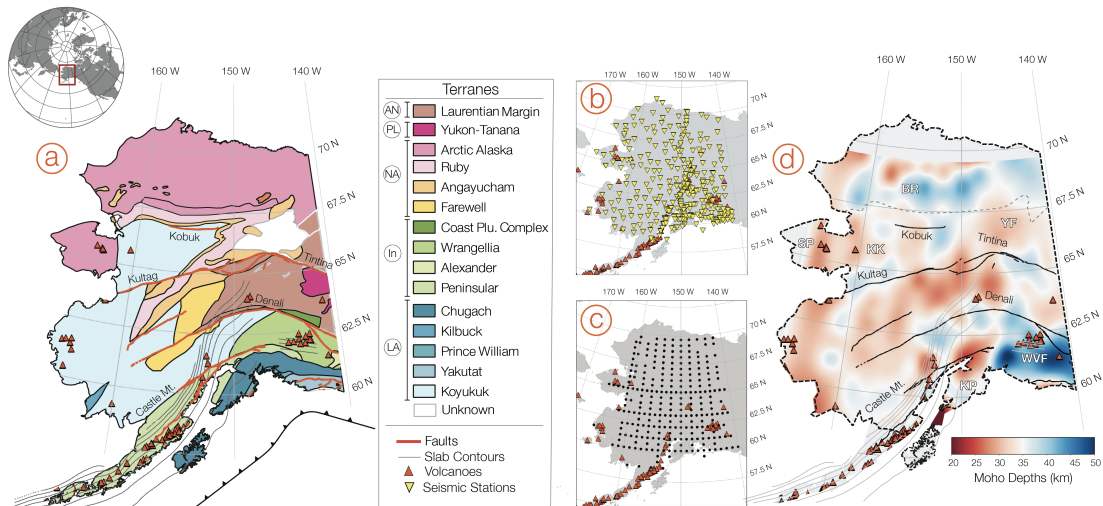


Figure 4.1: a) Geological setting of Alaska. Red thick lines are named and represent major fault zones. The tectonic terranes are based on age groups. AN: Ancestral North America, which includes both the craton and the continental margin of Laurentia. PL: Peri-Laurentia, which primarily represent Devonian to Jurassic arcs formed at the margin of ancestral North America. NA: Northern Alaska and In: Insular, which derived from other continents rather than the western margin of Laurentia, or possibly from further east in Laurentia in the case of the northern portion of the AA terrane that contains much of the Brooks Range. LA: Late Accreted terranes, which docked in the late Mesozoic to Cenozoic and includes the Yakutat (Colpron et al., 2007; Colpron and Nelson, 2011). b) The map shows in yellow all the seismic stations used to collect the data processed. c) Map of locations of the inversions shown in this paper. d) Map of interpolated Moho depths calculated in the inversion. Grey dashed line marks the boundary of the NA terranes. BR: Brooks Range; SP: Seward Peninsula; KP: Kenai Peninsula; YK: Yukon Flats; KK: Kobuk-Koyukuk basin; WVF: Wrangell Volcanic Field.

4.2 Results

4.2.1 Shear Wave Velocity Structure beneath Alaska

We generated shear wave velocity models by performing a Bayesian joint inversion of surface and body waves (see Methods section) (Figs. 4.2, 4.3, 4.4). The resulting models resolve a clear low velocity crust (shallow red to orange layer in Fig. 4.3), a higher velocity mantle lithosphere, and a pronounced low velocity layer that corresponds to the asthenosphere. Seismic velocities are valuable tools for detecting differences in the lithosphere and asthenosphere physical parameters such as temperature, pressure (Stixrude and Lithgow-Bertelloni et al., 2005), chemical composition (Selway et al., 2015), grain size (Faul and Jackson, 2005), elastically accommodated grain boundary sliding (Karato et al., 2015), near melt conditions (Yamauchi and Takei, 2016), partial melt in the asthenosphere (Debaille et al., 2020) and anisotropy (Beghein et al., 2014; Rychert et al., 2007). As a result, the contrast of shear-wave velocities (V_s) at depths ranging from 0-250 km gives important constraints on subduction zone structure, and significant structural features such as variations in the mantle lithosphere across terrane boundaries and fault systems.

4.2.2 Lithospheric Scale Shear Zones and Terrane Boundaries

The key results of our seismic models are the sharp and gradual changes in shear wave velocity across faults and terranes in Alaska's crust and mantle that can be seen in a map view (Fig. 4.2) and cross-sections that cut Alaska vertically and horizontally (Figs. 4.3 and 4.4). Both views show that strike-slip faults in Alaska offset the Moho (e.g., Fig. 4.1d, 4.2a) and disrupt the lithosphere-asthenosphere boundary (LAB) (e.g., Fig. 4.3 and 4.4) indicating that shear zones extend to the deeper crust and upper mantle.

Our results show thicker crust (Fig. 4.1d) with low velocity at depths of 40 km (Fig. 4.2a) in two locations separated by a fault or terrane boundary; In the latter map, lower velocities (< 4.1 km/s) largely correspond to crust, and higher velocities (> 4.4 km/s) represent mantle and delineate zones of particularly thin crust. One fault-related offset in

crustal thickness occurs in southeastern Alaska, where the northern margin of the thick crust of the underthrust Yakutat terrane corresponds to the Denali fault. This finding is consistent with prior studies that also documented Moho offsets at the Denali and Hines Creek faults (Allam et al., 2017; Miller et al., 2018; Haney et al. 2020). Another occurs in the far north, where the thick crust beneath the Brooks Range is bounded by the southern margin of the northern Arctic Alaska terrane (Fig. 4.1d). Between the Denali and Tintina-Kultag fault zones, a NE-SW band of the highest V_s values for 40 km depth indicates the upper mantle lithosphere beneath the collage of central Alaska terranes (Fig. 4.2a).

At depths of 100 km, the zone between the Kobuk fault and the southern boundary of the northern Arctic Alaska terrane now marks an abrupt change between thin lithosphere to the south, and thick lithosphere to the north beneath the northern Arctic Alaska terrane (Fig 4.2b). This boundary is also evident at depths of 150 km (Fig. 4.2c) albeit less consistently and dramatically. Elsewhere however, velocity variations are dominated by high velocities surrounding the zone of subducting lithosphere (e.g. around the contours marking the top of the slab in Figs 4.2b and 4.2c), and lower velocities in the surrounding asthenosphere. At 100 km, V_s is particularly low to the NE of the Wrangell Volcanic Field (WVF), and beneath SW Alaska and the Seward Peninsula. At 150 km (Fig. 4.2c), the localized lowest velocities in eastern Alaska have shifted to a position directly beneath the WVF.

As expected, cross-sections through the V_s model reveal higher velocities associated with the lithosphere subducting to the northwest in the main southern Alaska subduction zone (e.g. the high velocities surrounding the dipping zones of seismicity in Figs 4.3b and 4.4). In addition, a dipping high velocity anomaly is co-located with the zone of dipping seismicity found by Daly et al. (2021) beneath the WVF, similar to the high velocity feature resolved by Berg et al. (2020). The asthenosphere in the mantle wedge above the slab is characterized by a particularly low velocity layer (e.g. to the west of the slab in Fig. 4.3b and to the north of the slabs in Figs. 4.3c and 4.4). A more muted asthenospheric low velocity zone is also observed beneath the subducting lithosphere (Fig. 4.3b).

The cross-sections through the V_s model also show the abrupt increase in lithospheric

thickness across the southern boundary of the Arctic Alaska terrane that was evident in map-view (Fig. 4.2b). In contrast to the thin lithosphere and well-developed asthenospheric low velocity zone in central Alaska (e.g. Fig. 4.3a), the lithosphere to the north of the southern edge of the Arctic Alaska terrane reaches thicknesses of 150 km or more (Figs. 4.3c and 4.4).

However, additional and more subtle features correlated with major faults and terrane boundaries can also be seen in the cross-sections. In some locations in the relatively thin lithosphere of central Alaska, faults correlate with apparent changes in lithospheric thickness and/or local reduction in lithospheric velocities - for example, the Denali fault (Fig. 4.4), the Kullag-Tintina faults (Figs. 4.3a and 4.4), and the Kobuk fault (Fig. 4.4). Nevertheless, in some cases, the gaps in the base of the lithosphere are not as clear, such as at the Castle Mountain fault in Fig. 4.4. In addition, some variations in lithospheric structure between terranes occur independently of major faults. Overall, the flatter terranes and peninsulas, such as the Yukon flats (Fig. 4.3c), Kobuk-Koyukuk terrane or basin (Fig. 4.3a and Fig. 4.4), the Farewell terrane to the north of the Denali fault (Fig. 4.4), and the segment of the Arctic-Alaska terrane that makes up the Seward Peninsula (Fig. 4.3a) are distinguished by thin but clear high velocity lithosphere with a clear LAB at their center.

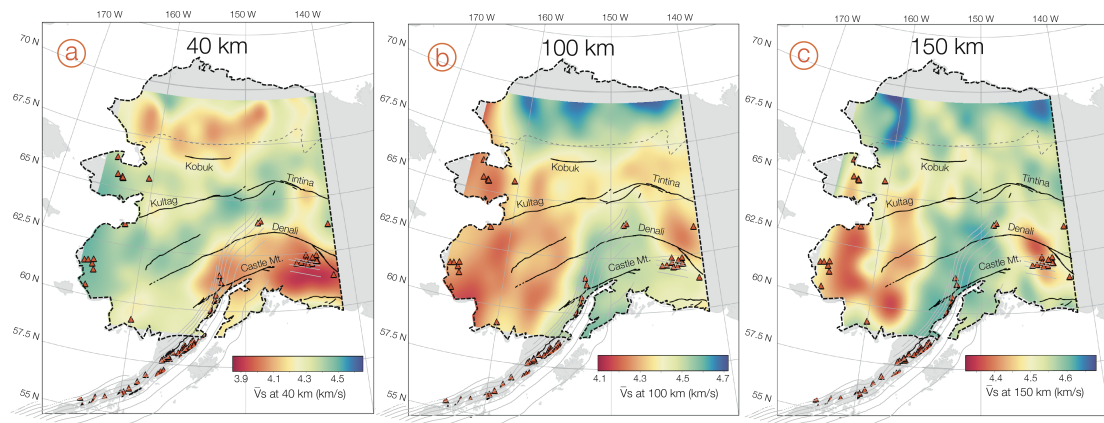


Figure 4.2: Average shear-wave velocity (V_s) at depths of a) 40 km, b) 100 km, and c) 150 km. The velocity maps are overlaid with major fault zones and slab contours. Grey dashed line marks the boundary of the NA terranes. Volcanoes are shown in red triangles.

4.2.3 Alaskan Subduction Zone - Past and Present

The velocity models reveal new features of present-day subduction zone structure. For example, the longer period phase velocities in this study enable imaging of a relatively rapid increase in velocity at base of the asthenospheric low velocity zone, in contrast to prior surface wave studies that employed only ambient noise data (e.g. Ward and Lin, 2018; Berg et al., 2020) and were not able to resolve sufficiently deep structure. Martin-Short et al. (2018) also captured this localized positive velocity gradient at the base of the asthenosphere, but only in the southern portion of our study area. However, our study shows that a well-delineated low velocity layer extends all the way to the northwest beneath the Seward Peninsula where it is particularly clear (Fig. 4.3a, similar to the findings of Gama et al. (2022), and to the north where it meets the thick lithosphere beneath the northern Arctic Alaska terrane (Figs. 4.3c and 4.4). The contrast between very low asthenospheric velocities in the mantle wedge above the slab, versus the higher minimum velocities on the ocean side of the slab (Fig. 4.3b), presumably reflects the addition of slab-derived fluids and the partial melting they produce in the mantle wedge (e.g. Rondenay et al., 2010). However, the presence of very low velocity asthenosphere ~800 km away beneath the Seward Peninsula where the signatures of slab-derived fluids are absent (e.g., Mukasa, et al, 2007) requires another mechanism, for example the introduction of hot upwelling mantle into the asthenosphere related to the strongly three-dimensional mantle wedge flow indicated by mantle anisotropy (Venereau et al., 2019; McPherson et al., 2020). A component of mantle upwelling mantle would be consistent with the signatures of decompression melting found in magmas erupted in the Seward Peninsula (Mukasa, et al, 2007; Andronikov and Mukasa, 2010; Gama et al., 2022).

More speculatively, the velocity models may also indicate the presence of a subducting or foundering fragment of lithosphere at the southern margin of the northern Arctic-Alaska terrane. At this terrane boundary in the N-S cross-section in Fig. 4.4, about 200 km south of the Brooks Range, a high-velocity feature dips to the south. One interpretation is that it is a fossil slab that has remained attached to the upper plate until recently, as has been suggested in southern California (Pikser et al., 2012; Wang et al., 2013). This

interpretation is consistent with past subduction, which in turn supports certain plate reconstruction models (Clennett et al., 2020). Alternatively, the high velocity anomaly could represent lithospheric foundering, perhaps associated with lithospheric instabilities that contributed to the steep margin of the thick lithosphere beneath the northern Arctic Alaska terrane, as has been suggested in Alaska with converted Sp waves (Gama et al., prep) and northwest Canada (Bao et al. 2014).

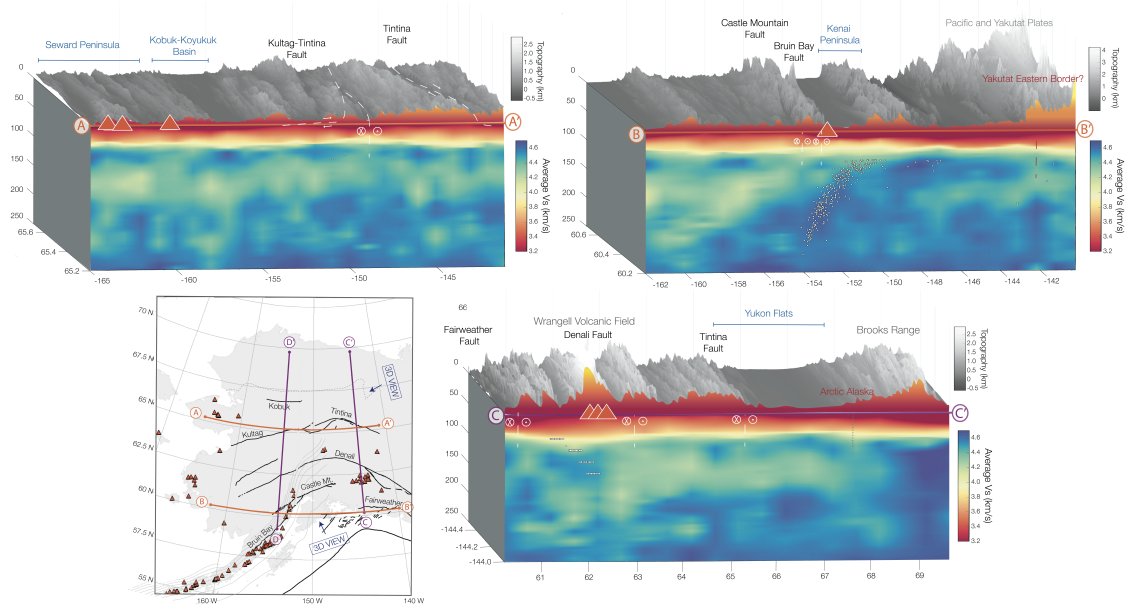


Figure 4.3: Map with profile location and 3D figures with topography based on ETOPO (Amante and Eakins, 2009) at the top in grey, and average shear-wave velocity (V_s) with depth. The V_s profiles have the same velocity bounds. Red triangles are volcanoes 50 km away from the profile, and white dots show seismicity at 50 km from the profile. In profile c-c', the white dots are slab contours at the Wrangell slab (Daly et al., 2020) 20 km from the profile. Annotation indicates major features near the profiles, including faults, terranes, and ranges.

4.3 Discussion

4.3.1 Implications for Upper Plate Deformation

The thicker lithosphere found beneath the northern Arctic Alaska terrane appears to represent high viscosity mantle with little internal deformation, as determined by geodetic and fault data (e.g., Finzel et al., 2015; Elliott and Freymueller, 2020). These craton-like

characteristics are corroborated by low heat flow and a zone of high magnetization in the deep crust of northern Alaska (e.g., Batir et al., 2016), both of which are typical of cratonic lithosphere. In contrast, the remainder of Alaska has higher heat flow and is believed to be more deformable and weaker (e.g., Saltus and Hudson, 2007; Finzel et al., 2015).

The steep southern boundary of the thick lithosphere has interesting implications for deformation. If this margin is old, for example dating from the time of terrane accretion at its southern edge, which ranges from 30-110 Ma based on different reconstruction models (Plafker and Berg, 1994; Johnston et al., 2001; Muller et al., 2019; Clennett et al., 2020), then geodynamic models (Currie and van Wijk, 2016) indicate that the mantle lithosphere must be chemically depleted and rheologically strong to have maintained this shape. Alternatively, the steep southern boundary could have been created more recently, perhaps by delamination which eroded the border of the thick lithosphere to the north, as suggested in western Canada (Bao et al., 2014).

The observed variation in lithospheric thickness at the southern boundary of the Arctic Alaska terrane, and the reductions in mantle lithosphere velocity across major faults and terrane boundaries in central and southern Alaska, add to the body of evidence that the signatures of terrane accretion and/or fault-related deformation can persist all the way to the LAB while remaining horizontally localized. The new examples provided in this study complement prior cases in Alaska (O'Driscoll and Miller, 2018), the Tibetan Plateau (Wittlinger et al., 1998), the Mackenzie craton from Canada to Alaska (Estève et al. 2020), a central part of the San Andreas Fault in California (Ford et al., 2012), Alpine Fault in New Zealand (Hua et al., 2018), Moroccan Atlas fault (Miller and Becker, 2014); and the Tanlu Fault Zone, eastern China (Chen et al., 2006).

How might ongoing faulting create localized zones of velocity reduction, such as those seen in the velocity models (Figs. 4.3 and 4.4)? One possibility is that shear-heating related to fault motion increases the temperature (e.g., Willis et al., 2019) enough to locally reduce lithospheric velocities. In addition, shear heating could be great enough to cause partial melting (Nicolas et al., 1977; Regenauer-Lieb et al., 2015), further reducing velocities and potentially affecting the distribution of Alaskan volcanism. However other

factors, such as alteration due to aqueous fluids, may also contribute to reducing V_s in the deeper crust or mantle lithosphere beneath faults. Near the Tibetan Plateau's northern edge, channels of low resistivity in magnetotelluric (MT) measurements (Zhang et al., 2015) have been used as evidence for fluids that formed by metamorphic events associated with oblique under-thrusting of distinct terrane blocks and migrated upward along the fault system, resulting in serpentinization which reduces seismic velocity (Hyndman and Peacock, 2003). Furthermore, anomalously low seismic velocity in this region of the upper mantle was interpreted as being due to the presence of material of partially underthrust crust, in the same location as the MT research (Wittlinger et al., 1998).

Overall, our observations of lithospheric thickness variations and mantle V_s reductions beneath terrane boundaries and major fault zones suggests that distinct lithospheric blocks are juxtaposed in these locations. This result supports the hypothesis that the mantle lithosphere can deform in concentrated shear zones, consistent with rheological models that allow localized strain in the upper mantle (Bercovici and Ricard, 2012).

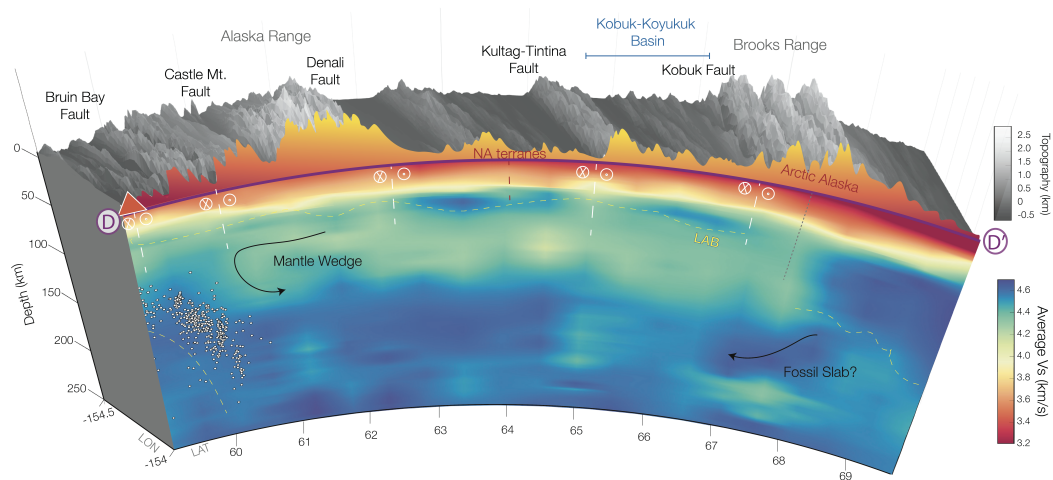


Figure 4.4: A 3-dimensional perspective of the Alaska crust and upper mantle from S-N showing average shear-wave velocity (V_s) with depth for profile d-d'. Profile location is plotted in the map in Fig. 4.3. Annotation indicates major features near the profiles, including faults, terranes, ranges, and interpretations.

4.4 Methods

We processed, inverted, and analyzed a large set of seismic data from 449 broadband seismic stations distributed in Alaska. We inverted two types of seismic data: (i) body waves converted phases and (ii) surface-wave phase velocities. (i) Sp converted phases (S waves that convert to P wave at boundaries) were used to enhance the resolution of velocity gradients such as the Moho, lithosphere-asthenosphere boundary, and a positive velocity layer at the mantle. We calculated 20210 receiver functions and created a 3D model of Sp converted phase amplitudes using teleseismic earthquake data and a common-conversion point method. (ii) Fundamental-mode Rayleigh wave phase velocities were used to constrain absolute velocities. We measured the phase velocities from ambient noise and teleseismic earthquakes between periods of 5 and 140 s. These seismic data offered complimentary information about the interior of Alaska. They were inverted to ~ 300 velocity profiles to depths of 300 km using a Bayesian joint inversion with a Monte Carlo Markov Chain search of the parameter space (Eilon et al., 2018; Gama et al., 2021).

Common conversion point stacking. Converted waves have been effective instruments for imaging velocity discontinuities in the Earth. An essential property of body waves is that these waves scatter, reflect, and refract in the presence of velocity change. We filtered (2-100s bandpass filter) and deconvolved Sp converted phases to isolate the Earth structure signals.

The common conversion point (CCP) stacking method assumes that converted waves happen near or at converted wave ray paths. The CCP result is receiver function images stacked positive and negative phases, and the amplitudes are recorded at the seismometers in the surface and projected along the ray paths. We attributed weights to these receiver functions according to the sum of amplitudes or rays in space. Since the CCP converted phases are depth domain, we migrated it to time using the reference velocity model used to generate the CCPs (Nayak et al., 2020).

From all the stations used in this study, 153 stations were the EarthScope Transportable Array project that ended in May 2021, and the remaining are other temporary

and permanent stations. We used earthquakes larger than Mw 5.8, recorded between 1986 and 2021. The distribution of back azimuths was dominated by two regions, from southeast Pacific events (215 - 255°) and north and northwest Pacific events (255 - 285°). The reader can find more details of the common conversion point method implemented by Gama et al. (2022).

Surface Wave Phase Velocities. We used fundamental-mode Rayleigh-wave phase velocities for Alaska from ambient noise and earthquake sources. For Rayleigh waves generated by earthquakes, we followed the approach used by Babikoff and Dalton (2019) to calculate phase velocities at periods 30 to 140 s. We incorporated ambient noise phase velocities for periods from 5 to 25 s, following the process of Ekström et al. (2009). The phase velocity data is shown in maps at different periods in the supplementary material (Fig. C.1) and is distributed online.

Bayesian Joint Inversion. We implemented a Bayesian joint inversion approach to determine 1D shear velocity models beneath Alaska. Our method framework has been applied and is more detailed in previous papers (Eilon et al., 2018; Gama et al., 2020). We used a Monte Carlo Markov Chain (MCMC) search on the model parameter space by randomly perturbing model parameters and using a Metropolis-Hastings criterion (e.g., Hastings, 1970) that controls the inversion progress. With this criterion, we calculated the likelihood values of each data type. We accepted models when their misfits were lower than in the previous iteration, but we allowed some higher misfits for the search to not fall in local minima. To account for sharp changes in the Moho and gradual changes in velocity that mantle features present, we parameterized our models with adaptive discontinuous cubic B-spline functions. The workflow of the method is shown in the supplement (Fig. C.2).

In addition to spline coefficients that define the velocity profiles, the parameters for which we solved include the number and depth of spline knots for the cubic B-spline functions, the hyperparameters that describe the relative uncertainties of the different data types, and the depth and velocity contrast at the Moho. The perturbed parameters are listed in table B1 with their probabilities of being perturbed.

For each of the twenty thousand iterations, we generated a synthetic Earth veloc-

ity profile, predicted the corresponding Sp converted phases and Rayleigh wave dispersion curves, and calculated the misfits of these predictions to the observed Sp RFs and Rayleigh wave phase velocities for the likelihood criteria. We generated synthetic waveforms using the propagator matrix algorithm to calculate synthetic Sp phases for each model (Keith and Crampin, 1977). Using the Mineos algorithm from Masters et al. (2011), we calculated synthetic fundamental-mode Rayleigh wave phase velocities. We solved for V_p , V_s , and density profiles using a single value of V_p/V_s in the crust was a free parameter, but we fixed this value in the mantle at each depth following the AK135 reference model (Kennett et al. 1995). We defined the density ratio to V_s after Brocher (2005) for the crust and Abers and Hacker (2016) for the mantle to solve for density.

We evaluated each candidate model using the posterior probability calculated with the prior and likelihood functions. The prior distributions for all parameters were determined empirically (Eilon et al., 2018), was roughly uniform, and were constrained by the bounds shown in Table C.1. The total likelihood function for a given model is the product of likelihoods for both data types.

We added the posterior distributions of all accepted chains to form a final probability distribution. The final velocity model was defined by the median shear velocity at each depth, and their respective median values defined crustal V_p/V_s and Moho depth.

4.5 References

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Appendix A

Shear Wave Velocity Structure beneath Alaska

Supplementary Material for: *Isabella Gama, Karen M. Fischer, Zachary Eilon, Hannah E. Krueger, Colleen A. Dalton, Lucy M. Flesch. **Shear Wave Velocity Structure beneath Alaska from a Bayesian Joint Inversion of Sp Receiver Functions and Rayleigh Wave Phase Velocities**, Earth and Planetary Science Letters, 2020. doi:10.1016/j.epsl.2021.116785*

A.1 Sp & Ps receiver functions and stacking methods

The waveforms in this study represent events with $M_w > 5.7$, depths < 100 km, and epicentral distances of $35^\circ - 80^\circ$ for Ps and $55^\circ - 80^\circ$ for Sp, resulting in thousands of receiver functions. The epicentral distance and depth windows were chosen to limit interference in the receiver functions from other teleseismic phases (e.g. Wilson et al., 2006). However, in practice receiver function stacks using data from the full range of depths are similar. In addition, we discard Sp phase time windows that correspond to post-critical incidence (e.g., Ford et al., 2010). To capture lateral variations in crust and mantle scattering, we grouped Sp and Ps phases into geographic clusters over 30° of

back azimuth. The distribution of back azimuths is dominated by two groups, 220 - 250° (southeast Pacific events) and 250 - 280° (north and northwest Pacific events).

To calculate receiver functions for individual event-station pairs, we first identified waveforms with incident/parent phases that met signal-to-noise and timing criteria. First, the parent phase was picked (S from the radial component, P from the vertical component) using Hilbert-transformed envelopes of waveforms that had been filtered with a bandpass of 0.5-100 s. This filter was chosen to retain shorter periods in order to more accurately pick phase arrivals. Picking was performed using moving signal (8 s) and noise (26 s) windows, and the phase arrival-time was identified as the time with the maximum signal-to-noise ratio (SNR, the mean amplitude in the signal window, divided by the mean amplitude in the noise window). If the picked phase arrival-time differed by more than 15 s from the arrival-time predicted by the SEMum2 model (French et al., 2013) the waveform was discarded. We then converted the waveforms to P-SV components using a free surface transform (Kennett, 1991; Bostock, 1998). The free surface velocities needed for the transform were obtained by first modeling the particle motions of all of the incident P phases in the geographic cluster to determine free surface V_s , and, assuming that free surface V_s , modeling the particle motions of all of the incident S phases to determine free surface V_p (Hua et al., submitted). As a second quality criterion, we calculated the SNR value of the parent phase component (SV for S_p receiver functions, P for P_s receiver functions) filtered with the 2-100 s bandpass. Waveforms with SNR values of less than 1.5 were rejected.

S_p and P_s receiver functions were obtained by deconvolving SV from P waveform components and P from SV components, respectively, using an iterative time-domain deconvolution (Ligorria and Ammon, 1999) with Gaussian half-widths 0.6 s for P_s and 0.9 s for S_p . Time-domain deconvolution avoids significant side-lobe energy around phases representing real structure, such as the Moho (e.g. Lekić and Fischer, 2017; Hua et al., submitted), and the Gaussian half-widths were chosen to improve stack signal-to-noise while minimizing additional filtering due to the deconvolution operation. We migrated the resulting receiver functions to the average ray parameter of all retained phases using the SEMum2 model (French et al., 2013).

Individual receiver functions were culled using K-means cluster analysis before stacking. This data mining technique groups receiver functions into clusters using a measure of their “distance” or differences. We employed five clusters with the cosine and cityblock distance functions provided by Matlab. We chose the clusters which contained a clear Moho phase (an assumption based on prior studies in Alaska which indicated clear Moho arrivals; e.g. Miller et al., 2018; Jiang et al., 2018) and which minimized energy at 0 km (as expected if the parent phase is successfully isolated on the parent waveform component). We did not make assumptions about mantle structure. After culling the receiver functions using K-means for the first time, we re-calculated the free-surface velocities with the waveforms that we selected to redo the free surface transform. This first selection of higher quality data is important to provide more reliable free surface velocities. Because the clustering process begins with a randomly selected waveform, the results of clustering vary slightly between iterations. We performed the clustering process, assuming the revised free surface velocities, five additional times, and each individual receiver function was weighted by the number of times it appeared in a selected cluster. In an example of the clustering process for Sp and Ps receiver functions for station MLY.AK (Fig. A1) individual receiver functions are shown in the leftmost panels, colored by the number of times they were selected. While this plot tends to accentuate outlier receiver functions, a heat map that shows how many times an amplitude-depth pixel is sampled by the individual receiver functions, weighted by the number of times they were selected, (second column in Fig. A1) indicates a greater degree of consistency among selected receiver functions.

After the selection of receiver functions through K-means, their median bootstrapped average was calculated to represent their stack. During the bootstrap process, we randomly selected individual receiver functions with the same number as the total number, but allowing the traces to repeat, and we calculated the median of those. We performed the same calculation one hundred times. The mean of these 100 medians and their standard deviation were taken to represent the receiver function stack. We interpreted phases only when their amplitudes significantly exceeded stack standard deviation. Finally, we migrated the stacks to depth, as seen in the third column of Fig. A1. For many stations, Sp and Ps receiver function stacks from the same station reveal Moho phases at

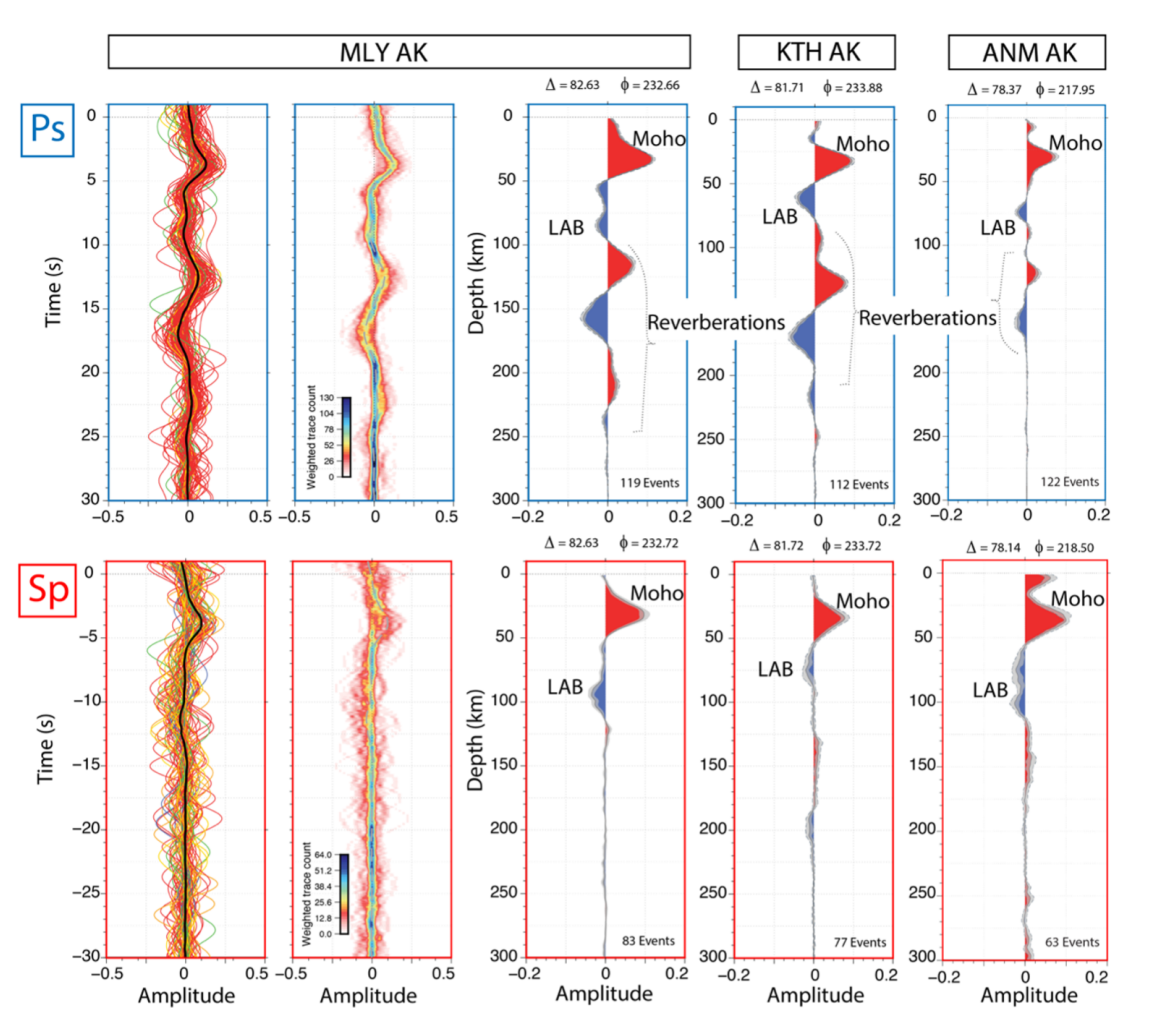


Figure A.1: (Columns 1-3) Receiver function processing step after K-means clustering for station MLY.AK for Ps and Sp. Column 1: colored lines show individual selected receiver functions colored by the number of times they were selected during K-means clustering; red (5 times) orange (4), yellow (3), green (2), blue (1); black line shows the median. Column 2: Heat map of individual receiver functions weighted by how many times they were selected. Column 3: Median of weighted individual receiver functions; grey shading shows one standard deviation uncertainties. Sp receiver functions were flipped to have the same polarity as Ps. Δ is the mean distance of the events, and ϕ is the mean back azimuth. Columns 4-5: Final stacks of Ps and Sp receiver functions for stations KTH.AK and ANM.AK. Depth-migrated receiver functions in this figure assume the SEMum2 velocity model (French et al, 2013).

compatible depths. For example, the stacks of receiver functions in Fig. A1 show that the positive (red) peaks of the Sp and Ps Moho phases, which appear between 25 and 50 km depth, agree within less than 3 km. The Sp receiver functions for these stations also contain phases (blue) consistent with negative velocity gradients, potentially at the LAB, at depths of 75-110 km. The Ps receiver functions show negative energy in simi-

lar depth ranges, but the largest phases at mantle depths are the positive and negative crustal reverberations, highlighting one of the challenges to incorporating Ps data into the joint inversion. Although in many cases Moho and LAB phases are similar for Ps and Sp receiver functions at a given station, we chose to use only Sp receiver functions in the inversion. This decision was motivated by the poor agreement between Ps and Sp at some stations, and the typical complexity in Ps stacks from reverberations. In addition, from synthetic tests, Eilon et al. (2018) demonstrated that adding Ps phases to inversions with Rayleigh waves and Sp phases does not improve resolution of shear velocity structure.

A.2 Diagnostics from Inversions with Synthetic Data

The results of the inversion with the deconvolved synthetic data indicate the stability and accuracy of the inversion process (Fig. A2). The data predicted by the ensemble of accepted models shown in Fig. 2.4d match the synthetic data from the input model well. In the upper left panel of Fig. A2, each grey line is a convolution of $SV_p(t,m)$ and the synthetic Sp receiver function and each cyan line is the convolution of $Pp(t,m)$ and the auto-deconvolved S phase. The large negative phase (~ 21 s) shows the Moho, and the positive phase (~ 17 s) is the LAB. The matches between the cross-convolutions for all accepted models illustrates the good agreement of the body-wave “data” and predictions. The bottom left plot shows good fits between the Rayleigh phase velocity “data” from the input model (circles) and the phase velocity predictions from each of the accepted models (orange lines). Misfit values for both Sp and surface wave data (shown in lower right plot for all six chains) decrease over the 8000-iteration burn-in period and are relatively stable over the subsequent iterations from which the accepted models are drawn. Comparable plots for the other two synthetic inversion cases in Fig. 2.4 can be found in Figs. A3 and A3.

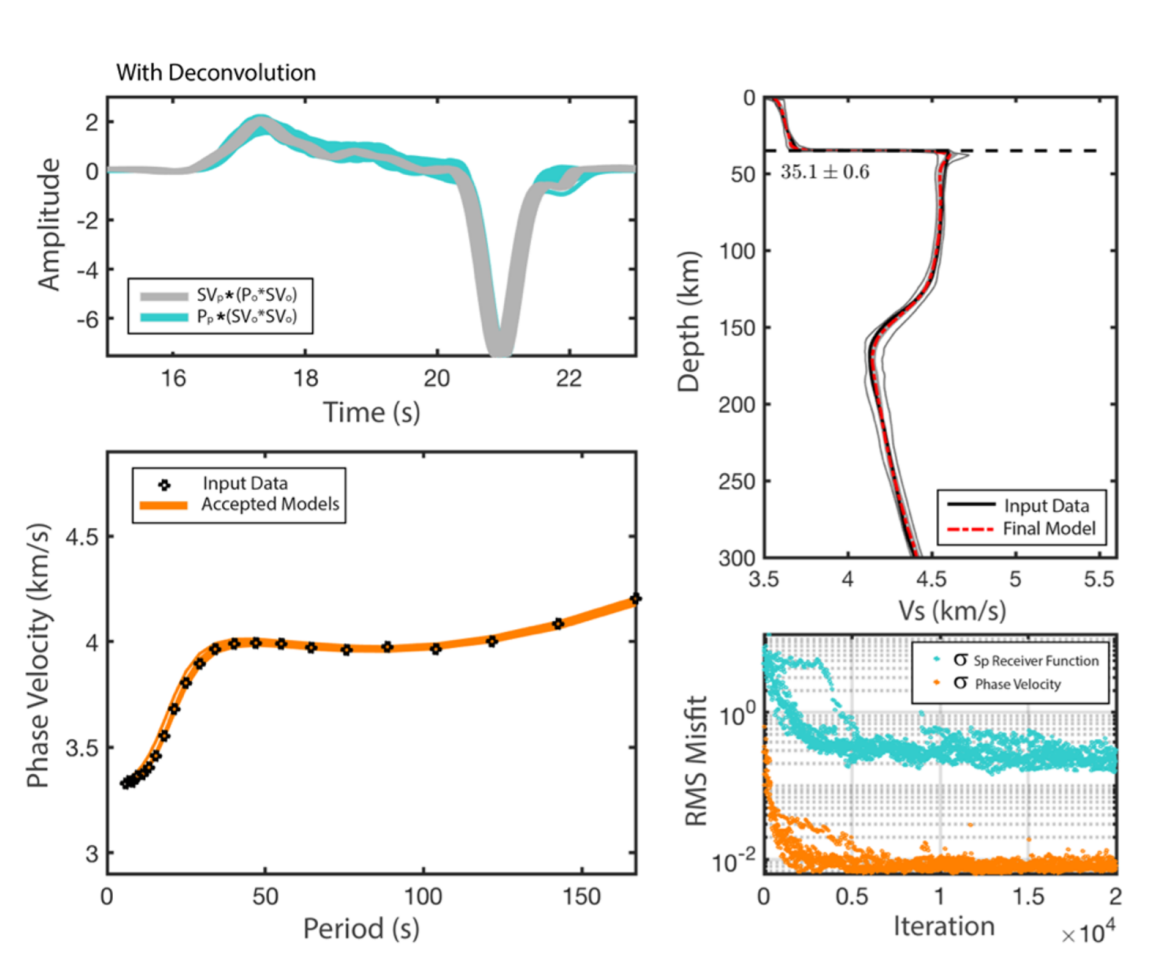


Figure A.2: Results of inversion with surface wave phase velocities and S_p receiver functions with deconvolution. Top left: cross convolution of synthetic data (S_p receiver functions, P_o*SV_o , and auto-deconvolved SV phases, SV_o*SV_o) and body wave predictions (P_p and SV_p) for all accepted models for the last chain. Bold asterisk (*) indicates convolution and asterisk (*) deconvolution. Bottom left: input phase velocity and phase velocity model curves of all accepted models of the last chain. Top right: 1D velocity model of the input data in black and average model as the final model in red. Bottom right: RMS misfits for both data types of all chains throughout the inversion.

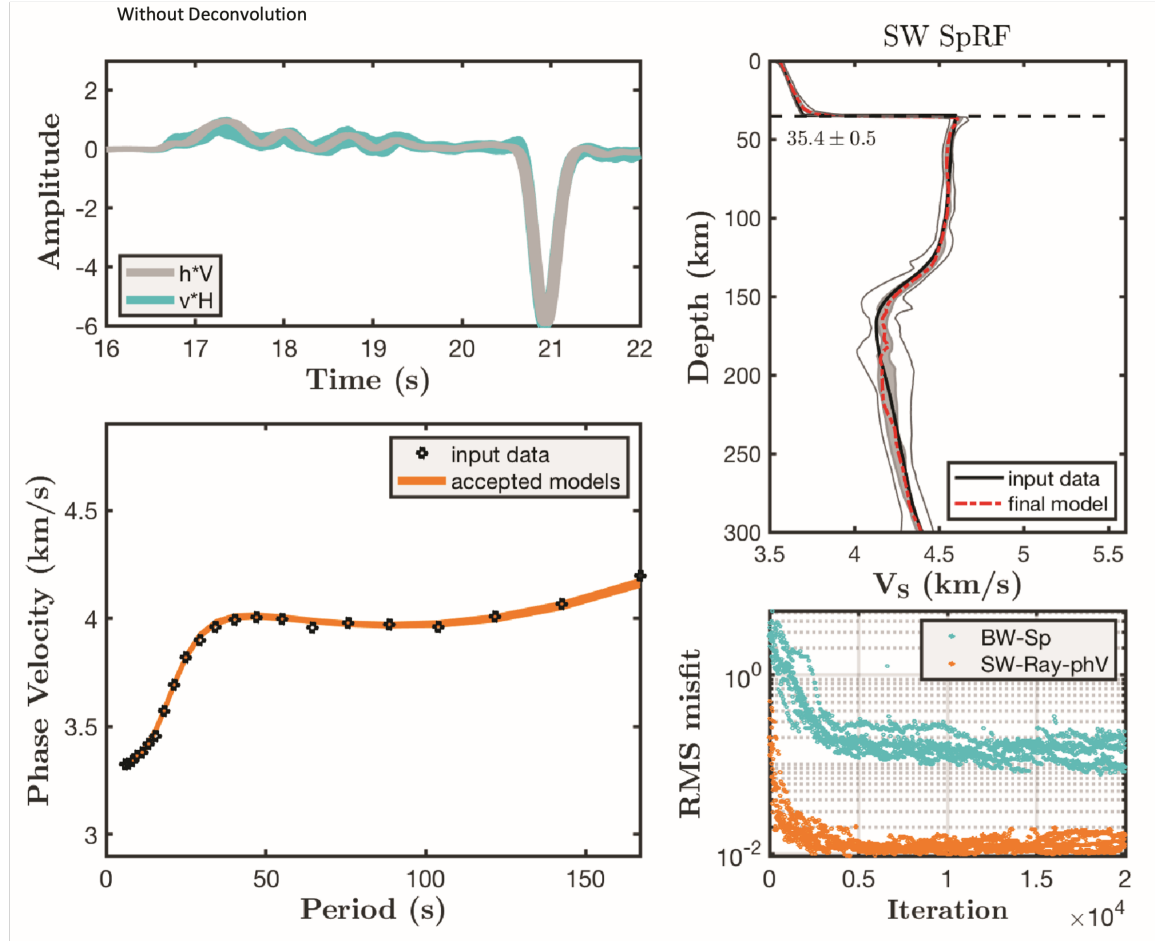


Figure A.3: Results of inversion with surface waves and Sp without deconvolution. Top left: cross convolution of body wave data and accepted models for the last chain. Bottom left: input phase velocity and phase velocity curves of all accepted models of the last chain. Top right: 1D velocity model of the input data and average model as the final model. Bottom right: RMS misfits for both data types of all chains throughout the inversion.

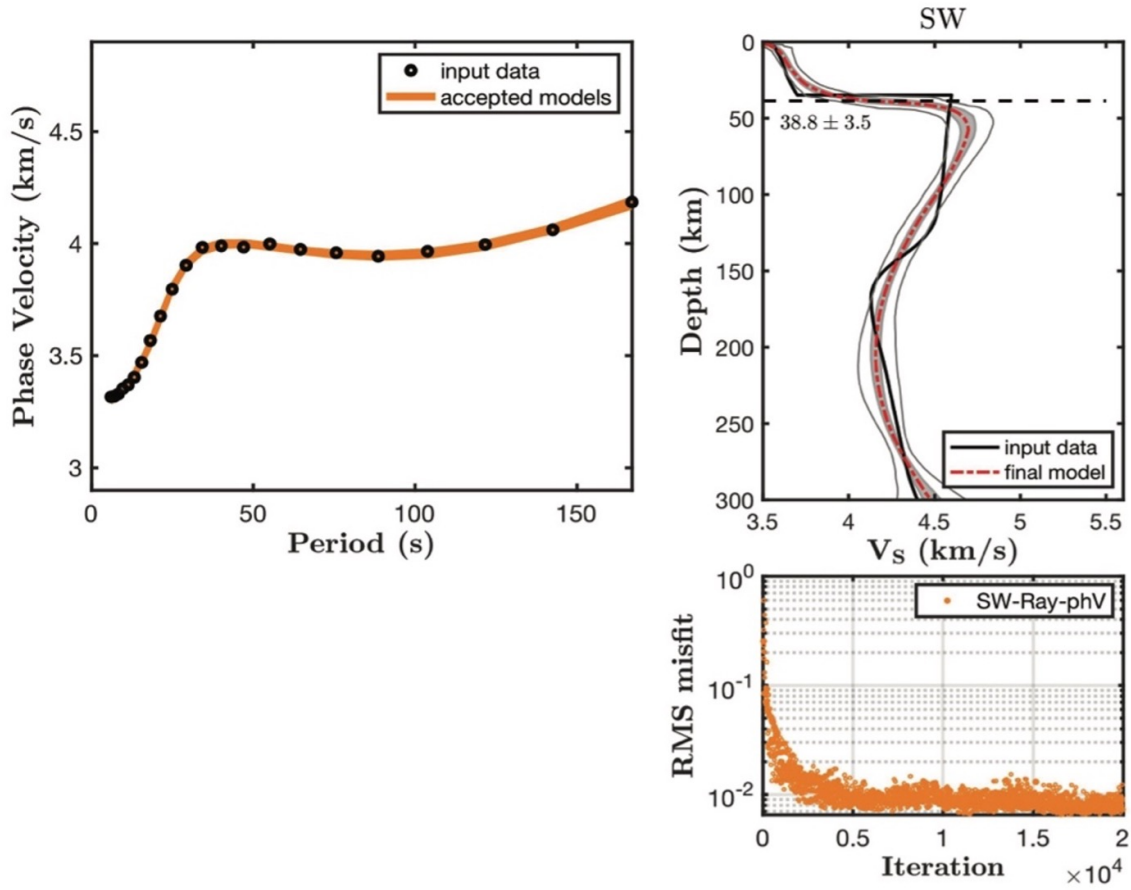


Figure A.4: Results of inversion with surface waves only. Top left: input phase velocity and phase velocity curves of all accepted models of the last chain. Top right: 1D velocity model of the input data and average 19 model as the final model. Bottom right: RMS misfits for surface waves of all chains throughout the inversion.

A.3 Shear Velocity Structure near the Wrangell Volcanic Field

Debate also exists regarding the origins of the Wrangell volcanic field (WVF) (Fig. 2.1a). In this paper we have adopted the interpretation that subduction is occurring beneath the WVF, based on new seismicity contours, which clearly show slab seismicity that deepens to the northeast (Daly, et al. 2019) (Fig. 2.1 and 2.7d). Earlier studies have invoked subduction beneath the WVF, although not necessarily with similar slab contours (e.g., Zhao et al., 1995; Bauer, et al., 2014; Wang and Tape, 2014; Jiang et al., 2018; Ward and Lin, 2018; Feng and Ritzwoller, 2019; Pavlis et al., 2019; Gou et al., 2019). Others have concluded that the WVF volcanism is not due to subduction directly beneath the WVF, but rather is influenced by the edge of the subducted Yakutat plate (Jadamec et al., 2013; Martin-Short et al., 2018; Berg et al., 2020). The results in this study are consistent with the inferred location of the Wrangell slab (Daly et al., 2019), but they do not sample close enough to the WVF to shed much light on this debate. One interesting point is that the upper plate lithosphere is relatively thick (~ 90 km) within approximately 100 km of horizontal distance of the WVF (Fig. 2.7d). Another is that upward asthenospheric flow around the eastern edge of the Yakutat slab could supply heat, and possibly small amounts of partial melt, to the upper plate, providing an explanation for the group of anomalously low lithospheric velocities in this area (Fig. 2.6c).

(a) Parameter	(b) Layer	(c) Variable	Min.	Max.
1. Node Velocity (3/8)	Crust (3/8)	Spline Velocity (3/5)	3.3 km/s	4.2 km/s
		Vp/Vs (1/5)	1.62	1.9
		Knot location (1/5)	0 km	Z_{max}
	Mantle (5/8)	Spline Velocity (4/5)	3.8 km/s	4.9 km/s
		Knot location (1/5)	Zmoho	200 km
2. Moho (2/8)		Velocity Jump (1/3)	-	25%
		Average Velocity (1/3)	Vs bounds	
		Moho Depth (1/3)	$\gamma - 10km$	$\gamma + 10km$
3. Nodes (2/8)	Crust (1/5)	Add (1/2) Remove (1/2)	2	5
	Mantle (4/5)	Add (1/2) Remove (1/2)	4	12
4. Hyperparameter (1/8)		Surface wave Error (1/2)	$10^{-7}m/s$	20 - 50 m/s
		Sp error (1/2)	$10^{-7}amp.$	-

Table A.1: Inversion parameter perturbation limits. Model perturbation starts by selecting which (a) parameter, then (b) layer, (c) variable that can vary within minimum maximum values. In parenthesis are the perturbation probabilities for each of (a), (b), (c). Zmoho represents the Moho depth in the calculated model. Bounds on Zmoho are $\gamma \pm 10$ km, where γ are depths from Zhang et al (2019). For stations that were not included in Zhang et al. (2019) Moho bounds are 20 km to 40 km. Consequently, Z_{max} varies across stations.

Station	Moho	M. err.	LAB	LAB err.	Asth. Vel.	Asth. err.	Lith. Vel	Lith. err.
(km)	(km)	(km)	(km)	(km/s)	(km/s)	(km/s)	(km/s)	(km/s)
AK.ANM	30.6	5.7	72.0	14.0	4.302	0.024	4.271	0.104
AK.BMR	37.7	4.8	101.0	34.0	4.347	0.041	4.574	0.073
AK.BPAW	24.3	1.2	70.0	22.0	4.346	0.035	4.461	0.019
AK.BRLK	35.1	1.2	91.0	24.0	4.479	0.042	4.458	0.111
AK.CAST	38.8	2.3	63.5	20.0	4.332	0.028	4.462	0.057
AK.CCB	28.9	2.5	45.0	6.0	4.291	0.031	4.358	0.072
AK.CHUM	26.3	2.3	72.0	34.0	4.318	0.030	4.416	0.010
AK.CNP	42.8	1.5	93.0	28.0	4.476	0.044	4.576	0.041
AK.COLD	32.7	0.7	118.0	18.0	4.343	0.032	4.523	0.098
AK.CRQ	31.9	1.0	119.0	18.0	4.348	0.047	4.361	0.204
AK.DHY	34.6	0.9	56.0	16.0	4.264	0.048	4.525	0.074
AK.DIV	34.8	1.1	79.0	10.0	4.405	0.031	4.447	0.101
AK.DOT	29.5	1.4	89.0	20.0	4.217	0.017	4.227	0.044
AK.GCSA	31.0	1.9	91.0	34.0	4.183	0.045	4.470	0.042
AK.KTH	30.2	2.3	48.0	44.0	4.313	0.031	4.352	0.065
AK.MDM	31.2	1.8	47.0	6.0	4.341	0.053	4.556	0.222
AK.MLY	26.1	1.9	91.0	6.0	4.200	0.062	4.445	0.088
AK.RDOG	31.4	0.3	120.0	10.0	4.433	0.044	4.528	0.017
AK.SCRK	30.0	2.3	81.0	18.0	4.325	0.024	4.291	0.015
AK.SII	26.0	1.8	98.0	16.0	4.427	0.050	4.381	0.198
AK.SKN	33.4	2.2	52.0	18.0	4.181	0.125	4.480	0.023
AK.SCM	39.6	0.5	98.0	8.0	4.332	0.032	4.443	0.026
AK.TNA	27.8	1.4	74.0	32.0	4.363	0.020	4.443	0.027
AK.WAX	23.4	0.9	121.0	26.0	4.328	0.062	4.346	0.163
AK.FYU	28.0	0.6	73.0	22.0	4.355	0.023	4.440	0.026
AK.WAT6	33.8	1.0	52.0	6.0	4.367	0.028	4.451	0.081
AK.EYAK	31.5	0.8	122.0	12.0	4.375	0.053	4.511	0.034
AK.SUCK	23.5	1.7	116.0	24.0	4.454	0.028	4.330	0.131
TA.D19K	34.0	1.6	132.0	18.0	4.591	0.054	4.590	0.021
US.EGAK	32.4	2.3	60.0	16.0	4.372	0.044	4.624	0.067
IU.COLA	27.0	2.2	57.0	20.0	4.386	0.029	4.414	0.014

Table A.2: Inversion Results. Moho column corresponds to Moho depth (km); M.err. are 2σ Moho depth uncertainties (km); LAB values are LAB depths (km); LAB. err. are 2σ LAB depth errors (km); Asth. Vel. and Lithos. Vel. are asthenospheric and lithospheric velocities (km/s); A. err. are the average of right and left bounds of 2σ asthenospheric velocity error (km/s); L. err. are average of right and left bounds of 2σ lithospheric velocity error (km/s).

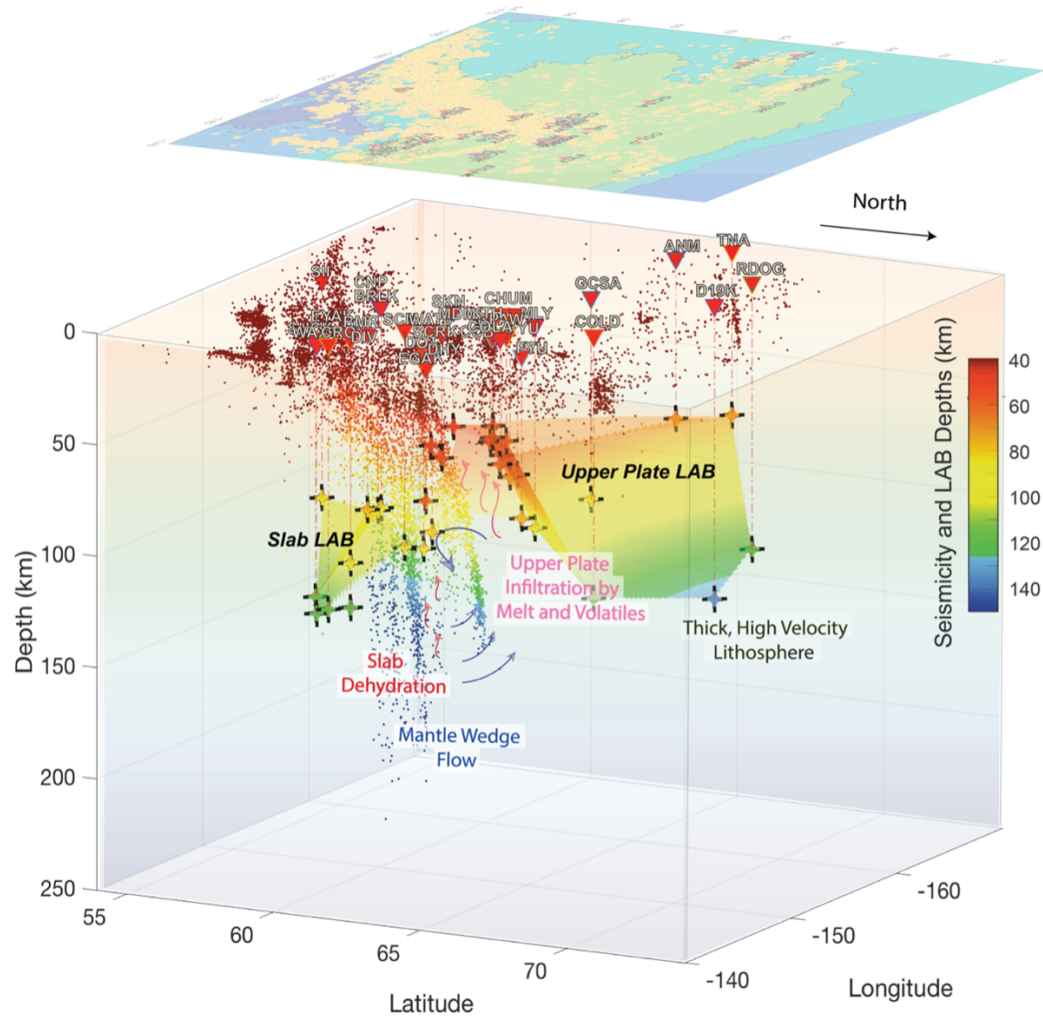


Figure A.5: 3D illustration of all the LAB depths calculated for this paper with annotation. Crosses show LB depth estimates, color-coded by depth. Colored surfaces correspond to interpolated LAB interfaces for the upper (North American) plate and the subducting plate. Small colored dots represent all earthquakes > 35 km depth scaled to be with the same color bar as LAB depths.

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Appendix B

Seismic Velocity Discontinuities beneath Alaska from Sp Phases

Supplementary Material for: *Isabella Gama, Karen M. Fischer, Junlin Hua. **Mapping the Lithosphere and Asthenosphere beneath Alaska with Sp Converted Waves**, Geochemistry, Geophysics, Geosystems.*

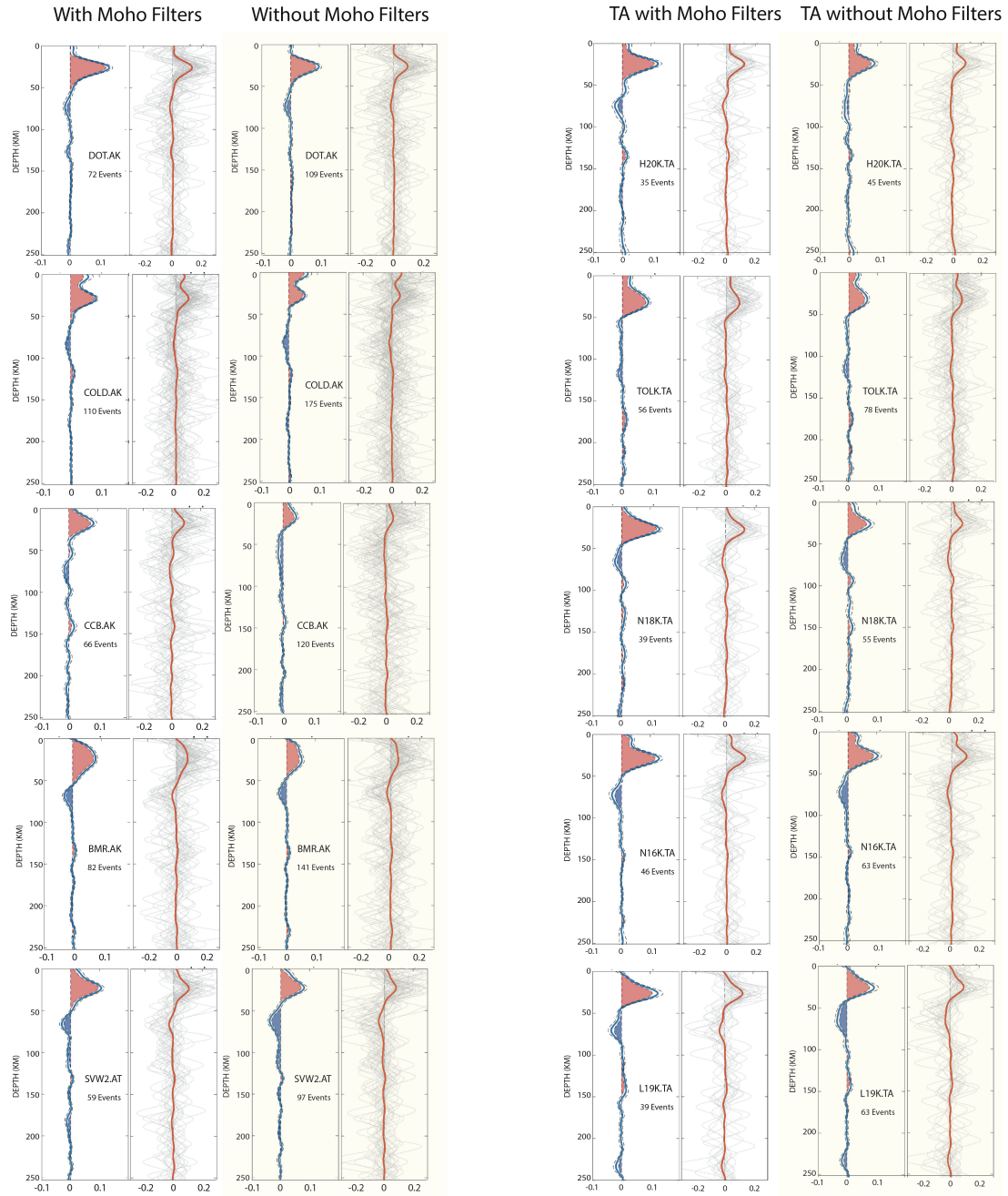


Figure B.1: Single stack S_p phases with and without Moho filters. Left column shows examples with permanent stations and the right column shows examples with transportable array station.

Appendix C

Mantle Signatures of Terrane and Fault Zones in Alaska

Supplementary Material for: *Isabella Gama and Karen M. Fischer. **Mantle Signatures of Terranes and Fault Zones and Implications for Deformation in Alaska**, Nature Geoscience.*

C.1 Online Supplement

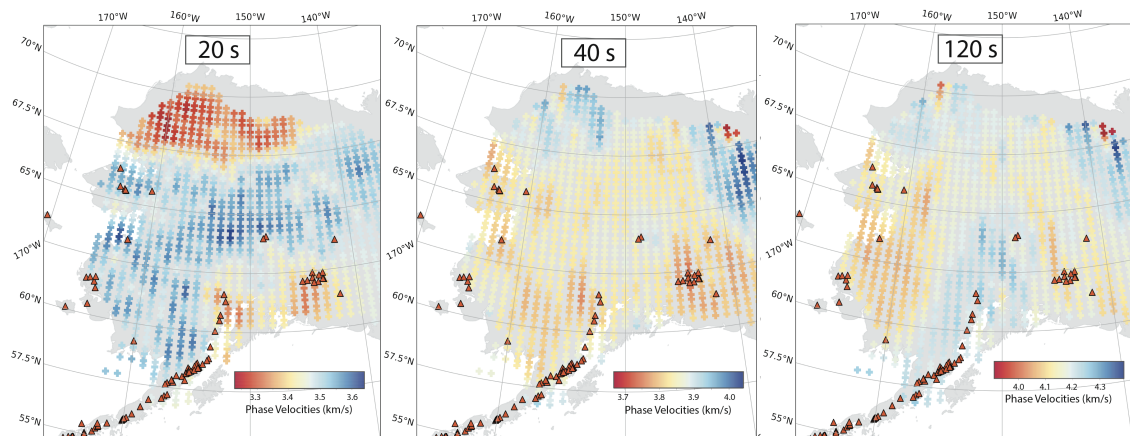


Figure C.1: Phase velocity maps of periods of 20, 40, and 120 s.

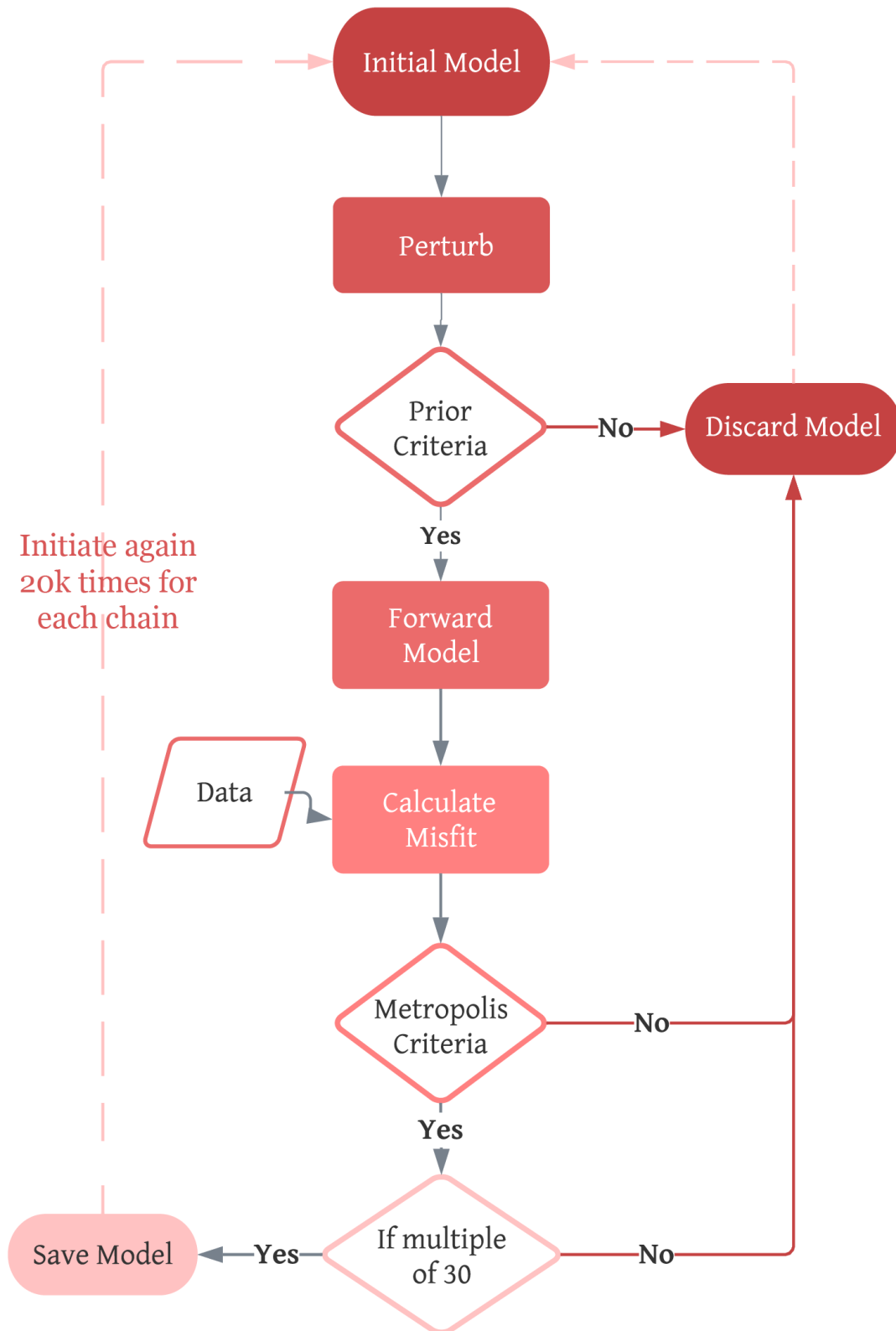


Figure C.2: Inversion workflow.

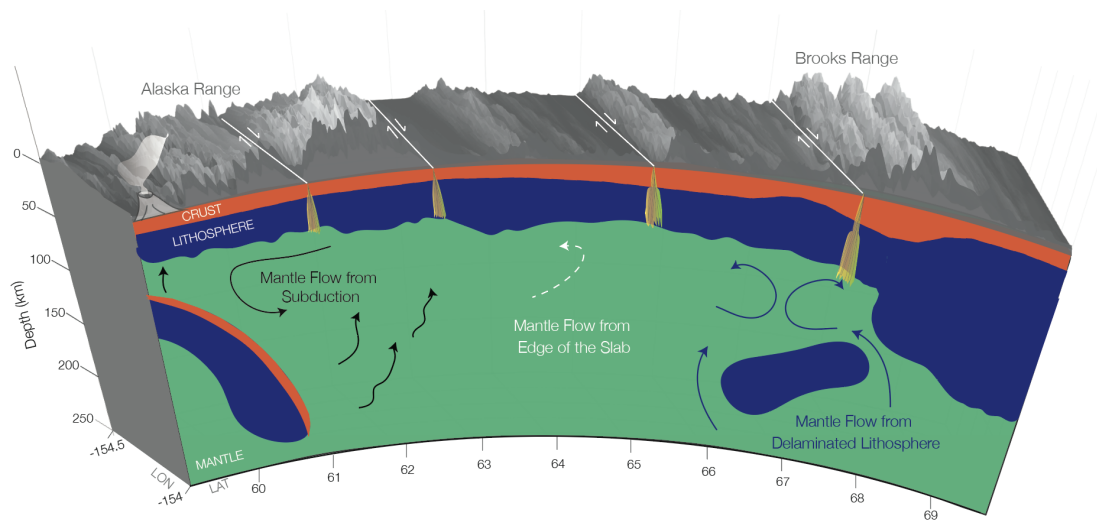


Figure C.3: 3D interpretation of a S-N profile across Alaska.

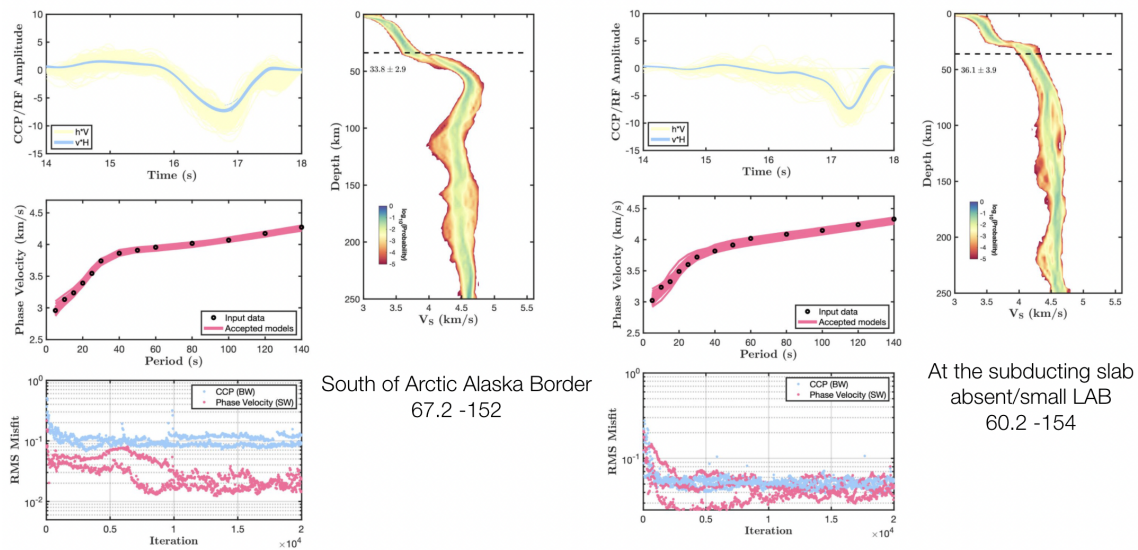


Figure C.4: Results of inversion showing examples of two distinct models at the coordinates shown below annotation. Top left: cross convolution of CCP data and accepted models for all chains after burn in period. Middle left: input phase velocity and phase velocity curves of all accepted models of all chains after burn in period. Bottom left: RMS misfits for both data types of all accepted chains throughout the inversion. Top right: probability of the resulting velocity models after burn in period.

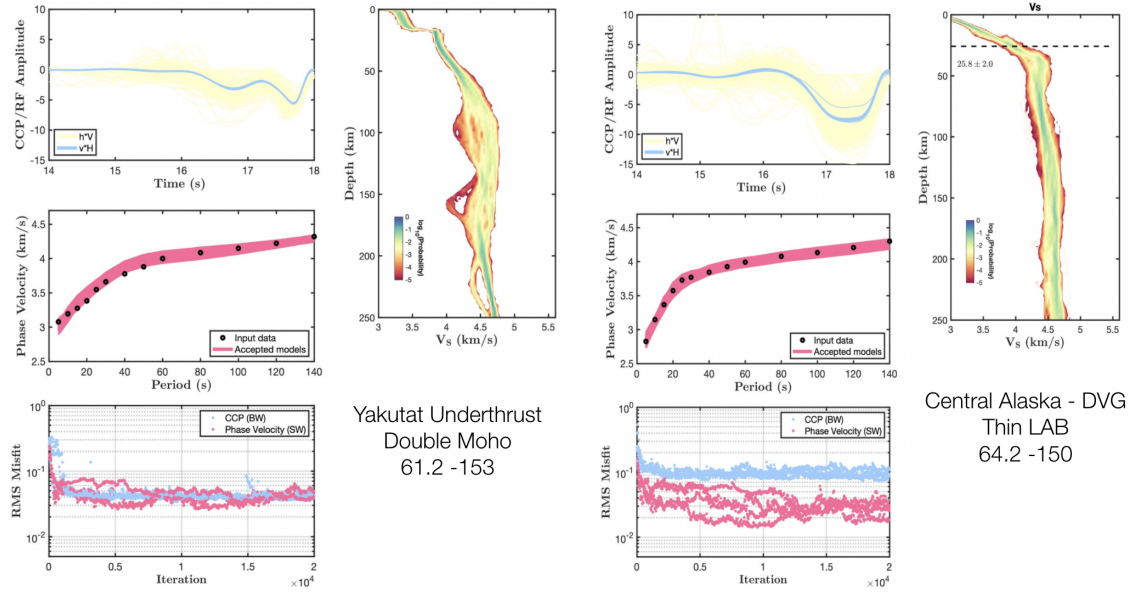


Figure C.5: Results of inversion showing examples of two distinct models at the coordinates shown below annotation. Top left: cross convolution of CCP data and accepted models for all chains after burn in period. Middle left: input phase velocity and phase velocity curves of all accepted models of all chains after burn in period. Bottom left: RMS misfits for both data types of all accepted chains throughout the inversion. Top right: probability of the resulting velocity models after burn in period.

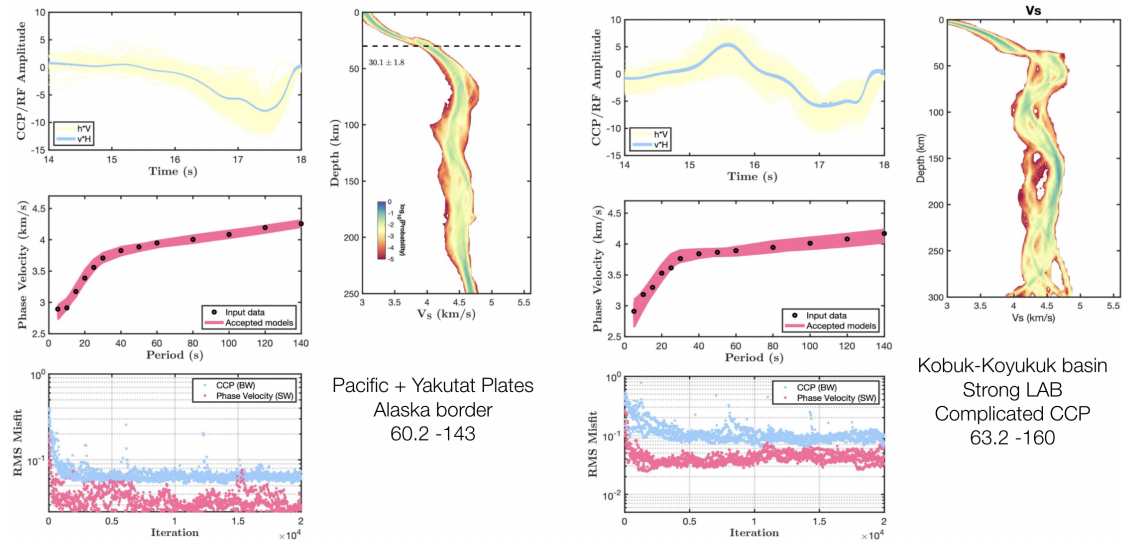


Figure C.6: Results of inversion showing examples of two distinct models at the coordinates shown below annotation. Top left: cross convolution of CCP data and accepted models for all chains after burn in period. Middle left: input phase velocity and phase velocity curves of all accepted models of all chains after burn in period. Bottom left: RMS misfits for both data types of all accepted chains throughout the inversion. Top right: probability of the resulting velocity models after burn in period.