

Exploring Complexities in the Earth's Upper Mantle and Connecting Approaches to Inclusivity in the Earth Sciences

by

Hannah E. Krueger

M.Sc. Brown University, 2019

B.Sc. Appalachian State University, 2017

A dissertation submitted in partial fulfillment of the requirements for the degree of Doctor of
Philosophy in the Department of Earth, Environmental and Planetary Sciences at Brown
University

Providence, Rhode Island

October 2023

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The dissertation by Hannah E. Krueger is accepted in its present form
by the Department of Earth, Environmental, and Planetary Sciences as
satisfying the dissertation requirement for the degree of Doctor of Philosophy.

Date 07/31/2023

Karen M. Fischer

Karen Fischer, Advisor

Recommended to the Graduate Council

Date 07/31/2023

Josh Russell

Josh Russell, External Reader

Date 7/31/2023

Colleen A Dalton

Colleen Dalton, Reader

Date 07/31/2023

Alberto Saal

Alberto Saal, Reader

Date 07/31/2023

Harriet Lau

Harriet Lau, Reader

Approved by the Graduate Council

Date _____

Thomas A. Lewis,
Dean of the Graduate School

Hannah E. Krueger

Curriculum Vitae

EDUCATION

Brown University, Providence, RI

Ph.D., Earth, Environmental, and Planetary Science – Anticipated August 2023

M.S., Earth, Environmental, and Planetary Science – May 2019

Appalachian State University, Boone, NC

B.S., Geology with Honors – August 2017, Summa Cum Laude

RESEARCH EXPERIENCE

Seismic Attenuation of the Antarctic lithosphere January 2023 – Ongoing

Dr. Karen Fischer & Dr. Colleen Dalton, **Brown University**, Providence, RI

Joint inversion of geophysical data for cratonic lithosphere structure October 2021 – Ongoing

Dr. Karen Fischer, **Brown University**, Providence, RI

Investigation of cratonic lithospheric structure using body wave RFs Fall 2017 – 2021

Dr. Karen Fischer, **Brown University**, Providence, RI

Investigating Along-Strike Anisotropy in Cascadia using RFs Summer 2015 – 2017

Incorporated Research Institutions for Seismology (IRIS) Internship

Dr. Erin Wirth, **University of Washington**, Seattle, WA

GPS Time Series Analysis of Seasonal Deformation in Southern CA Fall 2013 – 2017

Dr. Scott Marshall, **Appalachian State University**, Boone, NC

HONORS, AWARDS, & FELLOWSHIPS

Deans' Faculty Fellowship, Brown University May 2021

One-year visiting professorship in the Department of Earth, Env., and Planetary Sciences.

Presidential Award for Excellence in Teaching, Brown University May 2021

Recognition of excellence as a graduate teaching assistant and outreach coordinator.

Graduate Research Fellowship, National Science Foundation (NSF) Spring 2018 – 2023

Graduate research funding and support.

Quantitative Geoscience Award, Appalachian State University Spring 2017

High performance in quantitative sciences courses.

Chancellor's Scholarship, Honors College, Appalachian State University May 2013

Full-institutional cost undergraduate merit scholarship.

DIVERSITY, EQUITY, & INCLUSION

American Geophysical Union Fall Meeting 2020, 2021, 2022 (Co-created & Chaired)

Education Session - Inclusivity in Geosciences: A conversation between scientists and educators

Organized & led a diverse panel of educators with the goal of improving inclusivity in geoscience education.

THESES

Krueger, H. E. 2019. Do Mid-Lithospheric Discontinuities Exist? Master's Thesis, Brown University, Providence RI. Advisor: Karen M. Fischer

Krueger, H. E. 2017. Identification of secular velocity contamination using seasonal GPS signals Undergrad. Honors Thesis, Appalachian State Univ., Boone, NC. Advisor: Dr. Scott T. Marshall

PUBLICATIONS

Krueger, H. E., Dalton, C.A., Russell, J, Fischer K.M. *In Prep*. Investigating the complex tectonic history of Antarctica with phase velocity and attenuation.

Krueger, H. E., Gama, I., Fischer K.M., Golos, E.M. *In Prep*. A common origin for layering in the cratonic lithosphere.

Krueger, H. E., Garelick, S. *In Prep*. A comprehensive model for Inclusivity in the Geosciences: Reflecting on three years of AGU Fall Meeting panel sessions.

Garelick, S., Krueger, H. E. *In Prep*. Weather & Climate Next Generation Science Standards curriculum for third-grade classrooms in Providence, RI.

Krueger, H. E., Gama, I., Fischer K.M. 2021. Global patterns in cratonic mid-lithospheric discontinuities from Sp receiver functions. *Geochemistry, Geophysics, Geosystems*. DOI: 10.1029/2021GC009819.

Gama, I., Fischer, K. M., Eilon, Z., Krueger, H. E., Dalton, C. A., Flesch, L. M. 2020. 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*. DOI: 10.1016/j.epsl.2021.116785.

Krueger, H. E., Wirth, E. A. 2017. Investigating segmentation in Cascadia: Anisotropic crustal structure and mantle wedge serpentinization from receiver functions. *Geochemistry, Geophysics, Geosystems*. DOI: 10.1002/2017GC007064

Marshall, S. T., Funning, G. J., Krueger, H. E., Owen, S. E., Loveless, J. P. 2017. Mechanical models favor a ramp geometry for the Ventura-Pitas Point fault, California. *Geophysical Research Letters*; 44(3):1311-1319. DOI: 10.1002/2016GL072289.

TEACHING

Brown University, Department of Earth, Environmental, & Planetary Sciences, Providence, RI
Lecture: *Tectonic plate driving forces & mountain building* November 2022
Designed and taught an interactive & inquiry-driven 1.5-hour lecture for a 70-student Earth & Environmental Processes course.

Teaching Assistant: *Understanding Earth & Environmental Processes* Fall 2022
Co-taught labs, co-led day-long fieldtrip exercise, managed undergrad. TAs, and graded coursework.

Teaching Assistant: *Intro. to Methods in Data Analysis* Fall 2020
Assisted in remote teaching format, office hours, led weekly recitation, and graded coursework.

Graduate Assistant: *Undergraduate Spring Field Trip* Spring 2019
Provided geology & general assistance for the undergrad-run field trip to Arizona.

Brown University, Sheridan Teaching Center, Providence, RI

Teaching Certificate I

Fall 2018

Participated in a workshop-style, semester-long program for teaching practices in university settings.

Appalachian State University, Department of Geology, Boone, NC

Teaching Assistant: Intro. to Geophysics Lab

Spring 2017

Assisted students in quantitative geophysics laboratory.

OUTREACH & LEADERSHIP

Brown University, Department of Earth, Environmental, & Planetary Sciences, Providence, RI

DEEPS CORES Curriculum Developer

Spring 2021 – ongoing

Developed interactive curriculum to introduce ES topics & careers in classroom at Hope HS, Providence, RI.

DEEPS STEP 3rd Grade Coordinator & Volunteer Teacher

Fall 2018 – Summer 2022

Developed, implemented, & co-taught lesson plans for 3rd grade science. Module topic: Weather & Climate.

Graduate Student Summer Field Trip Leader

Fall 2018

Planned, organized, and led a 5-day trip for incoming graduate students to Vermont.

PROFESSIONAL ACTIVITIES

Teaching at Teaching Intensive Univ. Workshop, Westfield State Univ., MA

September 2022

Overview of teaching & scholarship at PUIs & community colleges.

Remote Online Sessions for Emerging Seismologists (ROSES)

Summer 2020

11-week workshop on using Python for seismology research.

FIELD EXPERIENCE

MUSICA project

Bogotá, Colombia – February 2023, 3 weeks

PI: Dr. Lara Wagner, Carnegie Science

Summer of Applied Geophysical Exp. (SAGE)

Santa Fe, NM – Summer 2016, 4 weeks

Prof: Dr. John Ferguson, U of Texas – Dallas

Field Geophysics, U of South Florida

Snake River Plain, ID – Summer 2017, 2 weeks

Prof: Dr. Jochen Braunmiller

Inv. Magma Under St. Helens (iMUSH)

Southern Washington – Summer 2015, 2 weeks

PI: Dr. Geoff Abers, Cornell University

SKILLS

Programming Languages:

MATLAB – Primary language used in graduate research. Retrieve, organize, process, & plot data.

Generic Mapping Tools (GMT) – Plot maps, DEMs, contour data, time-series, x-y plots, etc.

Perl – Primary language in undergrad. research. Parse ASCII data files.

Linux Shell Scripting (Bash) – Navigating the terminal, executing scripts, troubleshooting, etc.

Python – Primary language used in ML research project, courses, & ROSES workshop.

CONFERENCE PRESENTATIONS

1st authorship only

- Krueger, H. E., Gama, I., Fischer, K. M., Golos, E.M. 2022. Bayesian joint inversion of Sp receiver functions and Rayleigh wave phase velocities for the structure of the cratonic upper mantle. *American Geophysical Union Fall Meeting, Chicago, IL*. Poster presentation.
- Krueger, H. E., Gama, I., Fischer, K. M. 2021. Analyzing Global Patterns in Cratonic Mid-Lithospheric Discontinuities. *American Geophysical Union Fall Meeting, New Orleans, LA*. Oral presentation.
- Krueger, H. E., Fischer, K. M., Gama, I. 2020. Patterns in the Mantle Discontinuity Structure of the Cratonic Lithosphere. *American Geophysical Union Fall Meeting, Remote*. Oral presentation.
- Krueger, H. E., Fischer, K. M., Gama, I., Hua, J., Eilon, Z. 2019. Do Mid-Lithospheric Discontinuities exist? *American Geophysical Union Fall Meeting, San Francisco, CA*. Oral presentation.
- Krueger, H. E., Fischer, K. M., Gama, I., Hua, J., Eilon, Z. 2019. The Distribution of Mid-Lithospheric Discontinuities on Precambrian Continents. *IRIS SAGE/GAGE Workshop, Portland, OR*. Poster presentation.
- Krueger, H. E. 2018. Understanding the Origins of Mid-Lithospheric Discontinuities in the Kalahari Craton. *American Geophysical Union Fall Meeting, Washington D.C.*
- Krueger, H. E., Wirth, E. A., 2016. Investigating Segmentation in Cascadia: Anisotropic Crustal Structure and Mantle Wedge Serpentinization. *American Geophysical Union Fall Meeting, San Francisco, CA*.
- Krueger, H. E., Marshall, S. T., 2016. Seasonal Motions and Interseismic Strain Measured by Continuous GPS throughout the Transverse Ranges, CA. *Southern California Earthquake Center Annual Meeting, Palm Springs, CA*.
- Krueger, H. E., Wirth, E. A., 2015. Crustal Anisotropy in the Cascadia Subduction Zone – Evidence from Teleseismic Receiver Functions. *American Geophysical Union Fall Meeting, San Francisco, CA*.

Acknowledgements

First and foremost, I want to acknowledge my incredible dog, Maple. She has been by my side throughout my entire PhD; so, I believe she has also fulfilled the degree requirements. Thank you so much for your faithfulness and love, my **P**henomenal **D**og, Maple, PhD.

The first *person* I would like to acknowledge is my advisor, Karen. She is steadfast in her research ethics and personal values. It is truly invaluable to have worked with an advisor that not only provides excellent advice for research and teaching, but also cares about the emotional and mental health of her students. I could not imagine having a graduate experience with a more supportive advisor. At Brown I also relied on many other mentors. Notably, Colleen who co-advising my third thesis chapter and served as a constant voice of kindness and curiosity on my committee; Victor who provided an example of innovative teaching despite challenging conditions; and Alberto who always brought new ideas to our committee meetings.

I would not have pursued geophysics if it had not been for the incredible academic mentors I had prior to graduate school, Lara Wagner, Scott Marshall, and Erin Wirth. I am so thankful for their patience during my early days of research, and continued support.

Furthermore, the meaningful work I have done through outreach has fulfilled my graduate experience. In this work, Olga has served as an incredible mentor, Sloane has been wonderful collaborator, and the teachers and students in Providence Public Schools have provided an incredible environment for me to grow as an educator.

I also am grateful for my peers that have supported me like family. Most notably, I thank my “seismology twin,” Isabella Gama Dantas, for supporting me as a friend and lab mate through many highs, lows, and major life changes. Also, I thank Majid and Anant for always being excited to discuss science, and answer questions with no judgment. I also recognize all the friends that have helped me stay sane through spin class, wine nights, rock climbing, potlucks, hiking, and dancing. As well as my best friends from North Carolina, Laura and Kristen.

I greatly appreciate those that supported me during my year-long leave. At Brown, and in my family, I had immense support for my journey and recovery. I am thankful for Warren Doyle and the many other “trail angels” that directly supported my (and Maple’s) walk along the Appalachian Trail. Without their support, it is unlikely I would have continued in this program and made it to my defense. During this time, I met my amazing partner, Seany, whose care for those around him never ceases to amaze me. The list of things I thank him for include, but are not limited to, daily 7 am wakeup calls while working on my dissertation, spoiling Maple, reliably making goofy faces at little kids, and encouraging me to do the things I love.

My family has been a grounding force for me throughout this process. In particular, I am grateful for my mother, who is endlessly supportive but always encourages me to prioritize my own wellbeing; my brother, Isaac, for including me in his life as he has jumped fearlessly into adulthood; Grandma and Grandpa Krueger, who always show up to support me, even though I am only one of their twelve grandchildren, and a 10-hour drive away; Mommom for always providing a loving home full of quality time, even for months at a time; and all my extended family for reminding me how incredibly interesting, diverse, and loving the world is.

Finally, I remember my father who instilled in me grit, self-reflection, a love for the outdoors, the practice of questioning the world around me – and, unfortunately, a little bit of procrastination.

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Introduction

Upper mantle structure

The lithosphere-asthenosphere system

The Earth's upper mantle consists of two rheological layers: the asthenosphere and the lithosphere. The asthenosphere is characterized by low seismic velocities and convective heat transport. Overlying the asthenosphere is the lithosphere, which serves as a brittle/elastic "lid" atop the asthenosphere. The lithosphere is generally characterized with high seismic velocities, an iron-depleted composition, and is dominated by conductive heat transport (e.g. Fischer et al., 2020; Lee et al., 2011). The interface of these two layers is referred to as the lithosphere-asthenosphere boundary, or the LAB. The LAB at a regional scale often varies in complex ways and continues to be a feature of interest in geophysical studies (e.g. Gama et al., 2021; Lloyd et al., 2020). Generally, the depth of this interface correlates with the amount of time that has passed since evidence of tectonic activity near Earth's surface, or the thermotectonic age (e.g. Artemieva, 2006; Fischer et al., 2020). For example, the LAB is relatively deep (≥ 200 km) in tectonically stable East Antarctica, but shallow (~ 60 km) in the tectonically active West Antarctica (Lloyd et al., 2020). Improving our understanding of the complex lithosphere-asthenosphere system works towards unveiling the processes that form and modify continents.

Antarctica as a case study

Antarctica serves as an incredible case study for the complexities of upper mantle structure in the presence of various tectonic regimes. Antarctica is essentially divided into two parts: stable East Antarctica and active West Antarctica. The juxtaposition of these terranes has resulted in dynamic alterations to the lithosphere at their boundary in the Transantarctic Mountains (Shen et al., 2018). West Antarctica is an amalgamation of terranes with various histories. Notably, over

the past 250 million years, West Antarctica is thought to have experienced extensive rifting, widespread volcanism, subduction, the formation of a slab window at the cessation of subduction, and mantle plume interactions (Jordan et al., 2020; Wiens et al., 2023).

In addition to its intriguing geophysical structure, the study of the Antarctic upper mantle also has important implications for sea level rise. Mantle temperatures and viscosity control the rates of melting at the base of glaciers and isostatic rebound after the removal of ice (van der Wal et al., 2015; Wiens et al., 2023). Despite this importance, there are few samples of Antarctic rocks (e.g. Goodge, 2007) because of ice cover and general inaccessibility, so the use of geophysical techniques are incredibly valuable (Wiens et al., 2023 & references therein).

The cratonic upper mantle

While this thesis does explore the upper mantle in both tectonically active and stable continental regions, a majority of my graduate research has focused on the cratonic lithosphere. Therefore, below I provide a more in-depth contextualization of topics regarding the cratonic lithosphere.

Cratons are often generalized, as we did above, to be regions that have been tectonically stable since the Archean or Paleoproterozoic and have thick, iron-depleted, cold lithospheres that exhibit fast seismic velocities to depths of ≥ 150 km. These generalizations about the age, seismic character, and composition of the cratonic lithosphere commonly coincide, but can conflict. A clear conflict is the observation of layering in seismic velocity, particularly, layering that indicates a low seismic velocities within the cratonic upper mantle (e.g. Abt et al., 2010; Eilon et al., 2018; Ford et al., 2010; Geissler et al., 2010; Hansen et al., 2015; Hopper & Fischer, 2015; Hopper & Fischer, 2018; Kind & Yuan, 2018; Kind et al., 2012; Liu et al., 2021; Liu & Shearer, 2021; Miller & Eaton, 2010; Rychert & Shearer, 2009; Selway et al., 2015; Sun & Kennett, 2017; Wirth & Long, 2014; Wolbern et al., 2012).

One possibility for the origin of these observations is the modification of cratons from underlying or adjacent fluid infiltration. This can introduce slow seismic velocities through the crystallization of hydrous compositions despite little-to-no alteration of the apparent thermotectonic age measured at the surface. These processes may erode the LAB and/or cause compositional layering within the lithosphere (Jackson & Gibson, 2023; Sun & Kennett, 2017; Wolbern et al., 2012). Adding to the apparent complexity of the cratonic mantle is the observation of high seismic velocities in the lower lithosphere of cratons that exceeds the range of possible peridotite shear velocities at cratonic geotherms (e.g. Dalton et al., 2017; Dalton & Faul, 2010; French & Romanowicz, 2014; Garber et al., 2018; Lekic & Romanowicz, 2011; Schaeffer & Lebedev, 2013).

Understanding layering in the cratonic lithosphere is a collaborative effort that joins seismic and xenolith observations (Aulbach et al., 2020; Garber et al., 2018; Rader et al., 2015; Selway et al., 2015) with geochemical and geodynamic experiments and models (e.g. Fu et al., 2022; Saha et al., 2021; Yang et al., 2023). This work is done to illuminate the processes that formed and evolved cratons (Lee et al., 2011).

Seismic methods for studying the upper mantle

Despite this extensive volume of work characterizing cratonic lithosphere, the words “elusive” (Kind et al., 2020) and “ubiquitous” (Fu et al., 2022) are both used to describe the presence of layering, and specifically a low velocity layer, within the cratonic lithosphere. While the intrinsic heterogeneity of mantle structure provides one explanation for inconsistency in observations, another is that there are inherent methodological challenges to studying upper mantle layering.

Here, we emphasize a few key methodologies for studying the upper mantle structure. The methods outlined here are used in this thesis to characterize and investigate the origins of the layering within the cratonic lithospheric mantle.

Converted body waves are used to indicate the velocity gradient structure of the crust and mantle. A depth with a high velocity gradient likely indicates the top or bottom of a layer of anomalous seismic velocity. P-to-S converted body waves can resolve robust layering within the crust and mantle transition zone. Unfortunately, at upper mantle depths, reverberations of seismic energy within the crust can overprint signals of true Earth structure. Alternatively, S-to-P converted body waves do not experience the effects of reverberations, but do present challenges in data quality. Due to these challenges, careful handling of S-to-P data is crucial for robust results (See Chapter 1).

Rayleigh phase velocities provide complimentary observations to converted body waves. While converted body waves resolve discontinuity structure, Rayleigh phase velocities provide the complimentary observation of absolute shear velocities, but as smooth frequency-dependent profiles (Eilon et al., 2018). The joint interpretation or inversion of these data types provides a more holistic view of the upper mantle structure where layering and absolute shear velocity are constrained. Typically, these inversions implement a Bayesian framework to allow for the simple integration of multiple data types for misfit calculation (Eilon et al., 2018; Gama et al., 2022; Gama et al., 2021; Shen et al., 2018). These inversions are limited by the resolution of the Rayleigh phase velocity data set they utilize (See Chapter 2).

While shear velocity provides a powerful tool for interpreting mantle structure and composition (e.g. Altoe et al., 2020; Eeken et al., 2018), it is nonunique in its presentation of the effects of temperature and melt. This is of particular consequence in regions such as Antarctica

where resolution of mantle temperature and viscosity are crucial for understanding the societal impacts of sea level rise (van der Wal et al., 2015). Attenuation is an additional measurement that can be made using phase velocities and amplitude measurements. When analyzed together, the combined interpretation of attenuation and phase velocity allows for the evaluation of the collective influence of composition, temperature, and melt on seismic observations, considering their distinct sensitivities to mantle characteristics (Havlin et al., 2021).

Inclusivity in the geosciences

While geologic investigations often emphasize the Earth's heterogeneity, our field has historically neglected to scrutinize the origins of its own homogeneity. Many studies have emphasized the lack of diversity within the geosciences (Bernard & Cooperdock, 2018). This lack of diversity gained particular attention in the broader geoscience community during the nationwide Black Lives Matter protests in 2020. It is essential to avoid disconnecting discussions of diversity from the underlying culture that perpetuates a lack of diversity (Berhe et al., 2022; Dutt, 2020). To build a field that is not only more diverse but also inclusive and supportive of a wide range of scientists, we need to adopt a comprehensive strategy. This strategy should place emphasis on promoting inclusivity in our engagement with local communities (Medina Luna et al., 2019), educational methods (Bybee et al., 2006), and the overall environment within our departments and research groups. These focus points are not distinct, but intersect in a way that can center student and scientist identities, provide relevant and engaging geoscience experiences, and build symbiotic relationships with the community (Fortner et al., 2022; Gama et al., 2020; Garelick et al., 2021; Krueger et al., 2022). I discuss this strategy in detail in my final chapter.

Organization of thesis

This dissertation consists of three chapters that present scientific research. The first scientific research chapter has been peer reviewed. The second chapter is prepared for submission. The third chapter provides a comprehensive examination of the findings on phase velocity in Antarctica, while offering only initial insights into attenuation. Further efforts are necessary to finalize the attenuation analysis and ensure its suitability for publication.

In my first two chapters, I perform global studies of layering in the cratonic lithosphere. The first uses S-to-P converted body waves to investigate global patterns in lithospheric layering, finding patterns in the occurrence and magnitude of layering with thermotectonic age and proximity to active tectonic regions, respectively. The second chapter uses S-to-P converted body waves jointly with Rayleigh phase velocities to gain a more holistic perspective on the cratonic lithosphere. This work showed that the small velocity gradients we observe in converted body waves can oversimplify the structure of the upper mantle. When we combine observations of velocity gradients with absolute velocity from phase velocities, we see a much clearer picture of cratonic mantle that is heterogeneous at local, regional, and global scales. By resolving layered structure with absolute velocities, we can also analyze compositional explanations for our observations. We find that the presence of a velocity increase in the lower cratonic lithosphere is very common and often has an overlying low velocity layer. This layering can be explained by the presence of 1-3% diamond and 1-3% velocity reductions, respectively. These variations, along with the general heterogeneity we observe, are consistent with extensive infiltration of underlying carbonate melt from mantle upwellings.

In my third chapter, I investigate upper mantle structure in Antarctica at a regional scale using Rayleigh waves. Antarctica provides an excellent continent-wide case study of the Wilson

cycle. There, an extensive mountain chain divides the stable continental lithosphere of East Antarctica and recent tectonic activity of West Antarctica that includes complex subduction and continental rifting (Jordan et al., 2020). Throughout this region Rayleigh phase velocities provide observations of complex crustal and upper mantle structures that emphasize upper mantle heterogeneity. This complexity is further emphasized in preliminary attenuation maps for the region. Despite these complexities in small scale features, our big-picture observations remain consistent with the dichotomy of thick, cold, elastic lithosphere in stable East Antarctica and warm, attenuating upper mantle West Antarctica from recently active tectonics.

The fourth chapter is a commentary on approaches to inclusivity in the geosciences. This article was written co-equally with Dr. Sloane Garelick and has been reviewed by three additional co-authors. In this chapter, a Venn-diagram model for approaching inclusivity at an institutional level in the geosciences is presented. This article presents the “lessons-learned” from three years of organizing, planning, and moderating a panel session at the American Geophysical Union Fall Meeting titled “Inclusivity in the Geosciences: A conversation between educators and scientists” (Gama et al., 2020; Garelick et al., 2021; Krueger et al., 2022).

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Chapter 1 Global Patterns in Cratonic Mid-Lithospheric Discontinuities from Sp Receiver Functions

Hannah E. Krueger ^{1,*}, Isabella Gama ¹, Karen M. Fischer ¹

¹ Department of Earth, Environmental, and Planetary Sciences, Brown University

This chapter was published as:

Krueger, H. E., I. Gama, and K. M. Fischer (2021), **Global Patterns in Cratonic Mid-Lithospheric Discontinuities from Sp Receiver Functions**, *Geochemistry, Geophysics, Geosystems*, e2021GC009819, doi:10.1029/2021GC009819.

Key Points:

- Layering in the cratonic lithosphere is common but not ubiquitous; regions with lower mantle velocities have stronger shear velocity drops
- Layering is more common in regions of Mesoproterozoic age and younger, consistent with the transition from warm-to-cold subduction
- Ancient lithosphere can produce most layering, but some craton margins require post-formation alteration of cratonic mantle lithosphere

Abstract

The goal of this study is to constrain the origins of layering in seismic velocity structure within the cratonic mantle lithosphere (mid-lithospheric discontinuities or MLDs). For long-lived stations in cratons worldwide, we calculated S-to-P converted phase receiver function stacks using time domain deconvolution and a k-means algorithm to select robust, consistent receiver functions. Negative MLDs appear in only 50% of receiver function stacks, indicating that negative MLDs are common but intermittent. The negative MLDs correspond to shear velocity drops of 1-4%, which could be caused by layers of minerals created by metasomatism, although vertical layering in seismic anisotropy cannot be ruled out. In craton interiors, negative MLDs are lower amplitude (<3% velocity drops) and can be explained by metasomatism of the original Archean mantle. Negative MLD amplitudes increase with decreasing upper mantle shear velocity (towards the outer margins of cratons), but do not depend on the age of the craton. Thus negative MLD amplitudes are not dominated by age-related variations in cratonic mantle composition, and instead are more strongly correlated with proximity to tectonic and metasomatic activity that occurred long after craton formation. Negative MLDs are less numerous among stations that have Paleoproterozoic and Archean thermotectonic ages, consistent with the view that shallow release of slab-derived fluids during early ‘warm’ subduction was less favorable for negative MLD formation. We also observe velocity gradients below 150 km at stations in craton boundaries and interiors, indicating the presence of seismic velocity changes at the cratonic lithosphere-asthenosphere boundary and/or Lehmann discontinuity.

1 Introduction

Cratonic continental regions are characterized by thick, depleted lithospheres, high shear velocities, and low heat flow (e.g. Lee et al., 2011). While some studies characterize cratons as regions that have remained tectonically intact since the Archean (2.5 – 4 Ga), here we generalize cratons to include regions that have been tectonically stable since the Proterozoic which also exhibit thick, cool lithospheres (e.g. Artemieva, 2006).

Cratonic lithospheres often show evidence of layering in seismic velocities, which have been termed mid-lithospheric discontinuities (MLDs; e.g. Abt et al., 2010). Globally, this layering can be in the form of negative velocities gradients (NVGs), indicative of the top of a low seismic velocity layer, in the uppermost mantle, which has been observed by converted body waves. In tectonically active regions, this NVG is often interpreted as the lithosphere-asthenosphere boundary (LAB). However, in cratonic regions heat flow and xenolith constraints require the LAB to be much deeper, placing this NVG in the mid-lithosphere (e.g. Abt et al., 2010; Cooper & Miller, 2014; Eilon et al., 2018; Fischer et al., 2020; Hansen et al., 2015; Hopper & Fischer, 2015; Karato et al., 2015; Kind & Yuan, 2018; Miller & Eaton, 2010; Rader et al., 2015; Rychert & Shearer, 2009; Selway et al., 2015; Wolbern et al., 2012).

Low-velocity layers related to MLDs have key implications for the stability of the cratonic lithosphere as they are thought to be related to the deterioration of the cratonic lithosphere in locations such as the Wyoming and North China cratons (e.g. Wang & Kusky, 2019). Alternatively, some low-velocity MLDs are interpreted as ancient structures produced during craton amalgamation such as the hydrated surface of ancient subducted slabs or vertical variations in anisotropy (e.g. Bostock, 1998; Rader et al., 2015; Saha et al., 2021; Selway et al., 2015; Wirth & Long, 2014). Another proposed origin for MLDs is the global occurrence of

elastically accommodated grain boundary sliding (Karato, 2012). Some of these mechanisms are not mutually exclusive and, in reality, they may combine to produce MLDs. In this study we provide improved constraints on the (i) depth, (ii) magnitude (velocity change at discontinuity), and (iii) geographic and geologic distributions of MLDs, allowing us to test which of the proposed mechanisms are consistent with global observations.

Recently, a study of converted S-to-P body waves has called the existence of a mid-lithospheric NVG in North America into question, suggesting that the NVGs apparent in prior studies are artifacts of the use of deconvolution in data processing (Kind et al., 2020). A secondary goal of our paper is to reevaluate the impacts of different data processing steps on the arrivals observed in stacks of S-to-P receiver functions. However, we also note that a variety of seismic phases have indicated the presence of NVG and/or low velocity layer in the shallow cratonic lithosphere, including Rayleigh-waves (e.g. Dalton et al., 2017; Eeken et al., 2018), topside reflections of direct S phases (Shearer & Buehler, 2019), SS precursors (Tharimena et al., 2017), ScS reverberations (Revenaugh & Jordan, 1991), P reflectivity (Sun & Kennett, 2017), and long-range refraction surveys (Thybo & Perchuc, 1997).

In addition to MLDs, in this study we also examine deeper discontinuities in the cratonic mantle. One of these is a potential NVG associated with the base of the lithosphere, which is thought to lie at depths of 150-350 km based on velocity and attenuation tomography (e.g. Dalton et al., 2017; Fischer et al., 2020; Gung et al., 2003). A second mantle discontinuity is a positive velocity gradient (PVG) that has been observed beneath the continents in the 150-300 km depth range with both body waves (e.g. Revenaugh & Jordan, 1991) and surface waves (e.g. Dalton et al., 2017; Gung et al., 2003), a feature often associated with the Lehman discontinuity (Lehmann, 1959).

In this study, we characterize the structure of the cratonic upper mantle by analyzing NVGs and PVGs using S-to-P phases at globally distributed stations within and at the margins of cratons (Fig. 1-1). For the purposes of this study, we define cratonic mantle as those regions where shear wave velocities exceed 4.5 km/s at 150 km depth in the global shear wave model SEMum2 (French et al., 2013).

2 Data and Methods

In order to capture structure in the cratonic upper mantle (~50-300 km), we use deconvolved S-to-P converted body waves (Sp). Sp phases are advantageous at lithospheric depths because crustal multiples of the primary S phase do not overprint arrivals from the shallow mantle, in contrast to P-to-S converted waves (Ps) whose shallow mantle arrivals can be overprinted by crustal multiples. In addition, the sensitivities of Sp phases to velocity structure make them highly effective at resolving low angle ($<10^{\circ}$ - 15°) shear velocity gradients (Hua et al., 2020b). However, Sp phase signal-to-noise ratios are in general lower than those for Ps. To best resolve Sp receiver functions, in particular to avoid creating the illusion of discontinuities through data processing artifacts such as side-lobes, we adopt a strategy that involves: long-running and high-quality seismic stations; transformation to P and SV components; time domain deconvolution on broadband filtered data (2-100s); clustering analysis to objectively identify and select high-quality RFs; median weighted bootstrap stacking; and only interpreting final RF stacks with sufficient quantity and quality of individual RFs. Key aspects of these steps are described below, and further detail is provided in the Supporting Information.

2.1 Data qualifications and initial processing

We collect waveform data from 35 long-running (>8 years), broadband seismometers dispersed throughout cratons and their margins globally, plus a station in the tectonically active western

U.S. for comparison (Fig. 1-1, Table S1-1). Seismic network information is provided in the acknowledgements.

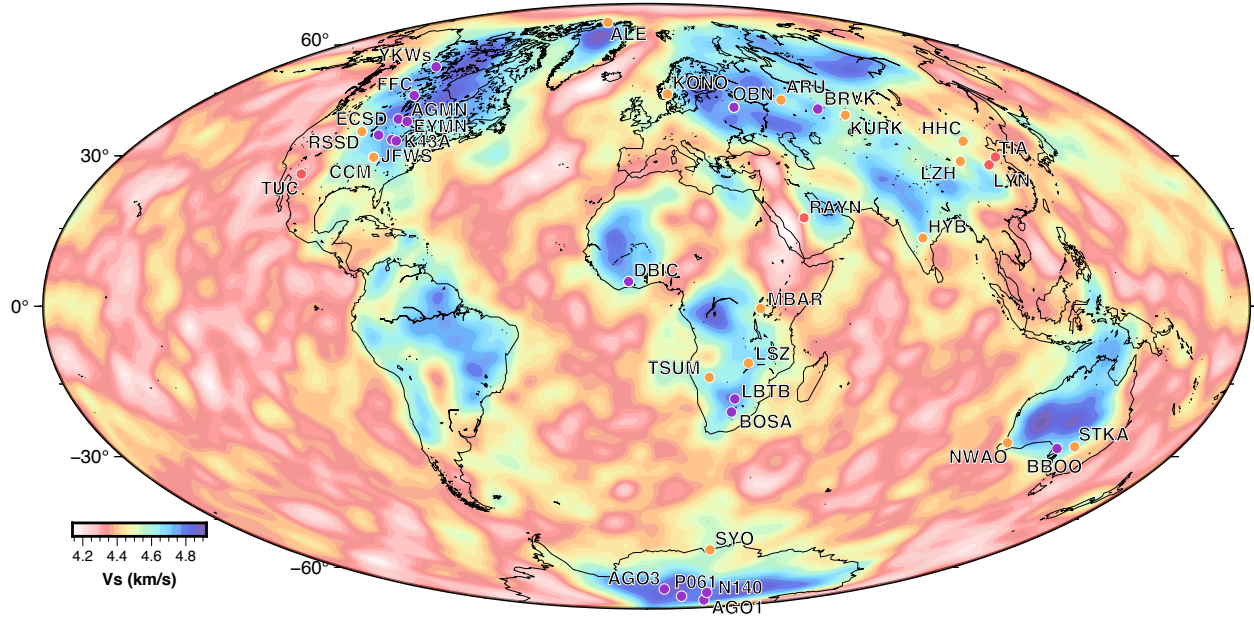


Figure 1-1. Global map of stations

Stations (circles) used in this study are underlain by the shear velocity at 150 km depth from SEMum2 (French et al., 2013). Circle colors indicate whether the station overlies the non-cratonic lithosphere (red), the edge of the cratonic lithosphere (orange), or the cratonic lithosphere (purple) as classified in Figure 1-6.

Waveforms are separated into 30° backazimuthal bins to account for potential lateral variations (e.g. Eilon et al., 2018). Waveforms are filtered from 2-100s (see Fig. SA-1 for details) and converted to P – SV components using the free-surface transformation (FST) and free-surface velocities determined from direct S and P waveforms (e.g. Hua et al., 2020a). Use of P and SV components for calculating RFs eliminates large amplitude direct phase arrivals at zero time, as can be seen by comparing Fig. 1-2 to Fig. SA-2. Prior to deconvolution, we discard waveforms if the signal-to-noise ratio (SNR) of the parent (S-wave) component is less than 1.5 and the onset time of the parent phase is more than 15 s from the predicted arrival-time (Text SA-1).

2.2 Receiver function processing and selection

We deconvolve the parent S component from the daughter P component of seismograms in the time domain to produce Sp receiver functions (RFs; Ligorria & Ammon, 1999), using a Gaussian shaping filter with a half-width of 0.9 s. Time domain deconvolution is more stable at long periods and is less likely to produce side-lobes around large phases than frequency domain deconvolution (Fig. 1-2; Lekic & Fischer, 2017; Ligorria & Ammon, 1999). For example, in the synthetic case shown in Fig. 1-2, time domain deconvolution yields clear phases from each of the discontinuities in the velocity model used to calculate the waveforms with virtually no side-lobe energy, even around the large Moho phase. In contrast, even more stable frequency domain approaches such as extended time multi-taper deconvolution tuned to minimize side-lobes (e.g. Mancinelli et al., 2017) cannot fully eliminate these artifacts (Fig. 1-2(c)).

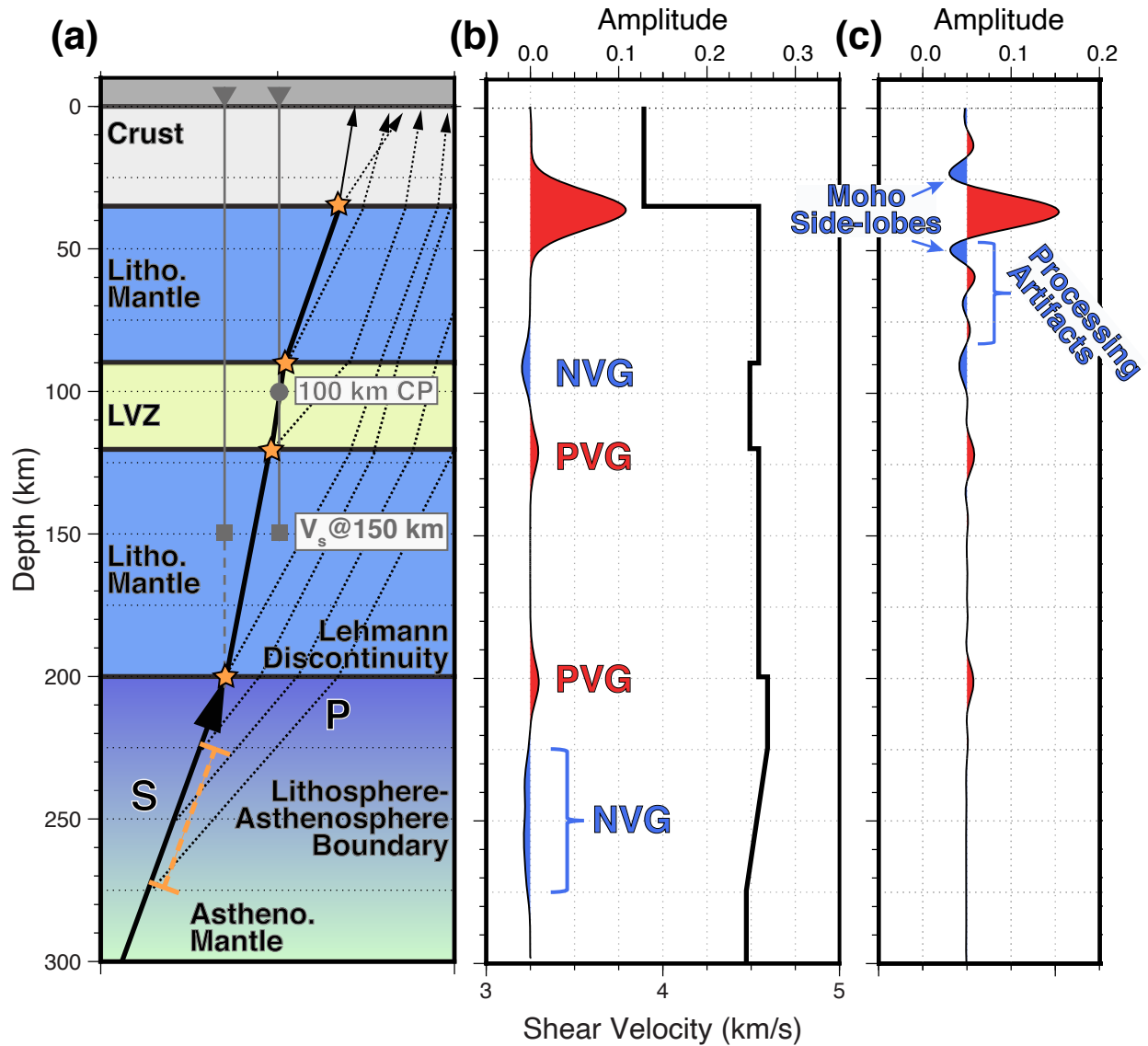


Figure 1-2. Synthetic RF example for cratonic lithosphere

(a) Geologic structure with ray paths for S (solid) and P (dotted) phases with (b) the resulting synthetic time and (c) frequency domain Sp RFs. (a) This schematic features velocity gradients hypothesized to exist in the cratonic mantle: MLD (a. yellow; b. low velocity layer), Lehmann Discontinuity (a. purple; b. high velocity layer), and LAB (a. purple to green; b. gradual negative velocity gradient). Orange marks (stars, dashed line) indicate the locations of S-to-P conversions. Hypothetical ray paths of an S-wave (solid black) and S-to-P converted phases (dotted black) are drawn. Gray lines illustrate the 150 km shear velocity (square) projected to the surface (inverted triangle: thermotectonic age) used for evaluating phases (left) and organizing RF stacks (right). (b) The synthetic time domain Sp RF (left) exhibits phases that clearly correspond to positive and negative velocity gradients from the forward-modeled shear velocity structure (right, black line). At discontinuities in the V_s structure, if velocity increases (decreases) downwards, the discontinuity is represented by a positive (negative) phase. (c) Using the same structure, a forward modeled Sp RF is calculated using deconvolution with an extended time multi-taper method. Note the additional side-lobe phases produced around the Moho phase, extraneous signals in the mantle, and the absence of a phase from the gradual cratonic LAB velocity gradient. The Sp RF polarity has been flipped to match the Ps convention.

Kind et al. (2020) argue that side-lobes around the Moho are a fundamental drawback of converted body wave studies that employ deconvolution, leading to artifacts that can be misinterpreted as discontinuities. However, the example of time domain deconvolution shown in Fig. 1-2 and our experience in general indicate that carefully applied time domain deconvolution can avoid Moho side-lobes. Moreover, deconvolution has the great advantage of removing the effects of variable source-time functions and allowing each vertical change in velocity to be represented as a single impulse.

To enable further selection of high-quality RFs, we apply the k-means++ algorithm (Arthur & Vassilvitskii, 2007). We use this algorithm to separate RFs into four groups based on similarity in the 9 s before the parent (S) arrival (plotted as 0 s) (Fig. 1-3). These times should encompass the arrival of an S-to-P conversion from the crust-mantle boundary (the Moho) as well as any residual energy at 0 s remaining if the FST was not effective. The separation of data based on amplitude of the Moho conversion and energy present at 0 s allows us to identify traces that clearly exhibit S-to-P converted energy from a Moho velocity increase (to keep in our data set) as well as traces that have been mis-transformed to P-SV (to avoid in our data set). The user picks the clusters of data that exhibit low noise, low 0 s energy, and prominent Moho arrivals (Fig. 1-3(a)). Note that the selection explicitly avoids signals at mantle depths, in order to avoid bias in mantle arrivals in the resulting RF stacks.

Continuing with the seismograms selected through k-means cluster analysis, we repeat the process of solving for surface velocity values using the same method as in section 2.1, which is improved with the exclusion of poor-quality data. Seismograms are then migrated to the average ray parameter of all remaining events. We calculate RFs from newly transformed P-SV component seismograms and perform five consecutive k-means clustering analyses (i.e. Fig. 1-

3(b)). Selected seismograms are weighted using the number of times they are selected in the five iterations of k-means. This method prioritizes data identified as consistent through k-means, which helps to minimize the influence of side-lobe energy that is more prominent in low-quality RFs.

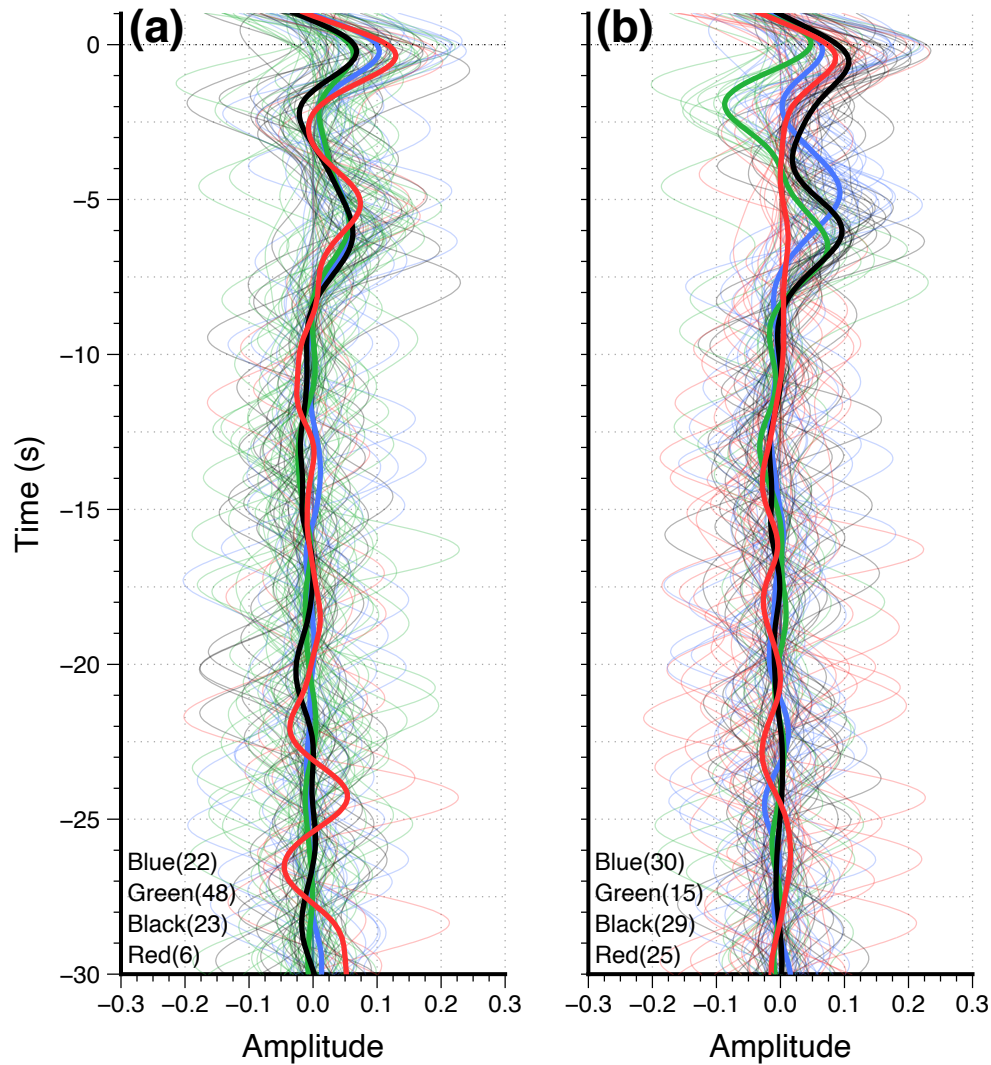


Figure 1-3. Example of the k-means process

Example of the k-means process for selecting high-quality traces with strong signals from the Moho discontinuity for RF stack CCM 1. Individual traces (thin lines) are grouped into clusters (colors) using the k-means algorithm. Medians of bootstrapped stacks for each cluster are shown by the bold lines. We provide examples for (a) the initial data culling process and (b) the second use of k-means in the final stacking procedure. Here we show examples from the cityblock clustering method. In this example, we chose (a) the blue, green, & black clusters. For one iteration of the final stacking procedure, we chose (b) the blue, green, and black clusters. The number of traces in each cluster is indicated in the bottom left.

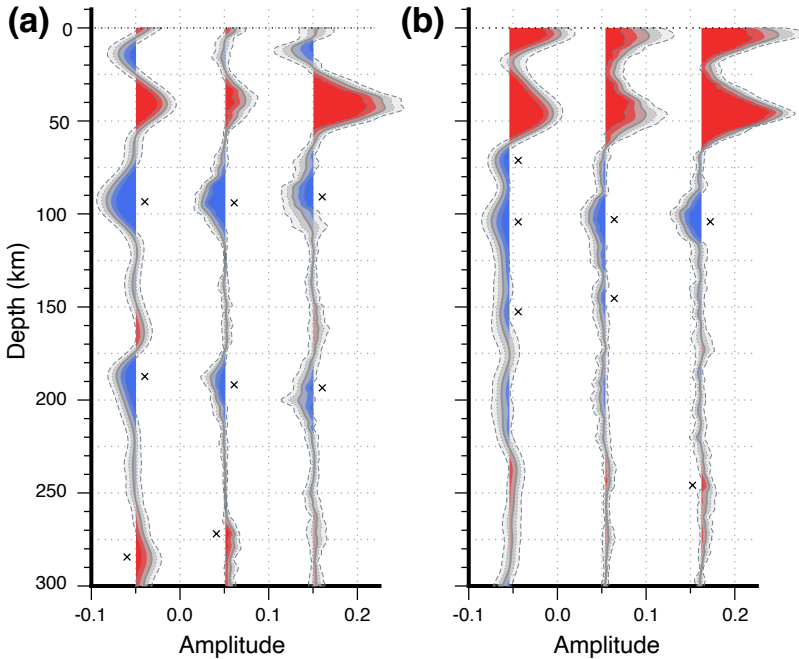


Figure 1-4. Analysis of processing procedures

These figures illustrate the variation in resolved phases using different processing procedures for RF stacks **(a)** RSSD 1 and **(b)** CCM 1. The data is culled and stacked in various ways. From left to right: all data SNR>3, mean bootstrap stacked; all data SNR>3, median bootstrap stacked; and data picked and weighted with additional 5 iterations of k-means (our final RF stack). **(a)** The RF stack for RSSD 1 does not change significantly at mantle depths with the different processing choices, while **(b)** the RF stack for CCM 1 does. Most notably, the small negative phase below the Moho is insignificant after processing with k-means and use of the median in each bootstrap iteration. The combination of k-means data culling and median bootstrap stacking represents a conservative choice that minimizes energy at mantle depths. RF stacks are plotted as in Fig. 3-1 to 3-5(b).

Some caveats regarding using k-means clustering for data selection with a focus on Moho arrivals are that Moho amplitudes (Fig. SA-16 to SA-58; Fig. 1-4) may be over-accentuated. In addition, this approach would not be sensible in regions known to lack significant Moho arrivals, but this is not the case for any of the stations used in this study.

The RFs resulting from the k-means analysis are bootstrapped in the sense that they are randomly selected 100 times. The median amplitude of each random sample is calculated, then the mean of the 100 medians is taken and smoothed with a Savitzky-Golay filter (Fig. 1-5(b)). The use of the median reduces the impact of individual RFs that are outliers and tends to produce less energy at mantle depths (Fig. 1-4). This is particularly important, as we often observe strong

side-lobe energy on individual RFs that are outliers (Fig 1-5(a), red). We only interpret RF phases whose amplitudes exceed 2-standard deviation uncertainty (black x's and o's, Fig. 1-5(b), Fig. 1-6, Figs. SA-14 to SA-56). Ps RF stacks for the stations in this study were also calculated using the same approach (Text A-2.1 and Fig. SA-4).

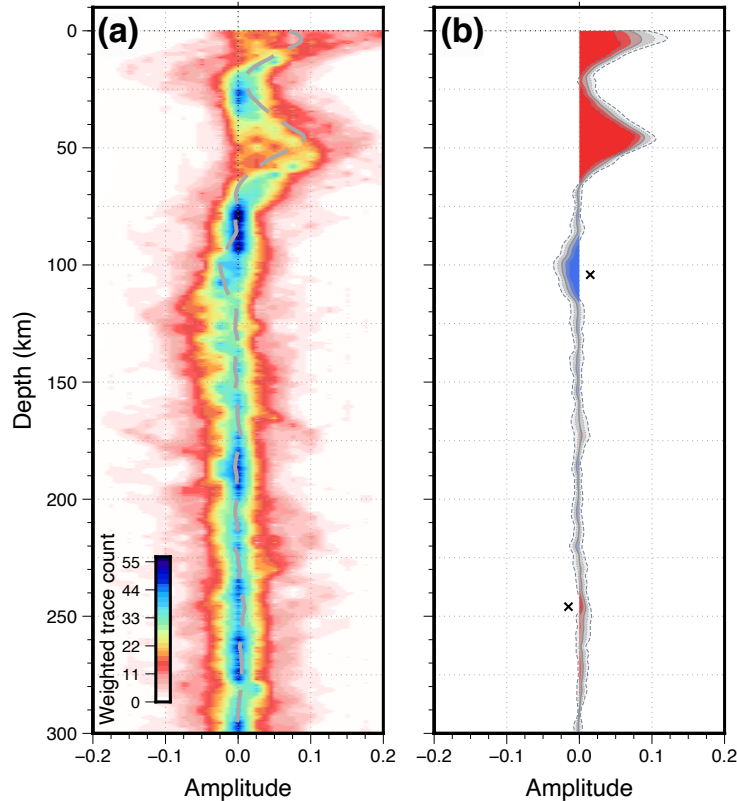


Figure 1-5. Example of results for CCM 1

Example of the results of our stacking procedure for CCM 1. This stack contains 90 waveforms. **(a)** A heat map of the number of traces that overlap at a particular amplitude and depth indicates data consistency. The RF stack amplitude is plotted as a gray dashed line, which follows the amplitude where most RFs are represented. **(b)** RFs are weighted, bootstrapped stacked, and migrated to depth to produce this RF stack. The gray shaded regions represented one and two standard deviations on either side of the stacked amplitude. Sp RF polarity has been flipped to match the Ps convention.

3 Receiver Function Stack Results

We characterize our results by the shear velocity at 150 km depth from the SEMum2 shear velocity model (French et al., 2013) below the 100 km depth conversion point of the RF

stack (Fig. 1-1, 1-2, 1-6; Table S1-1). We are interested in the shear velocity (V_s) at this depth because, at 150 km depth, cratonic regions tend to exhibit high lithospheric velocities whereas non-cratonic continental regions have lower asthenospheric velocities (Fig. SA-5). We are considering velocities below the 100 km conversion point of the RF stack because we observe most MLDs around this depth (Fig. 1-7). We classify stations as non-cratonic (Fig. 1-6(a); $V_s < 4.5$ km/s), craton boundary (Fig. 1-6(b); 4.5 km/s $< V_s < 4.675$ km/s), and craton interior (Fig. 1-6(c); $V_s > 4.675$ km/s) based on this shear velocity. Some stations lie within craton boundaries based on regional geology (Fig. SA-5, outlines) but do not overly cratonic lithosphere (i.e. Fig. SA-5(d), TIA) or exist on the edge of cratonic lithosphere (i.e. Fig. SA-5, (b) MBAR & (c) KONO) as defined by V_s .

For the purposes of plotting, we migrate RFs to depth using the SEMum2 earth structure model (French et al., 2013) and reverse the polarity of Sp RFs to match the convention of Ps RFs. Positive polarities correspond to positive velocity gradients (PVGs) and negative polarities correspond to negative velocity gradients (NVGs). All of the RF stacks presented have strong PVGs associated with the Moho discontinuity that begins near 20 km depth and continues until 50-75 km depth (Fig. 1-6; Fig. SA-16 to SA-58). We focus on conversions from below this high amplitude, positive energy as mantle discontinuities.

A first-order result of the RF stacks is that mantle phases in the Sp RF stacks differ between these groups based on V_s at 150 km, although variation exists within each group. The majority of non-cratonic RF stacks (Fig. 1-6 (a)) have large amplitude NVGs below the Moho, craton boundary RF stacks (Fig. 1-6 (b)) have moderate amplitude NVGs distributed variably throughout the mantle, and craton interior RF stacks (Fig. 1-6(c)) have low amplitude velocity gradients that are positive and negative.

We note that some RFs have high amplitude energy at 0 s. This energy could be due to real crustal structure or possibly to difficulties in determining free-surface velocities that minimize incident wave energy. However, most mantle phases identified as significant are relatively insensitive to uncertainties in the FST. Phases that have the potential to be side-lobe phases sometimes appear as negative amplitudes directly below the Moho (i.e. they look similar to the negative-positive swings of a sine wave; Fig. 1-2). We have confidence that negative phases below the Moho represent real structure only if they have relatively large negative amplitudes, distributed negative amplitudes with depth, and/or the presence of some zero-amplitude energy between the Moho and NVG phase. We find that our careful processing and conservative interpretation of mantle phases remove virtually all suspicious phases from consideration (i.e. RAYN 2, HHC 1, TSUM 1), although we remain skeptical of the shallowmost negative mantle phases in RF stacks ARU 1, LSZ 1, DBIC 1, and OBN 1 (Fig. 1-6, boxed names). Because of this skepticism, we also perform a full analysis of trends in the RF phases without the stacks for these stations (Fig. SA-7) and find that our conclusions do not significantly change.

Figure 1-6. Sp RF stacks for all stations. Stacks are assumed to overly (a) non-cratonic lithosphere ($V_s < 4.5$ km/s), (b) the edge of the cratonic lithosphere ($4.5 < V_s < 4.675$ km/s), and (c) cratonic lithosphere ($V_s > 4.675$ km/s) based on the shear velocity at 150 km depth beneath the 100 km conversion point of the stack (Table S1). RF stacks are plotted as in Fig. 1-5(b). White circles indicate phases that are outside of two standard deviation error. Black x's indicate the most robust phases, in which the amplitude out of two standard deviations is at least 50% of one standard deviation. For a full analysis with only use these most robust phases, see Figure S6. Light blue boxes around stack names indicate RF stacks with possible side-lobe energy below the Moho phase, although some of this energy could actually be derived from mantle structure.

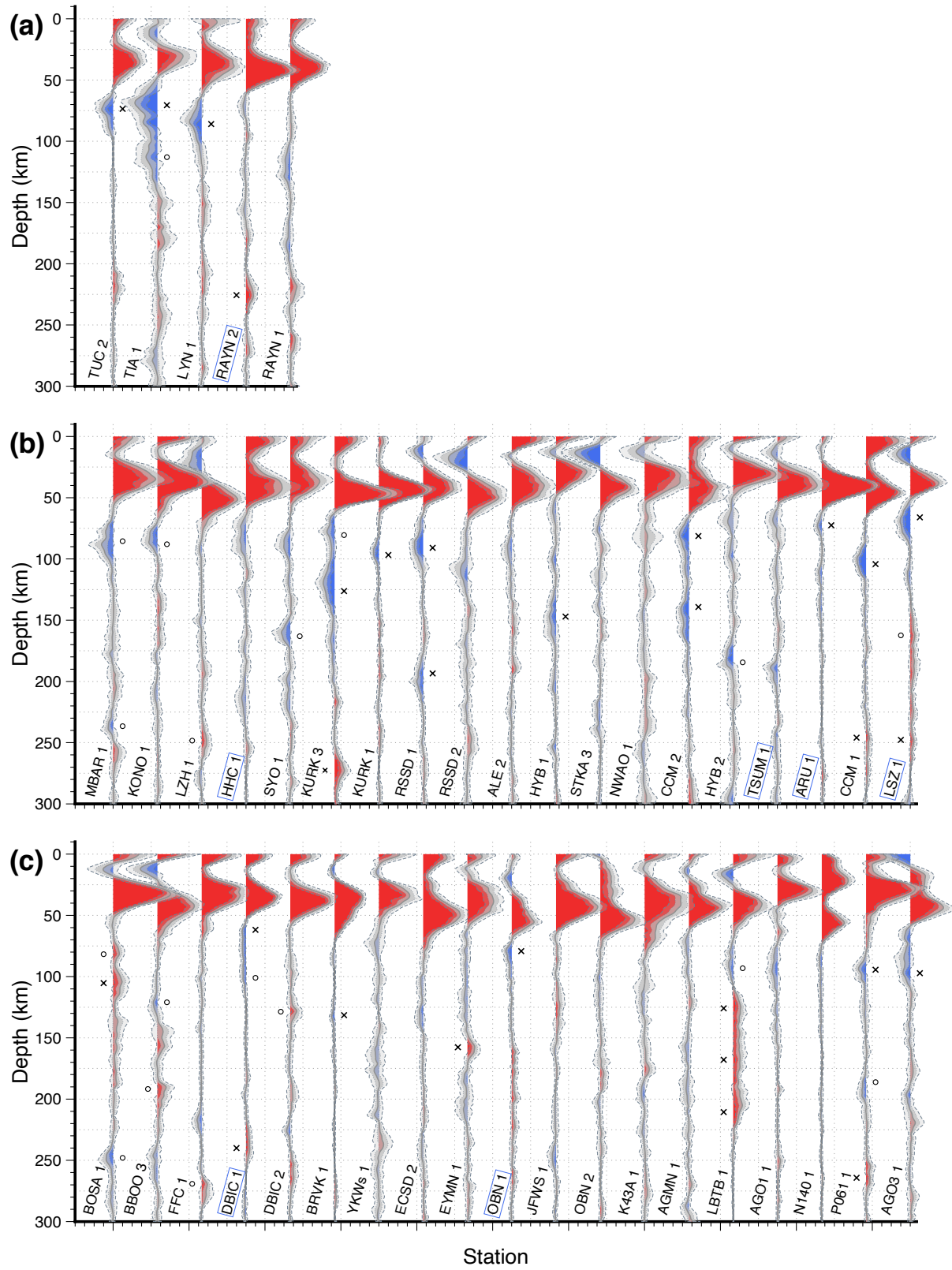


Figure 1-6. Sp RF stacks for all stations

3.1 Non-Cratonic continental regions

Non-cratonic continental regions lack an underlying high-velocity lithosphere and are identified by shear velocities of less than 4.5 km/s at 150 km depth using SEMum2 (French et al., 2013). Station TUC is located in the tectonically active western US, which we present for comparison. The RF stack for TUC 2 is characterized by a high amplitude NVG below the Moho to 100 km depths (Fig. 1-6(a)), which is thought to represent the LAB (Abt et al., 2010). Stations TIA and LYN in the eastern North China craton also exhibit relatively high amplitude NVGs below the Moho.

RF stacks for station RAYN correspond to slow SEMum2 mantle velocities but are an exception in that they exhibit little mantle energy. A possible explanation is that mantle velocities beneath RAYN are higher velocity than can be resolved in the global SEMum2 model, because of the large lateral velocity gradient from the Red Sea rift into the Arabian platform (Fig. SA-5(b)). Celli et al. (2020) observe a shear velocity increase of $\sim 16\%$ at 150 km depth over ~ 500 km in the center of the Arabian Peninsula near the location of station RAYN. This observation is supported by Liu et al. (2020), who only observe deep (~ 130 km) NVGs in a similar location.

3.2 Cratonic boundaries

We identify cratonic boundaries as the transition between regions that lack high-velocity cratonic lithosphere ($V_s < 4.5$ km/s) and regions that definitively overly high-velocity cratonic lithosphere ($V_s > 4.675$ km/s) at 150 km depth (French et al., 2013). We note that MBAR 1, KONO 1, and RSSD 1 all exhibit negative phases that appear similar to the phase observed at TUC 2 for a non-cratonic LAB. These RF stacks are all sampling mantle lithosphere that is adjacent to or potentially interacting with asthenosphere. MBAR exists at the eastern edge of the

Tanzanian craton which is surrounded by the east African rift (Fig. SA-5(b)). Celli et al. (2020) observe a high velocity cratonic lithosphere below the Tanzanian craton extending to ~150 km depth that is surrounded by a low velocity anomaly associated with the east African rift system. Similarly, the location of the Oslo rift in Scandinavia coincides with the lithosphere sampled by KONO 1 (Neumann et al., 1992) where Rayleigh group velocities observe low seismic velocities in the upper mantle (Weidle & Maupin, 2008). Station RSSD is on the eastern edge of the Wyoming craton, which has lost mantle lithosphere from its western edge and has experienced multiple episodes of metasomatism (e.g. Carlson et al., 2004; Eilon et al., 2018). It is apparent from higher resolution regional models (Babikoff & Dalton, 2019) that mantle velocities vary significantly at the edge of the Wyoming craton at a relatively fine scale. The RSSD RF stacks also show lateral variations. The RSSD 1 stack that contains conversion points to the northwest of the station contains a high amplitude NVG at ~90km, whereas RSSD 2 with conversion points to the southeast does not (Fig. SA-5a).

The potential MLD observed in RF stack ARU 1 is possibly a Moho side-lobe, since the negative phase is low-amplitude and concentrated below the Moho. Also, the negative phase below the Moho in RF stack LSZ 1 may be the result of constructive interference of a real negative MLD and a Moho side-lobe based on its large amplitude and position below the Moho (Fig. 1-6(b)).

3.3 Craton Interiors

We consider regions with shear velocities greater than 4.675 km/s at 150 km depth to have high velocity cratonic lithosphere (French et al., 2013). In this group, NVGs in the upper

mantle are relatively weak. We also observe significant PVGs in the mid-lithosphere of the Kaapvaal craton at BOSA 1 and LBTB 1.

Stations AGO1, AGO3, P061, N140, and SYO sample the east Antarctic craton. Based on SEMum2, these stations overly fast cratonic lithosphere. However, regional models of the east Antarctic craton (An et al., 2015) reveal more lateral variability in shear velocity structure than resolved in SEMum2, indicating that the entire lithosphere underlying the Antarctic craton may not have such high velocities. Furthermore, Shen et al. (2018) suggest that the lithosphere beneath the central Transantarctic mountains and adjacent east Antarctic craton has delaminated. Although not directly sampled by our RF stacks, it is possible that lithospheric foundering has influenced the local structure in east Antarctica. While we retain the Antarctic stations in the craton interior group in the following discussion, we also show results with and without them (Fig. 1-9 & S1-10).

The NVG at DBIC 1 is distributed in depth, so at least the deeper portion of this arrival likely represents mantle-derived energy. However, the higher amplitude of this phase directly below the Moho may represent a Moho-related artifact. We are also skeptical of the existence of the potential MLD observed at OBN 1 as the phase is low-amplitude and concentrated in depth directly below the Moho (Fig. 1-6(b)).

3.5 Results summary

We observe NVGs in the shallow cratonic lithosphere in 17 of 38 craton boundary and interior RF stacks and are confident that at least 15 of these NVGs are due to Earth structure. We also observe PVGs in the mid-lithosphere of three craton interior RF stacks. In addition to shallow velocity gradients in a large proportion of cratonic lithosphere seismic stations, the Sp

RF stacks also contain phases consistent with velocity gradients beginning between 150 and 300 km depths. We observe these velocity gradients out of two standard deviations in 15 of our craton boundary and interior RF stacks. We acknowledge that some of the phases we interpret are small in amplitude and/or are barely out of error. To add additional confidence to our interpretation in the following section, we repeat the analysis using the criterion that phase amplitude must exceed twice the uncertainty and find results that are even more compelling in support of our conclusions (Fig. SA-6).

4 Discussion

4.1 Mid-lithosphere structure

We systematically analyze mid-lithospheric structure to draw conclusions about MLDs on a global scale (Fig. 1-7 to 1-9). The new MLD observations in this study delineate previously unrecognized global patterns in MLD occurrence that shed new light on the origins of layering in the mid-lithosphere of cratons.

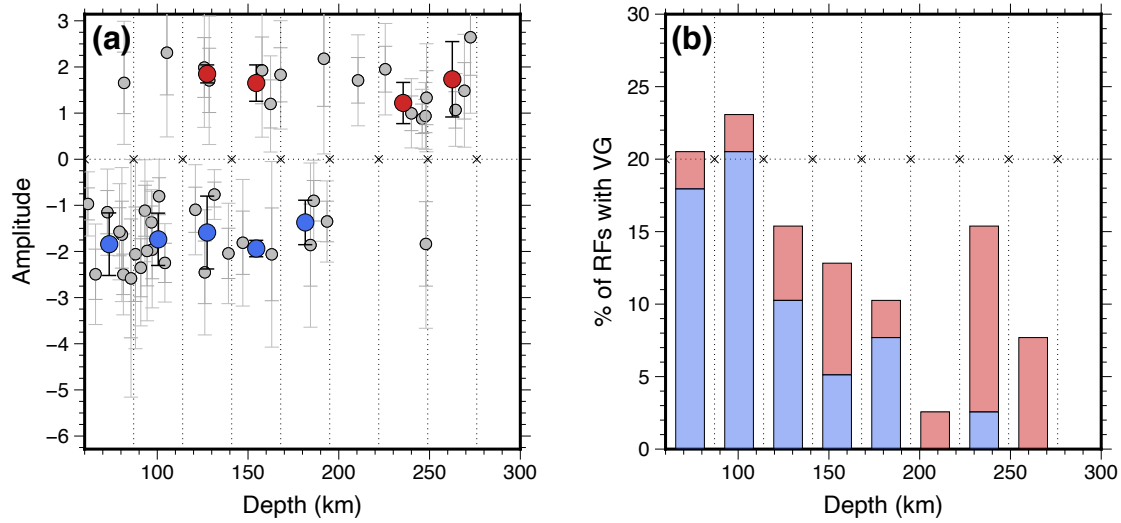


Figure 1-7. Velocity gradient distribution with depth

Velocity gradient distribution with depth in craton interiors and boundaries (>4.5 km/s). (a) Gray circles represent the amplitude of individual phases. Data is binned (black X) by depth to show averages of NVGs (blue circles) and PVGs (red circles) within each bin if at least four phases are present. (b) The number of PVGs and NVGs within a single depth bin relative to the total number of RF stacks are represented by red and blue bars, respectively.

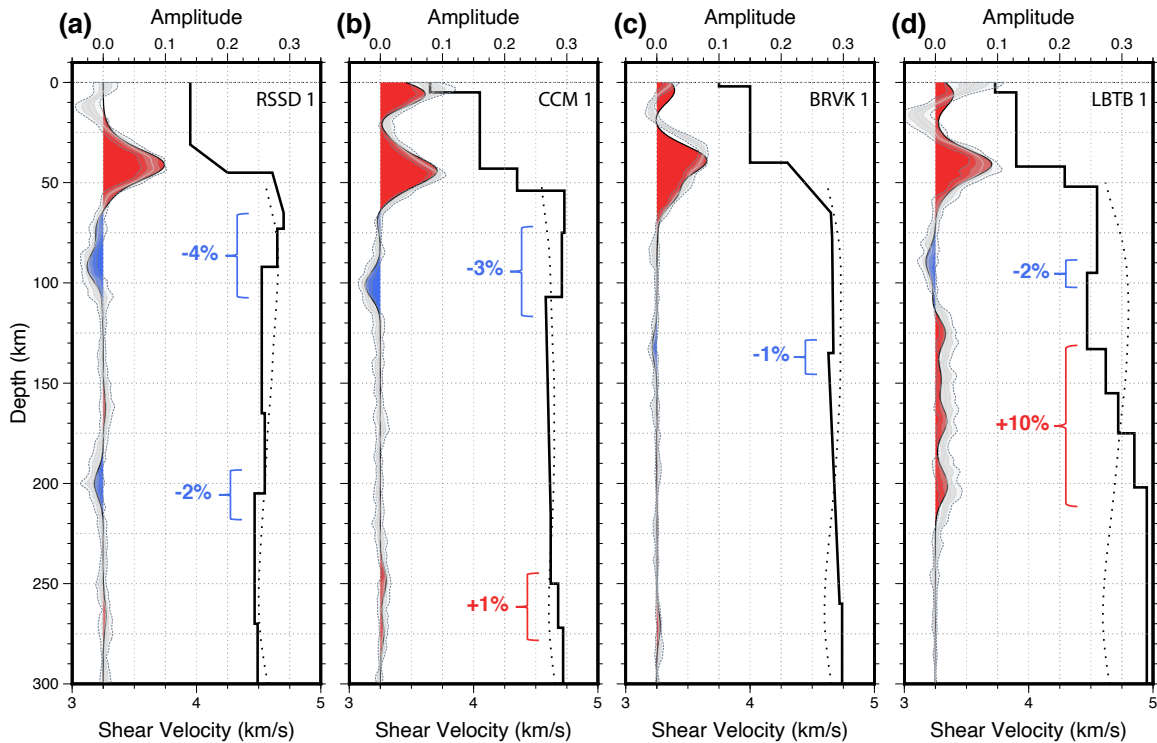


Figure 1-8. RF model predictions

Observed RFs (left plot in each panel, gray dotted line) are compared to model predictions (left, red/blue RF). The corresponding velocity model (right, black line) is made to match the velocity gradients observed while maintaining consistency with the SEMum2 velocity profile at the RF stack location (right, black dotted line). One (light gray) and two (gray) standard deviation errors encompass the observed results, similar to plotting in Fig. 1-6. The percent velocity change across velocity gradients are annotated. This is done for RF stacks that sample regions classified as craton boundaries; (a) RSSD 1, (b) CCM 1; and craton interiors; (d) BRVK 1, and (e) LBTB 1.

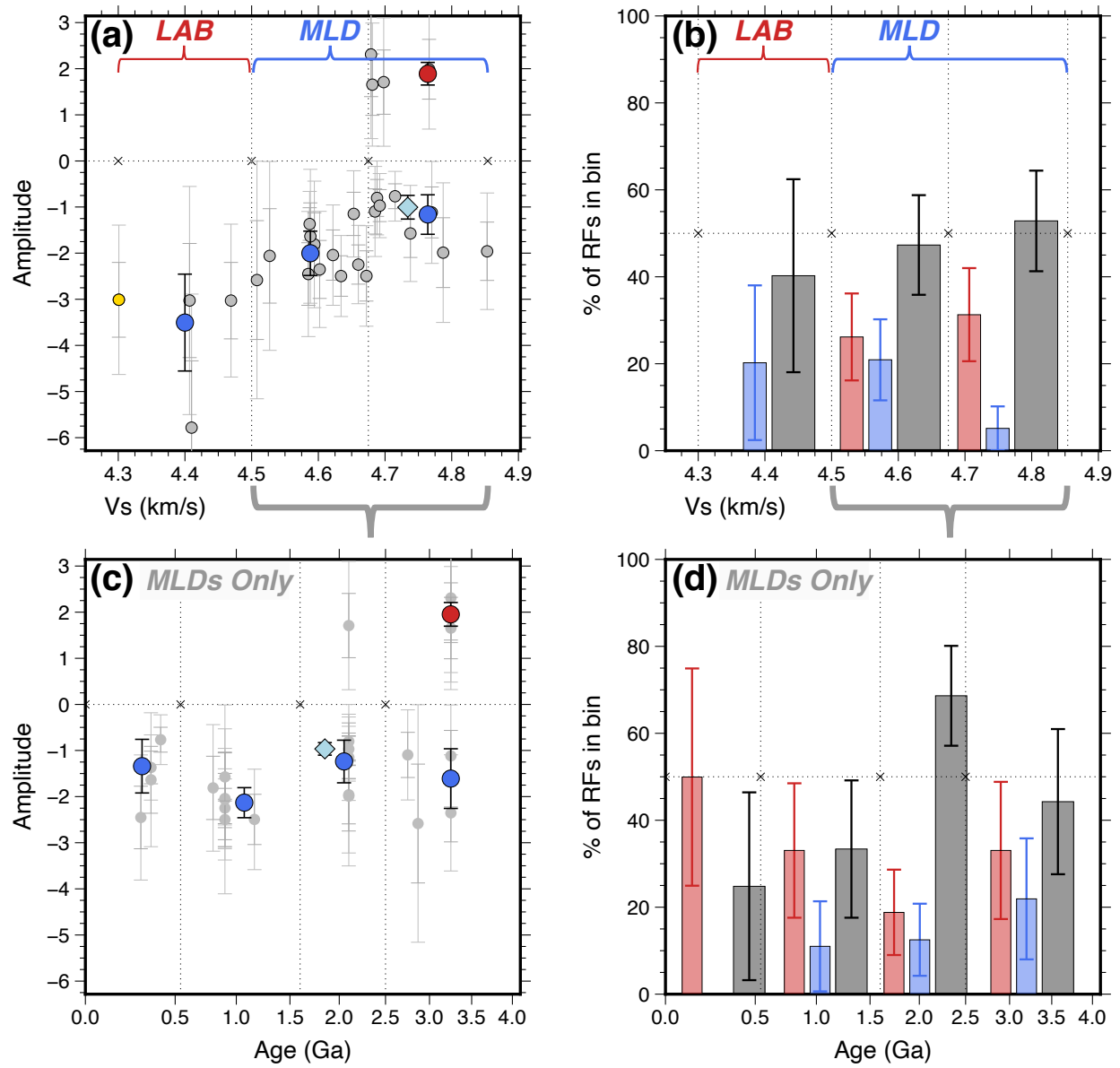


Figure 1-9. Negative velocity gradient phases

Negative velocity gradient (NVG) phases identified for RFs presented in Fig. 1-7. **(a)** All observed phases in the Moho to 150 km depth range, plotted by their amplitude and the shear velocity at 150 km depth at the location of the phase conversion point (gray circles) with one and two standard deviations error bars. To evaluate this scattered data, we bin (black X) phases by shear velocity to correspond with the categories of non-cratonic (<4.5 km/s), cratonic boundary (4.5-4.675 km/s), and cratonic interior (>4.675 km/s). NVG (blue circles) and PVG (red circles) amplitudes are averaged within each bin. Light blue diamonds are the average negative MLD amplitude in the bin excluding phases from central Antarctica. The yellow point indicates the NVG from TUC 2 which previous studies agree represents a LAB. **(b)** RF stacks with no MLDs in this depth range are represented as a bootstrapped percentage of the total number of RF stacks within each Vs bin (dark gray bars). RF stacks with PVGs (red bars) and NVGs (blue bars) from 150-300 km depth are plotted as a bootstrapped percentage of the total number of RF stacks within each Vs bin. Errors bars represent one standard deviation. Phases in **(c/d)** are the same as **(a/b)**, but only presents phases that are greater than 4.5 km/s at the 150 km depth conversion point. Data in **(c/d)** are organized by the thermotectonic age at the location of the phase conversion point. Data is binned by age: Phanerozoic (0-541 Ma), Meso/Neo-Proterozoic (541-1600 Ma), Paleoproterozoic (1600-2500 Ma), and Archean (2500-4000 Ma).

4.1.1 Global observations of MLDs

It is clear that mid-lithospheric structure varies significantly at global, regional, and local scales (Fig. 1-6, Fig. SA-5). Cratonic RFs exhibit mid-lithospheric phases in only ~50% of stacks (Fig. 1-9(b), black error bars). In contrast, if station RAYN is excluded (see section 3.1, Fig. S1-12), all non-cratonic stacks in this study have NVG phases in the LAB depth range. At a local scale, in some cases results for different backazimuthal stacks at the same station are variable (i.e. RSSD 1 & 2, DBIC 1 & 2). This could be due to anisotropy or lateral changes in structure (over ~100 km).

We use the Propagator Matrix method (Keith & Crampin, 1977) to forward model a subset of high-quality cratonic RF stacks in order to better understand their implications for elastic velocity changes in the upper mantle (Fig. 1-8). With the local structure from SEMum2 as a starting model, we computed synthetic RFs for a range of candidate velocity structures, and matched the amplitudes and shapes of the predicted and observed mantle phases. Both sharp and gradual velocity changes are required to replicate the variety of phase widths we observe. To match the range of observed mantle phases, shear velocity drops from 1% (BRVK 1; Fig. 1-8(d)) to 4% (RSSD 1; Fig. 1-8(a)) are required. The velocity decrease at RSSD 1 is similar to the ~5% velocity reduction found in a joint inversion of Sp, Ps and Rayleigh wave data by Eilon et al. (2018). We also model a RF stack with moderate (CCM 1; Fig. 1-8(b-c)) NVGs below the Moho, which requires a shear velocity drop of 3% to replicate the observed phases. We only observe high amplitude NVGs, similar to RSSD 1, at craton boundaries. The estimated Vs drops are lower than some prior estimates of Vs drops at cratonic MLDs (2-12%, Hopper & Fischer, 2018; 12%, Wolbern et al., 2012). However, in the case of Hopper and Fischer (2018) the

inferred velocity drops were often distributed over larger depth ranges, in some cases explaining the higher velocity reductions.

In order to characterize MLDs further, we use thermotectonic age (Artemieva, 2006) at the conversion point of a phase to compare stacked phases of RFs that sample craton boundaries and interiors ($V_s > 4.5$ km/s; Fig. 1-9(c-d)). There is no significant trend of negative MLD amplitude with thermotectonic age (Fig. 1-9(c)). However, there is a clear trend of weakening negative MLD amplitude (Fig. 1-9(a), blue circles) with increasing shear velocity. This trend becomes statistically significant if Antarctic stations are excluded (Fig. 1-9(a), light blue diamond; see section 3.3), in the sense that the two-standard deviation uncertainties for the mean amplitudes in the two MLD bins do not overlap. The trend indicated by the binned amplitude measurements is corroborated by a linear regression of the individual amplitudes as a function of V_s (Fig. SA-8).

An interesting pattern emerges when we count the number of stacks with negative MLD velocity gradients in each age bin (Fig. 1-9(d)). RF stacks without any velocity gradients between the Moho and 150 km depth are more prevalent in regions of Paleoproterozoic (~75%) and Archean (~45%) age than in younger regions (<30%).

We find that there are some discrepancies between the global trends we observe and individual regions. As noted in sections 3.1-3.3, many of these discrepancies are explained by the resolution of the global SEMum2 velocity model which does not capture local variations in shear velocity structure. Some of these regions are also sparsely sampled with seismic instrumentation, particularly Antarctica and northern Africa, which inherently decreases velocity model resolution. To verify that these regional exceptions are not skewing our results, we

reanalyze MLD data as in Figures 1-7 & 1-9 without these exceptional stations and find there are not significant changes in our overall conclusions (Fig. SA-12, SA-13, SA-15).

Positive MLDs are observed only in craton interiors (Fig. 1-9(a)), emphasizing the unusual properties of the thick, high velocity lithosphere in these regions. These PVGs are all observed in Africa, with the most significant positive phases from the Kaapvaal craton (stations LBTB and BOSA).

4.1.2 Do MLDs Exist?

We resolve MLDs in ~50% of RF stacks that sample cratonic lithosphere (Fig. 1-6(b); > 4.5 km/s at 150 km depth). Contrary to the conclusions of Kind et al. (2020) that widespread MLDs may not exist, we observe negative MLDs globally which we are confident are not artifacts of our waveform processing. However, our observations are also not consistent with the view that MLDs are a nearly continuous feature of the cratonic mantle. Rather, they are intermittent. MLDs are most common at approximately 100 km depth (Fig. 1-7), but we find that they vary significantly in shape, size, and occurrence at global and local scales.

In regard to the Kind et al. (2020) suggestion that many prior MLD observations may be artificially created by Moho side-lobes related to deconvolution, we counter this assertion by demonstrating with synthetic seismograms that our use of time domain deconvolution generates virtually no side-lobe energy (Fig. 1-2) and by finding significant, if intermittent, MLD energy in the globally distributed stations of this study. We also note that Eilon et al. (2018) stacked un-deconvolved Sp waveforms, and still found MLDs in some cratonic settings, for example matching our RF results for RF stack RSSD 1 (Fig. 1-6).

We also compared the results of this study to amplitudes in the Sp common conversion point stack of Hopper and Fischer (2018). For six out of nine backazimuthal bins in the northern

Great Plains (Fig. SA-14) where we observe significant negative MLD phases, Hopper and Fischer (2018) also found larger MLD amplitudes, and where we see no resolvable MLD energy, the Hopper and Fischer (2018) amplitudes are small.

4.1.3 Mechanisms for mid-lithospheric velocity gradients

Quantitative evaluation of cratonic lithosphere structure sheds light on trends that potential mechanisms/origins of MLDs must satisfy. These qualifications, described in section 4.1.1, are as follows: (1) upper mantle structure varies between RF stacks sampling different backazimuthal swaths from a single station; (2) negative MLDs are predominately present at ~100 km depth; (3) MLDs are only present in ~50% of cratonic RF stacks ($V_s \geq 4.5$ km/s at 150 km); (4) craton boundaries exhibit shear velocity decreases up to 4%; (5) craton interiors exhibit shear velocity decreases up to 3%; (6) MLD magnitude does *not* depend on thermotectonic age; (7) MLDs have lower magnitude in regions with higher local shear velocity; (8) MLDs are less common in Paleoproterozoic and Archean age regions.

Based on the analysis of the RF stacks in this study, we rule out attenuation due to elastically accommodated grain boundary sliding and layering in degree of melt depletion as likely mechanisms for the MLD phases we observe. Attenuation due to elastically accommodated grain boundary sliding is proposed to cause velocity drops of a few percent in the upper mantle globally (Karato et al., 2015; Karato and Park, 2019). Although we may fail to resolve small MLDs at stations with relatively low data quality (i.e. YKWs 1), we also do not observe MLDs in some high-quality RF stacks (i.e. FFC 1) in which a 2-3% velocity drop should be resolved if it exists (Fig. 1-6). To explore whether some regions of the cratonic mantle might not experience a velocity reduction due to elastically accommodated grain boundary sliding due to lithospheric conditions, we compared ranges of cratonic geotherms based on xenolith data

(Lee et al., 2011) to the full range of temperature and pressure conditions where the attenuation peak that produces the velocity reduction is predicted to occur, given uncertainties in mantle water content, grain size, and activation volume (Karato et al., 2015). This comparison indicates that the velocity reduction should occur in the 50-150 km depth range, and thus appear in our MLD observations, except for the coldest cratonic mantles with conditions at the very dry, large grain size, and high activation volume end of the cases considered by Karato et al. (2015). Since the failure to observe an MLD is common (50%) in the receiver functions presented here and includes regions where mantle velocities are not especially high and temperatures are not particularly cold, we conclude that elastically accommodated grain boundary sliding is not a likely explanation for the data. Furthermore, layering in melt depletion is hypothesized to cause shear velocity changes of 1-1.7% (Afonso & Schutt, 2012), which too small to explain many NVGs we observe at craton boundaries and in craton interiors, although it is possible that layering in depletion could contribute to some of the structures we observe (Fig. 1-8).

In craton interior regions, all shear velocity changes we observe ($< -3\%$) can potentially be explained by anisotropy frozen in the lithosphere (Wirth & Long, 2014). Mineral alignment during deformation has been shown to cause anisotropy in the lithosphere, contributing to MLDs observed at station CCM (Wirth & Long, 2014). The presence of anisotropy is consistent with the variation in lithosphere structure we see from RF stacks of different backazimuthal swaths at the same seismic station (i.e. DBIC) as well as the regional variations we observe, since the amplitude of phases produced by anisotropy depend strongly on mineral orientation relative to backazimuth. While some MLDs likely have contributions from anisotropy, the global trend of MLDs being less common in regions with older thermotectonic ages (Fig. 1-9(d)) is *not* explained by anisotropic MLDs. Unfortunately, a complete analysis of the contribution of

anisotropy to the phases we observe is beyond the scope of this study as the backazimuthal bins for which we have high-quality stacks do not span a sufficient backazimuthal distribution. Additionally, the S-to-P conversion points from different backazimuths measured at a single station are 50-150 km apart, so we cannot differentiate between lateral variations and anisotropic structure.

MLD phases in craton interiors are compatible with metasomatism (Fig. 1-10; Dasgupta & Hirschmann, 2010). Metasomatic fluids in the cratonic lithosphere could have been introduced at ancient convergent boundaries, potentially during amalgamation of the cratonic mantle (e.g. Lee et al., 2011) or after its formation. In particular, the introduction of hydrous silicic or carbonatitic melt into the upper mantle through subduction could result in minerals with relatively low seismic velocities (e.g. Hansen et al., 2015; Rader et al., 2015; Selway et al., 2015). Saha et al. (2021) model the shear velocity drop that can be caused by the presence of metasomatic minerals in the cratonic lithosphere. With typical Archean mantle compositions and geothermal gradients, a layer with ~5 vol% phlogopite and ~10 vol% amphibole can produce shear velocity drops of 1.8-2.38%, which can explain our observations for all MLDs in craton interiors and most MLDs observed at craton boundaries (Fig. 1-8(b,c)).

The creation of MLDs by metasomatism is further supported by a pattern we see in MLD presence with thermotectonic age. A higher occurrence of MLDs in mantle younger than the Paleoproterozoic indicates that either MLDs are formed through processes that are more abundant since the Paleoproterozoic or that there is a process that erases MLDs over time. One mechanism in the former category is that due to higher geothermal gradients during the Archean and Paleoproterozoic, volatiles from the subducting slab may have been released at shallower depths where they were unable to affect large volumes of the upper mantle (Dasgupta &

Hirschmann, 2010; Sun & Dasgupta, 2020). It is possible that this shallower release of metasomatic fluids could result layers of metasomatic material being preserved less often in the upper mantle. The transition of MLDs in only ~25% of Paleoproterozoic RF stacks to ~75% of Mesoproterozoic and younger RF stacks coincides with the global transition from ‘warm’ to ‘cold’ subduction, where the introduction of metasomatic fluids could occur deeper in subduction (Sun & Dasgupta, 2020). Thus the small and moderate amplitude MLDs observed in craton interiors could be remnant of ancient subduction.

We also observe PVGs in the cratonic mid-lithosphere. These PVGs are most pronounced in the lithosphere of the Kaapvaal craton (stations LBTB and BOSA). This result is consistent with other geophysical and geochemical observations from the Kaapvaal craton that indicate that the Kaapvaal mantle contains complexities not widely observed in other Archean cratons (e.g. Text S2.5; Hoal, 2003; Pedersen et al., 2009). For example, the Kaapvaal craton experienced significant Paleoproterozoic volcanism during the emplacement of the Bushveld Igneous Complex, which is the largest layered igneous intrusion in the world (Hoal, 2003; Kgaswane et al., 2012), and station LBTB, the station with the largest PVGs, is adjacent to the Bushveld Igneous Complex. If modeled at face value, the LBTB PVGs, including those at depths greater than 150 km, imply a 10% velocity increase from depths of 130 km to 200 km (Fig. 1-8(d)). The observed PVGs could be due to a phase transition from spinel to garnet peridotite, or an increase in depth of high velocity eclogite and/or carbon preferentially stored in the Archean mantle in the form of diamond (Text S2.5; Aulbach et al., 2020; Dasgupta & Hirschmann, 2010; Garber et al., 2018; Lebedev et al., 2009).

At craton boundaries, we observe some velocity changes ($> -3\%$) that exceed what is possible from metasomatism of a mantle with an Archean bulk composition (Saha et al., 2021).

While the potential for creating larger volumes of minerals such as amphibole and phlogopite from younger cratonic mantle has not yet been evaluated, our results argue against this possibility. The amplitudes of observed MLDs do not correlate strongly with thermotectonic age (Fig. 1-9(c)), suggesting that larger MLDs observed at craton boundaries are *not* due to temporal changes in bulk mantle compositions during craton formation. We do, however, observe a correlation between MLD amplitude and modern upper mantle shear velocity structure (Fig. 1-9(a)), in which proximity to low shear velocity regions corresponds to larger MLD amplitudes. Regions with relatively low shear velocities at 150 km depth are more likely to be located near non-cratonic regions, where the LAB is relatively shallow. Infiltration of metasomatic fluids from adjacent shallow asthenosphere (e.g. Sun & Kennett, 2017; Wolbern et al., 2012) following craton formation, potentially through subduction or rifting processes, is a logical source for large MLDs ($> -3\%$) at craton boundaries (Fig. 1-10). This is further supported by looking at individual cases of these larger MLDs. We observe large negative phases in RFs that sample the Wyoming craton (RSSD 1), Tanzania craton (MBAR 1), the eastern block of the North China Craton (TIA 1, LYN 1), and the Oslo rift (KONO 1), all of which have multi-disciplinary evidence for post-craton formation alteration of the lithospheric mantle (see Results section).

Geochemical data that support the creation of MLDs by metasomatism due to tectonic activity at the margins of existing cratonic mantle lithosphere include an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 1-1.25 Ga for phlogopite from Kaapvaal craton kimberlites (Hopp et al., 2008). This age is consistent with the minerals in the MLD-related layer being generated by metasomatism related to Proterozoic orogenic activity at the margins of the Archean Kaapvaal cratonic mantle.

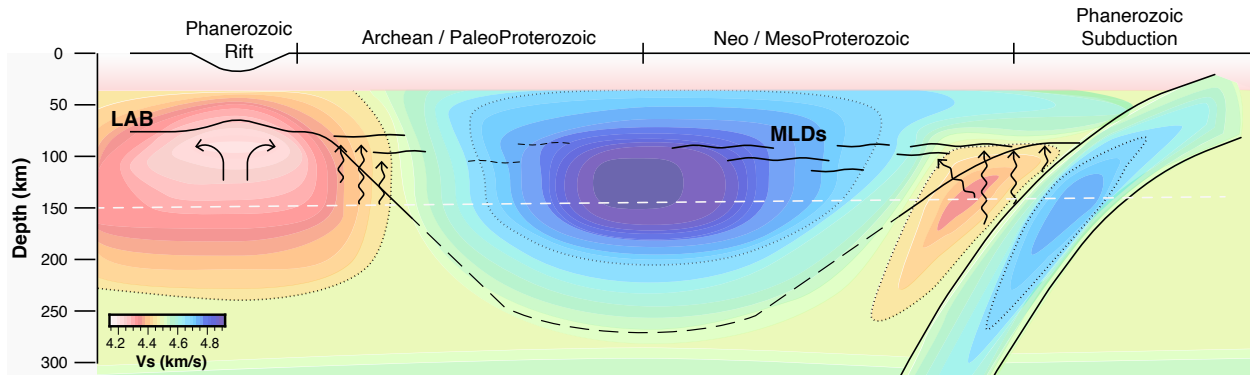


Figure 1-10. Schematic view of conclusions regarding MLD occurrence

Thermotectonic age is labeled at surface, and upper mantle shear velocity is shown by colored contours. Dotted lines indicate the 4.5 and 4.675 km/s shear velocity contours, which are used to bound cratons and craton interiors, respectively, where the contour intersects with 150 km depth (white dashed line). Wavy lines represent MLDs and squiggly lines indicate recent metasomatism.

4.2 The lower-lithosphere and lithosphere-asthenosphere boundary

4.2.1 Global observations of the lower-lithosphere and LAB

At cratonic stations, negative and positive velocity gradients are present in the depth range of 150-300 km. In cratons, small NVGs from 150-300 km exist and are more prevalent at craton boundaries (~20%) than interiors (~5%). We primarily observe positive phases at depths of 220-275 km (Fig. 1-7) in craton boundaries and interiors. The observation of PVGs at this depth is more striking if RF stacks from the Kaapvaal craton are excluded (see section 4.1.2; Text S2.5; Fig. SA-13).

Velocity gradients from 150-300 km also show a pattern with thermotectonic age. PVGs are significantly more common (50%) in stacks that sample regions with Phanerozoic thermotectonic ages, while Proterozoic and Archean regions only have PVGs in 20-30% of RF stacks (Fig. 1-9(d)). Phanerozoic RF stacks indicate no NVGs from 150-300 km. In Proterozoic regions we observe deep NVGs in <20% of RF stacks, while Archean regions have NVGs in this depth range in ~25% of stacks (Fig. 1-9(d)). With more conservative amplitude criteria, we do not observe any NVGs below 200 km depth (Fig. SA-6).

Based on forward modelling results, we find that these lower lithosphere velocity gradients can be due to velocity increases or decreases of 1-2% in most cratonic regions (Fig. 1-8(a-d)). In the Kaapvaal craton, positive phases in the lower lithosphere require up to 10% velocity increases over ~100 vertical kilometers (Fig. 1-8(e)). These strong and distributed PVGs are observed at stations LBTB 1 and SYO 1, and they could be explained by the same mechanisms as PVGs in the mid-lithosphere (sections 4.1.2 and Text S2.5).

4.2.2 Mechanisms for lower-lithospheric velocity gradients

The cratonic LAB, estimated to exist at depths greater than 150 km (e.g. Dalton et al., 2017; Fischer et al., 2020), is one possible explanation for the weak NVGs observed in this depth range. We find that negative phases in this depth range are absent (with one exception) in non-cratonic regions, presumably because the lithosphere is thinner. Lower-lithosphere NVGs are more prevalent at craton boundaries (~20%) than interiors (~5%), consistent with the idea that proximity to regions thinner lithosphere and shallower asthenosphere can produce a larger velocity contrast at the cratonic LAB. The NVG at ~245 km depth at station MBAR is one such example (Fig. SA-5(b); 1-6(b)). Despite infiltration of metasomatic fluids from the East African Rift, some of the altered cratonic keel of the Tanzanian craton remains intact (Celli et al., 2020; Wolbern et al., 2012), and it is possible that the ~245 km depth NVG is remnant of the cratonic LAB of the Tanzanian craton.

The NVGs we observe at depths greater than 150 km can be modelled by shear velocity decreases of 1-2% (Fig. 1-8(a,d)), indicating that the cratonic LAB could be the juxtaposition of the depleted lithosphere over fertile asthenosphere (Afonso & Schutt, 2012). Deep NVGs that indicate velocity reductions exceeding ~2% require additional compositional changes (Hopper & Fischer, 2018) or an anisotropic contribution from the transition from asthenospheric flow to

static (in a relative sense) lithosphere (Gung et al., 2003; Vinnik et al., 2005). In regions with RF stacks that do not contain a deep NVG, it is possible that refertilization at the base of the lithosphere (Lee et al., 2011) has caused the transition from depleted to fertile mantle to be gradual enough that an observable NVG is not produced.

Previous studies have also observed negative phases which they interpret as cratonic LABs at stations HYB and RSSD. A negative phase is observed in RF stack HYB 2 at ~180 km depth and similar phase is observed at HYB 1 just above ~150 km depth (Fig. 1-6(b)). These observations coincide with the results of Mitra et al. (2006), who observe a phase at ~155 km at station HYB which they interpret as the LAB in the Dharwar craton. Additionally, Mancinelli et al. (2017) observe a NVG they interpret as the cratonic LAB at RSSD at ~200 km depth, corresponding to our results in RF stack RSSD 1 (Fig. 1-6(b)).

We also observe PVGs in the depth range 150-300 km in many RF stacks that sample craton interiors and boundaries. This PVG may represent the Lehmann Discontinuity (Lehmann, 1959) which has been proposed to be (1) anisotropy at the cratonic LAB (Gung et al., 2003), (2) the bottom of a low velocity layer (Vinnik et al., 2016), or (3) the top of a high-velocity layer, possibly within the cratonic mantle lithosphere (Dalton et al., 2017; Eeken et al., 2018; Garber et al., 2018). The existence of eclogite xenoliths as well as diamond xenocrysts within kimberlites and xenoliths provide an explanation for a layer of high seismic velocity in the cratonic lithosphere (Garber et al., 2018).

5 Conclusions

Through meticulous processing, we resolve previously undetected patterns in upper mantle velocity gradients using Sp RF stacks. Regions with cratonic crustal geology that lack an underlying cratonic lithosphere are characterized by large negative amplitude MLDs below the

Moho that we interpret as the LAB. In regions with thick cratonic lithosphere, low amplitude negative lithospheric velocity gradient(s) are observed in ~50% of sampled locations, indicating these MLD phases are common but intermittent. The negative lithospheric velocity gradients are most easily explained by the presence of volatile-rich minerals produced by metasomatic processes, although vertical gradients in anisotropy cannot be ruled out as an alternative origin. MLDs that correspond to shear velocity decreases of less than 3%, including all cases in craton interiors, could represent metasomatism of lithosphere with an Archean bulk composition (Dasgupta & Hirschmann, 2010; Saha et al., 2021), although this condition is not required. Negative velocity gradients in the shallow lithosphere that indicate shear velocity decreases of more than 3% are likely due to interactions of the cratonic lithosphere with the asthenosphere following craton formation. An increase in the observation of MLDs in RF stacks from ~30-55% to 70% at the end of the Paleoproterozoic corresponds with the shift from ‘warm’ to ‘cold’ subduction.

We also observe shear velocity changes from 1-2% at 150-300 km depths. Velocity decreases in this depth range can be attributed to the layering of depleted lithosphere over fertile asthenosphere at the cratonic LAB. Positive velocity gradients at lower lithospheric depths could represent the Lehmann discontinuity, and could be attributed to the top of a higher-velocity layer (Dalton et al., 2017; Eeken et al., 2018; Garber et al., 2018). The presence of both a shallow low velocity MLD and a deeper positive gradient corresponds with Rayleigh wave phase velocity measurements in cratons (Dalton et al., 2017; Eeken et al., 2018; Garber et al., 2018).

7 Acknowledgements and Data

We thank Zachary Eilon, Junlin Hua, and Colleen A. Dalton for their useful advice regarding methodology and interpretations. We also thank Dr. Ebru Bozdog for providing us

with the GLAD-M15 model. This work was funded by the U.S. National Science Foundation's Graduate Research Fellowship Program and the Cooperative Studies of the Earth's Deep Interior Program (EAR-1763243). Generic Mapping Tools is used to plot figures in the manuscript and supplementary information (Wessel & Smith, 1991).

Seismic waveform data were acquired from the Integrated Research Institutions in Seismology Data Management Center (IRIS DMC). These data are from the following networks: the IRIS/USGS Global Seismic Network (II: <https://doi.org/10.7914/SN/II>, IU: <https://doi.org/10.7914/SN/IU>); the Global Telemetered Seismograph Network (GT: <https://doi.org/10.7914/SN/GT>); GEOSCOPE (G: <https://doi.org/10.18715/GEOSCOPE.G>); the United States National Seismic Network (US: <https://doi.org/10.7914/SN/US>); the Canadian National Seismograph Network (CN: <https://doi.org/10.7914/SN/CN>); GEOFON (GE: <https://doi.org/10.14470/TR560404>); the Australian National Seismograph Network (AU); the Central and Eastern US Network (N4: <https://doi.org/10.7914/SN/N4>).

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Chapter 2 A Common Origin for Layering in the Cratonic Lithosphere

Hannah E. Krueger¹, Karen M. Fischer¹, Eva Golos², & Isabella Gama¹

¹ Department of Earth, Environmental, and Planetary Sciences, Brown University, 324 Brook St, Providence, RI 02912

² University of Wisconsin – Madison

Key Points:

- The presence of a low velocity layer with a 1-3% velocity reduction is needed to explain the velocity structure of approximately one half of the locations we analyze
- 1-3% diamond can explain velocity increases in the lithosphere in many cratons globally, particularly those without thinned lithospheres.
- Sublithospheric infiltration of carbonate melts explains the common observation of velocity increases in the lower lithosphere of cratons as well as the intermittent observation of overlying low velocity layers

Abstract

Layering in the seismic structure of the cratonic lithosphere, and its origins, are debated. Some prior work considers similar layering to be ubiquitous in cratons globally, while other work observes significant heterogeneity in cratonic structure. To address this debate, we resolve shear wave velocity structure for the upper 250 km of the mantle at 42 locations globally. Of these, 39 locations are underlain by thick lithosphere extending below 150 km depth. Within these profiles we observe a wide variety of lithospheric structure that varies on local and global scales. We model simple layering in a peridotite mantle with a single velocity increase and/or a single upper lithosphere low velocity layer. We explain our models with three levels of complexity: (1) peridotite-only compositions, (2) peridotite with a single mantle velocity increase or a low velocity layer in the uppermost mantle, and (3) a combination of a single velocity increase underlying a low velocity layer. With this scheme we adequately model 28 of our shear velocity profiles. All velocity increases we model are explained by ≤ 3 vol% diamond with some contributions from eclogite. All low velocity layers we model are explained by a $\leq 3\%$ reduction in velocity from surrounding peridotite which can be explained by the hydration of minerals within the Archean mantle. The infiltration of carbonate melt from asthenospheric upwellings offers a common origin for the deposition of diamond in the lower lithosphere and the presence of hydrous minerals in the mid-lithosphere.

1 Introduction

Cratons compose the tectonically stable cores of continents. In general, the uppermost mantle of cratons consists of thick, seismically fast, and iron-depleted lithosphere with cool geothermal gradients. The timeline of and processes involved in the formation and evolution of cratons remain unclear (e.g. Lee et al., 2011). The cratonic lithosphere has experienced large-scale tectonic stability since the Precambrian, although infiltration of melt from nearby tectonic activity or underlying mantle upwellings is evident, if not pervasive, in most cratons (e.g. Jackson & Gibson, 2023).

Layering in seismic velocity within the cratonic lithosphere has been observed in many regions around the globe. For example, negative velocity gradients and/or low velocity layers in the upper lithosphere of many cratons have been imaged in numerous studies of converted and reflected body waves (e.g. Abt et al., 2010; Ford et al., 2010; Geissler et al., 2010; Hansen et al., 2015; Hopper & Fischer, 2015; Hopper & Fischer, 2018; Kind & Yuan, 2018; Kind et al., 2012; Krueger et al., 2021; Liu & Shearer, 2021; Miller & Eaton, 2010; Rychert & Shearer, 2009; Sun & Kennett, 2017; Wirth & Long, 2014; Wolbern et al., 2012) and joint inversions of body and surface wave data (e.g. Calo et al., 2016; Eilon et al., 2018). Geophysical studies have also found interfaces within the cratonic lithosphere with contrasting seismic anisotropy and electrical conductivity (Chen et al., 2009; Eilon et al., 2018; Selway et al., 2015; Sodoudi et al., 2013; Wirth & Long, 2014).

A negative velocity gradient in the upper lithosphere (Hopper & Fischer, 2018; Rader et al., 2015; Selway et al., 2015), often referred to as a “mid-lithospheric discontinuity” (Abt et al., 2010) is commonly enough observed that a negative velocity gradient, thought to be the top of a low velocity layer, is considered to be a ubiquitous feature of the cratonic lithosphere by some

studies (e.g. Fu et al., 2022; Yang et al., 2023). However, the nature, origin, and ubiquity of this layering is debated (e.g. Karato & Park, 2018; Kind et al., 2020; Krueger et al., 2021). The mechanisms proposed to explain this feature include layering in melt depletion, the presence of hydrous minerals (e.g. Hansen et al., 2015; Liu et al., 2018; Rader et al., 2015; Saha et al., 2021; Selway et al., 2015) or melt (Dasgupta, 2018; Thybo & Perchuc, 1997), seismic anisotropy (Wirth & Long, 2014), and attenuation due to elastically accommodated grain boundary sliding (Karato, 2012; Karato & Park, 2018). One test of these mechanisms is whether they predict that a mid-lithospheric discontinuity (or “MLD”) is present everywhere in the cratonic mantle lithosphere. Some recent studies have found MLD structures to be spatially intermittent (e.g. Hopper & Fischer, 2018; Krueger et al., 2021; Liu & Shearer, 2021), pointing to a compositionally heterogeneous cratonic upper mantle (e.g. Altoe et al., 2020; Eeken et al., 2018), while others have concluded that MLD structures are nearly continuous (Liu & Gao, 2018), or have suggested it does not exist (e.g. Kind et al., 2020). Another test of these mechanisms is the magnitude of the velocity reductions that they can produce (e.g. Karato & Park, 2018; Rader et al., 2015; Saha et al., 2021; Selway et al., 2015). Examples of typical MLD-related velocity reductions in recent studies include 5-6 % (Hopper & Fischer, 2018) and 1-4% (Krueger et al., 2021).

An additional puzzle about the cratonic mantle lithosphere are the high seismic velocities apparent in the lower lithosphere of cratons (e.g. Dalton & Faul, 2010; French & Romanowicz, 2014; Lekic & Romanowicz, 2011; Schaeffer & Lebedev, 2013), which in many cases exceed the shear velocity ranges predicted for peridotite even assuming cool cratonic geothermal gradients (Dalton et al., 2017; Dalton & Faul, 2010; Garber et al., 2018). Previous work has

proposed these high velocities could be explained by the presence of intrinsically high velocity materials such as diamond or eclogite or by anisotropy (Garber et al., 2018; Gung et al., 2003).

To investigate the character and origins of layering in the cratonic lithosphere, we solve for absolute shear velocity and discontinuity structure at 39 different cratonic locations globally (Fig. 2-1) using a Bayesian joint inversion of S-to-P receiver functions (Sp RFs) and Rayleigh phase velocities (Eilon et al., 2018; Gama et al., 2021). Layering in the cratonic lithosphere is commonly observed with methods that solve for velocity gradients, but interpretation of velocity gradients leaves significant ambiguity for interpreting potential sources of velocity contrasts, such as mineralogy. The information provided with both the absolute shear velocity structure and the discontinuity structure allows us to thoroughly investigate the discrepancies between previously proposed mechanisms for velocity discontinuities within the cratonic lithospheric mantle. By jointly inverting Sp RFs (Krueger et al., 2021) and Rayleigh phase velocities (Ekstrom, 2011), we achieve the following goals:

- (1) Solve for the shear velocity structure of the uppermost cratonic mantle (Moho – 250km).
- (2) Characterize global distributions of velocities within the cratonic mantle that cannot be explained by simple peridotite mantle compositions.
- (3) Offer possible explanations for the high and low cratonic shear velocities we observe.

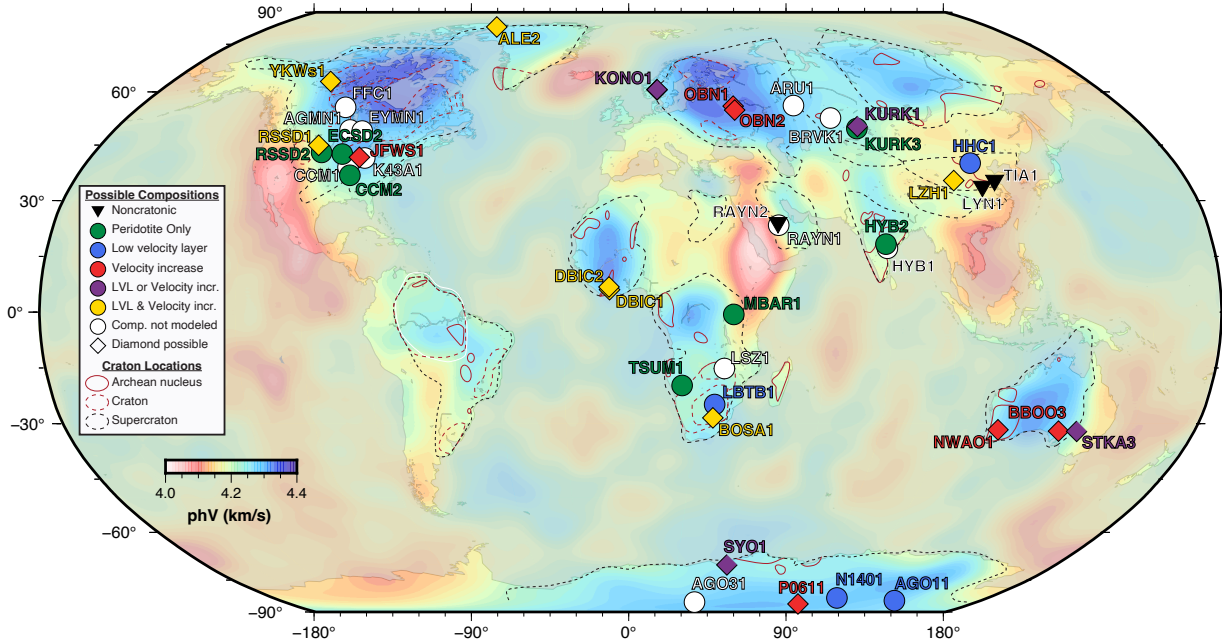


Figure 2-1. Global distribution of Vs profiles

Global distribution of the locations of the Vs profiles from this study plotted with phase velocity from GDM52 at 125s (Ekstrom, 2011). Shape and color indicate the possible compositions at each location (see section 4.1). Diamonds indicate locations where Vs structure does require the presence of diamond in the lithosphere. The Vs structure at each location was compared to peridotite (green). If a Vs profile is not adequately modeled with only peridotite, additional velocity increases (red) and low velocity layers (blue) are added to peridotite mantles. Some profiles could be satisfied with either a velocity increase or a low velocity layer (purple). Some Vs structures we observe are complicated and require both a low velocity layer and a velocity increase within the lithosphere (yellow) or cannot be effectively modeled with this simple structure and limited band of compositions (white). Craton locations are outlined based on Woolley and Kjarsgaard (2008) (Figure 24).

2 Data & Methods

2.1 Body wave data

We use 42 Sp receiver functions (Sp RFs) from Krueger et al. (2021) calculated at long-running broadband seismic stations located in a variety of cratons across the globe (Fig. 2-1). We note that in this context, we consider stations from locations with high velocity lithosphere at 150 km depth (e.g. French et al., 2013) or Paleoproterozoic or older thermotectonic age (Artemieva, 2006). With this classification, we include stations overlying Phanerozoic thermotectonic age material that exhibit characteristics of cratonic lithosphere (Fig. 2-1, KURK1). We also include stations at locations with Archean thermotectonic ages but overly slow upper mantle (Fig. 2-1,

LYN1, TIA1). Traces containing Sp phases are separated into bins spanning 30° in backazimuth to isolate lateral variations in upper mantle structure. S-to-P conversions cause the ray path to flatten resulting in upper mantle conversions that may occur 10s of kilometers lateral distance from the station and 100s of kilometers from conversions with opposing backazimuths. Due to S-to-P ray paths, phases observed in an Sp RF are more representative of the structure below the 100 km conversion point location of an S-to-P phase than the structure directly below the corresponding station (Fig. 2-2). We consider Vs profiles resulting from inversions of these S-to-P receiver functions to best represent the location of the 100 km S-to-P conversion point found with the TauP Toolkit (Fig. 2-2) (Crotwell et al., 1999).

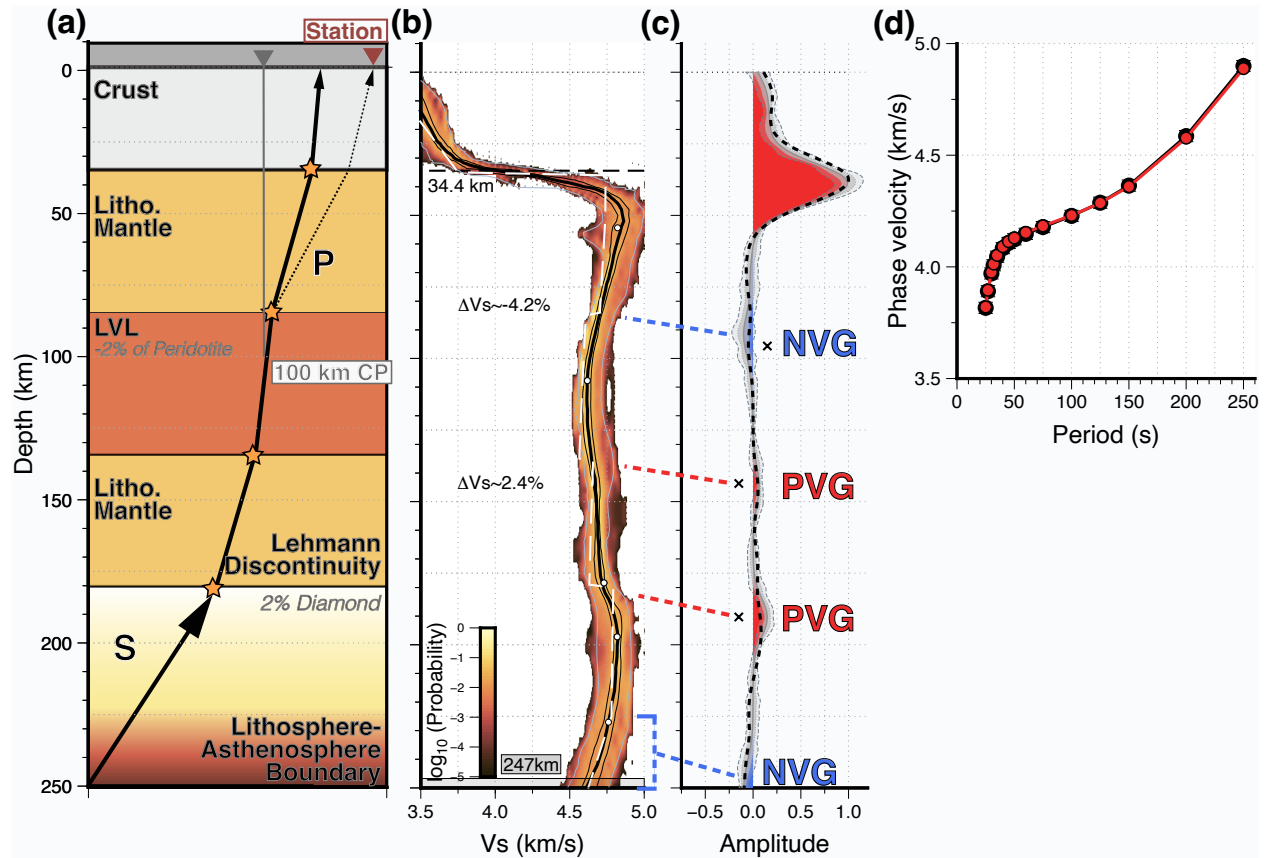


Figure 2-2. Example of inversion with synthetic data

(a) A schematic of potential craton structure, including the crust-mantle boundary, a low velocity layer (LVL) within the cratonic lithosphere, a velocity increase in the lower lithosphere due to 2% diamond in peridotite, and a gradual transition from cratonic lithosphere to underlying lower-Vs asthenosphere (225-265 km). A star indicates the location at which an S-to-P conversion from the top of a low velocity zone at 100 km depth which would then be measured at the station (red inverted triangle). The gray inverted triangle indicates the projected surface location of this conversion point, demonstrating our reasoning for using Rayleigh phase velocities at the location of this conversion point instead of the station location. (b) Input Vs profile (white dashed line) used in synthetic inversion, calculated assuming a 35 mW/m² geotherm and the 2% low velocity layer and 2% diamond shown in (a). The posterior probability (color scale) and final Vs model (black dash) of an inversion of synthetic Sp RF and Rayleigh phase velocity data generated for the input Vs profile. Gray lines show one standard deviation. Horizontal black line is the estimated Moho depth. Small white circles indicate changes in velocity gradient between which velocity changes are calculated, with changes $\geq 2\%$ annotated. (c) Synthetic Sp RF (black-dashed) from the model result compared to “observed” synthetic Sp RF data from the input model (red/blue). Polarity has been flipped to match the Ps convention. (d) Synthetic phase velocities from the model result (red) compared to “observed” synthetic Rayleigh phase velocities (black). Synthetic phase velocities are from the Mineos algorithm (Masters et al., 2007).

In this study, we re-process Sp RFs from Krueger et al. (2021) with data up to July 2022. Parent, SV, and daughter, P, components for individual events are deconvolved iteratively in the time domain (Hua et al., 2020; Ligorria & Ammon, 1999). Individual Sp RFs undergo a rigorous culling and stacking procedure to minimize processing artefacts and confidently resolve phases

due to velocity gradients in the upper mantle (Krueger et al., 2021). We have updated the culling procedure from Krueger et al. (2021) to include hierarchical clustering as an initial data culling step prior to the weighting algorithm from Krueger et al. (2021) which iterates K-means clustering to weight individual Sp RFs before final bootstrap stacking. Bootstrapped data error (Fig. 2-3, grey shading on RF) is included in the joint inversion through the Sp RF misfit calculation to prevent overfitting to uncertain phase amplitudes.

While Ps RFs are calculated for a majority of the stacks we invert (Krueger et al., 2021), we do not incorporate Ps RFs into our final inversions because some regions have Ps RFs that are not compatible with their Sp counterparts due to complicated and laterally variable crustal structures. We neglect Ps RFs for two primary reasons: (1) reverberations due to overlying ice or sedimentary basins can complicate the apparent depth and amplitude of Moho phase arrivals and (2) regions with laterally variable Moho features (depth & velocity change) can have different Moho characteristics in Sp & Ps RFs due to differences in the incoming angle of Ps and Sp converted phases. Conversion points from Ps Moho and crustal phases can be due to structure 10s of kilometers lateral distance from Sp conversion points of the same stack. Thus, while Ps RFs in regions with simple crustal seismic velocity structure and consistent crustal thickness help to constrain Moho characteristics in this type of joint inversion (Eilon et al., 2018), the inclusion of Ps RFs in inversions for regions with more complicated crustal structure can introduce inconsistencies into the inversion results.

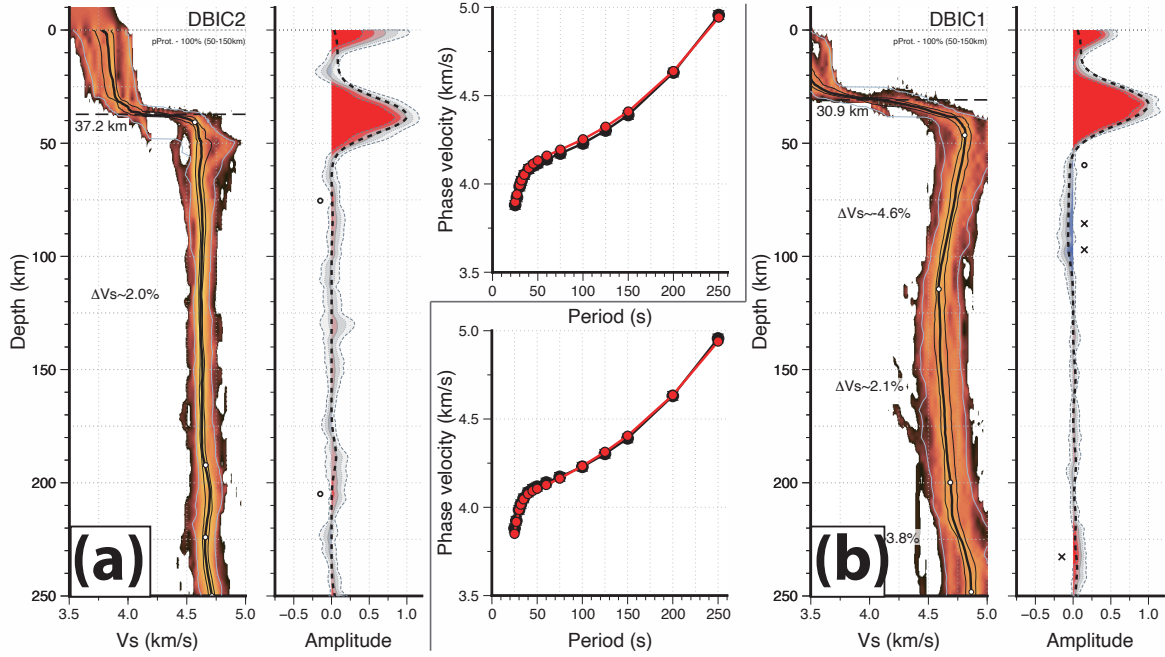


Figure 2-3. Examples of joint inversions with real data

Median Vs profiles (bold black line) and posterior probabilities (yellow-red colormap) and comparisons of observed and predicted receiver functions and phase velocities for cases with (a) negligible velocity gradients and (c) significant velocity gradients. These Vs models correspond to a single station (DBIC) but different backazimuthal bins, exemplifying lateral variation within cratonic regions. Phase velocities from GDM52 (black circles, where uncertainties are approximately the size of the symbols) are compared to the median model predictions (red circles). Receiver function data (red and blue; Krueger et al., 2021) with 1- and 2- standard deviations (gray) plotted against synthetic data from the median Vs model (black dotted). The probability density function (pdf) for the models plotted in the background indicates the consistency of individual model chains. Even broader pdfs still produce median models that fit body and surface wave data well.

2.2 Surface wave data

We carried out one set of inversions using fundamental mode Rayleigh wave phase velocities from the global model GDM52 (Ekstrom, 2011) at periods of 25-250s (Fig. 2-1, background) (Ekstrom, 2011). We use Rayleigh wave phase velocities at location of the 100 km S-to-P conversion point of the corresponding S-to-P receiver function (Fig. 2-2) (Crotwell et al., 1999). This assumption is especially important in regions with laterally variable mantle structure (Fig. 2-3).

The global GDM52 model (Ekstrom, 2011) allows for consistent comparisons between stations distributed across the globe. However, we also recognize that in many regions higher-resolution phase velocity data sets exist, but using phase velocity data from separate studies may introduce inconsistencies in our Vs profile results between separate regions. To address this, we provide additional velocity models obtained with Rayleigh phase velocities calculated by regional studies in the supplementary materials for stations in Southern Africa (Malory et al., 2022; White-Gaynor et al., 2021), the United States (Babikoff & Dalton, 2019; Herrmann et al., 2021), and Antarctica (Heeszel et al., 2016; Shen et al., 2018). We use these inversions to validate and evaluate the limitations of Vs profiles produced with GDM52 (Fig. 2-4 to 2-7, aquamarine-dashed lines).

Rayleigh phase velocities at shorter periods from ambient noise tomography (ANT) provide constraints on crustal and uppermost mantle velocity structure, but, to our knowledge, no global model of Rayleigh phase velocities at short periods (<20 s) exist. Therefore, we do not supplement phase velocities from GDM52 with phase velocities from ANT. The lack of shorter period information results in poorly constrained crustal velocity structure in our Vs models. Due to these limitations, we only interpret the structure of the mantle lithosphere in this study. In regions where we also inverted regional surface wave data, we combined Rayleigh phase velocities from both earthquake and ambient noise tomography. These data sets are averaged at overlapping periods. Inversions with ANT data included show significantly better constrained crustal velocity structures.

2.3 Inversion Methodology

We apply the transdimensional hierarchical Bayesian inversion method previously used in Eilon et al. (2018) and Gama et al. (2021) to jointly invert Sp RFs and Rayleigh wave phase velocities

for shear wave velocity. Bayes' theorem (eq. 1) is used to evaluate the probability of a proposed Earth structure at a given location considering observed seismic data from that location. This posterior probability, $p(m|d_{obs})$, is calculated using the prior probability, $p(m)$, of model parameters and a likelihood function, $p(d_{obs}|m)$, found using the misfit, Φ , between forward modeled and observed data (eq. 2).

$$p(m|d_{obs}) \propto p(m) \times p(d_{obs}|m) \quad \text{eq. 1}$$

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

Where the subscript i indicates an individual data type (i.e. Sp RF or Rayleigh phase velocities), σ is the data error hyperparameter, and n is degrees of freedom (number of frequencies for surface waves, first zero-crossing of data auto-correlation function for body waves).

The Bayesian framework allows us to easily incorporate multiple data types into the inversion by simply multiplying the likelihood of each data type. To constrain both absolute shear velocity and discontinuity structure in the upper mantle, we combine Rayleigh wave phase velocities, and Sp receiver functions, respectively (eq. 3). Data misfits are fully explained in section 2.3.2 and 2.3.3.

$$p(d_{obs}|m) = p(d_{SpRF}|m)p(d_{SW}|m) \quad \text{eq. 3}$$

We use a Monte Carlo Markov Chain (MCMC) algorithm to randomly explore the model parameter space. At each iteration, we calculate misfit to observed data from the predictions of the proposed Earth model obtained by forward modeling. The goals of decreasing misfit between predicted and observed data and a thorough search of model parameter space are attained by using a Metropolis Hastings acceptance criterion to evaluate the model tested in each iteration (Hastings, 1970). The method results in a posterior probability distribution from an ensemble of

acceptable models. The median of the acceptable shear-velocity models determines a smooth 1D shear velocity profile representing the shear velocity structure adjacent to each station in the direction of the Sp RF binned backazimuth (Fig. 2-1, 2-2). Gama et al. (2021) and Gama et al. (2022) use this methodology to clearly characterize the LAB in both tectonically active (southern) and stable (northern) regions of Alaska. The LAB depths they observe are within the depth range where we expect to resolve velocity gradients due to MLDs based on the depths of phases observed in Sp RFs (Gama et al., 2022; Gama et al., 2021; Krueger et al., 2021).

2.3.1 Model Parameterization

Following Eilon et al. (2018), Gama et al. (2022), and Gama et al. (2021) we use adaptive discontinuous cubic B-spline functions to parameterize smooth seismic velocity profiles. The crust and mantle are represented by separate spline functions with the depth of the sharp Moho velocity contrast as an additional model parameter to define the transition between these functions. The coefficients of each spline as well as the depth and number of spline knots are model parameters that shape the resulting velocity profiles. The relative uncertainty of each data type is also solved for as a hyperparameter that controls the weight of each data type in the likelihood term of Bayes' Theorem (eq. 2).

A variety of limitations are placed on both the inversion parameters and resulting acceptable velocity profiles. (1) A monotonic increase of seismic velocity in the crust is required, while both negative and positive velocity gradients are allowed in the mantle. (2) The crust contains one V_p/V_s value which is solved for as a model parameter. (3) Velocity gradients at the Moho are required to be positive and cannot exceed a 30% change in velocity. (4) Strong positive velocity gradients are not allowed on either side of the Moho to isolate the Moho velocity gradient to the location of the Moho. (5) Spline knots are allowed to exist for the full

depth of the model (300 km) to resolve sharp velocity gradients throughout the upper mantle but must be spaced at least 2 km in depth from one another. (6) The V_p/V_s ratio within the mantle as well as mantle and crustal density are determined using the AK135 reference model (Kennett et al., 1995), the scaling of Brocher (2005), and Abers and Hacker (2016), respectively.

2.3.2 Misfit calculation for body waves

Synthetic S-to-P converted phases are calculated with the propagator matrix algorithm (Keith & Crampin, 1977a, 1977b, 1977c) using a source period of 2 s and ray parameter equal to the mean ray parameter of the data stacked in the Sp RF data. Synthetic traces are processed in the same method as the real Sp RFs (Krueger et al., 2021), including a Gaussian shaping function with a 0.9s half-width that is applied after deconvolution. This shaping function acts as a filter, helping to equalize the frequency content of the observed and synthetic RFs. In a previous application of this methodology, Eilon et al. (2018) chose to use undeconvolved S-to-P converted phases for both the real and synthetic data due to fear of introducing deconvolution processing artefacts in the observed data and to increase computational efficiency during forward modeling in MCMC iterations. Using the methods of Krueger et al. (2021) for processing Sp RFs, Gama et al. (2021) subsequently used deconvolved observed S-to-P converted phases with minimal processing artefacts, but continued to use undeconvolved synthetic data in forward modeling for computational efficiency. We find that using deconvolution for both real and synthetic data significantly improves final fits to body wave data, despite decreasing the computational efficiency of the inversion (Eilon et al., 2018; Gama et al., 2021).

While deconvolving both the modeled and observed Sp RFs makes their time series more compatible, the observed Sp RFs often contain more high frequency energy than the synthetic Sp RF, which is likely due to defining the observed RF by the median amplitude of the RF stack. To

better equilibrate frequency content when calculating misfit, we retain the practice of computing cross-convolutional misfits instead of simply taking the difference between the two RFs (eq. 4) (Eilon et al., 2018). Comparing the cross-convolution of the parent (daughter) of the observed data and the daughter (parent) of the predicted data accounts for any remaining frequency content differences between the two RFs (Eilon et al., 2018). The body wave misfit is the l^2 -norm of the differences in cross-convolution functions, using the window from -1.5 to -40 s in the daughter component (the Sp RF) and 10 to -29.5 s in the parent component (the auto-deconvolved S phase). These time windows are chosen to isolate the Moho and upper mantle phases of interest from non-zero energy at 0 s in the Sp RF while maintaining the same length time window in the parent auto-deconvolved trace. The observed data is also normalized to a parent amplitude of one and so that the largest phase in the daughter Sp RF (presumably the Moho phase) also has an amplitude of one.

$$\Phi_{SpRF} = \left\| \frac{1}{\epsilon_{SpRF}} ([parRF(t, m) * SpRF(t)] - [SpRF(t, m) * parRF(t)]) \right\|^2 \quad \text{eq. 4}$$

$$\epsilon_{SpRF} = \epsilon * parRF(t, m) \quad \text{eq. 5}$$

We incorporate the Sp RF bootstrapped data error, ϵ , (Krueger et al., 2021) into the inversion framework while calculating the body-wave cross-convolution misfit, Φ_{SpRF} (eq. 4). To make Sp RF error compatible with the cross-convolution misfit, we convolve the data error with the modeled parent phase (eq. 5). The addition of the Sp RF error causes times with large error to have a lower weight in the misfit than times with small error. This misfit (eq. 4) is then used in eq. 2 for calculating the likelihood that the modeled Earth structure could be described by the observed Sp RF data.

2.3.3 Misfit calculation for surface waves

We forward calculate synthetic Rayleigh wave phase velocities for all periods corresponding to GDM52 data (Fig. 2-3) using the Mineos algorithm from Computational Infrastructure for Geodynamics (Masters et al., 2007). Mineos requires velocity information for the full radius of the Earth for Rayleigh wave phase velocity calculations, which we represent using the current velocity model until 300 km depth, a gradient between the current model and PREM from 300-400 km depth, and PREM below 400 km depth (Dziewonski & Anderson, 1981). Following (Eilon et al., 2018), we periodically calculate V_p , V_s , and density perturbation kernels which are integrated over depth to determine Rayleigh phase velocity. This is preferred to calculating Rayleigh phase velocities directly from Mineos at each iteration for computational efficiency. The surface wave misfit is the l^2 -norm of the differences in the observed and forward modeled phase velocities across all observed frequencies (eq. 6). This value is used to calculate the likelihood that the modeled Earth structure is being observed by the given Rayleigh phase velocities.

$$\Phi_{sw} = \|C(f) - c_p(f, m)\|^2 \quad \text{eq. 6}$$

The publicly available GDM52 phase velocities (Ekstrom, 2011) does not provide data uncertainties, so we do not incorporate observed surface wave data uncertainties into our inversions with global Rayleigh phase velocities.

2.3.3 Exploration of parameter space & final ensemble of models

A Monte-Carlo Markov Chain (MCMC) with a Metropolis Hastings acceptance criteria (Hastings, 1970) is used to explore the model space for 20,000 iterations. Parameters in the starting model are randomly chosen from the prior distributions. These distributions are roughly uniform within the ranges specified in Table S1. Subsequent iterations randomly choose a single

model parameter to perturb. The perturbation is chosen randomly from a Gaussian distribution (Table S1). An initial burn-in period stabilizes model parameters in likelihood until 8,000 iterations. During this burn-in period, parameters are allowed to vary more when perturbed and are more likely to be accepted than during subsequent iterations. This allows a fuller exploration of the model space and avoids parameters from being limited by local minima in misfit (Eilon et al., 2018). Following the burn-in period, models are saved every 25 iterations to contribute to the final ensemble of acceptable models for this individual chain. This sequence is performed for six independent chains in parallel. At the end of the inversion, all chains with acceptably low RMS misfits for both Sp RF and Rayleigh phase velocity data (within 1.3 of the mean RMS misfit of chains) are considered for the final ensemble for acceptable models. If there are fewer than four chains remaining for the final model ensemble, the unacceptable chains are re-calculated with new random starting model parameters. The posterior distribution of each accepted chain is summed for the final probability distribution. To reduce the influence of outlying models, the median shear velocity of all models in the final model ensemble defines the final shear velocity profile (as opposed to the mean). Median values of crustal V_p/V_s and Moho depth from all models are also chosen.

3 Results

When categorizing the results of the inversions for shear velocity, a first step is to determine which shear velocity profiles are consistent with expectations for intact cratonic mantle based on global averages. To evaluate the extent of cratonic lithospheric mantle in each profile, we compare velocity profiles to the median velocities in the SEMum2 global velocity model (French et al., 2013) for Archean and Paleoproterozoic regions identified from Artemieva (2006). If the two-sigma bounds of a V_s profile overlap with the median absolute deviation V_s bounds for

Archean and Paleoproterozoic regions for at least 50% of the depth range 50-150 km, we consider the uppermost mantle sampled by the Vs profile to be cratonic lithosphere (Fig. SB-2). Vs profiles that are not considered to have cratonic lithosphere are shown in Figure 2-7(b). For Vs profiles with cratonic lithosphere (Fig. 2-4 to 2-6), we identify an LAB depth which represents the maximum depth of mantle we interpret as cratonic lithosphere and shade all depths we consider asthenosphere in gray (Fig. 2-4 to 2-6). We also identify distinct lithospheric velocity gradients in each profile (Fig. 2-4 to 2-6, white circles and ΔV s annotations). Full explanation of LAB and velocity gradient identification are provided in supplementary material section 1.2.

3.1 Lithosphere structure

3.1.1 Cratonic lithosphere

Out of 42 Vs profiles calculated with global Rayleigh phase velocities and Sp RFs, we classify 39 as representative of the cratonic uppermost mantle. We observe significant variation in lithosphere structure (Fig. 2-4 to 2-6). Overall, we observe 22 of the 39 cratonic profiles to have PVGs $>2\%$ in the lithosphere, excluding PVGs that are a result of a poorly resolved Moho depth. We also observe significant low velocity gradients in the upper- and mid-lithosphere (Moho – 150 km) in 19 profiles (Fig. 2-4 to 2-6 annotations). We acknowledge that crustal Vs and the sharpness of the crust-mantle transition are poorly constrained in the global inversions due to the lack of the shorter period phase velocities from ambient noise tomography. We are therefore careful to not interpret PVGs that are an artifact of poorly constrained crust-mantle transitions as velocity gradients in the mantle lithosphere by avoiding PVGs that exist completely within 25 km of the Moho depth (Fig. 2-4 to 2-6, dashed horizontal black line).

In addition to individual velocity gradients, we see a wide array of overall structure. PVGs often appear in the mid- or lower-lithosphere, and many of these PVGs do not appear to be the bottom of a low velocity layer because they have no NVG overlying them; in nine of these cases the PVG is $> 2\%$, and in four $< 2\%$. Numerous PVGs that are $> 2\%$ (13/39) have overlying NVGs, and these gradients appear to represent the upper and lower boundaries of a low velocity layer. Two more profiles appear to exhibit this structure, but without VGs $> 2\%$. Velocity decreases without underlying PVGs appear in 4/39 profiles. Additionally, 7/39 profiles lack sharp lithospheric velocity gradients altogether. These observations are summarized in Table S2.

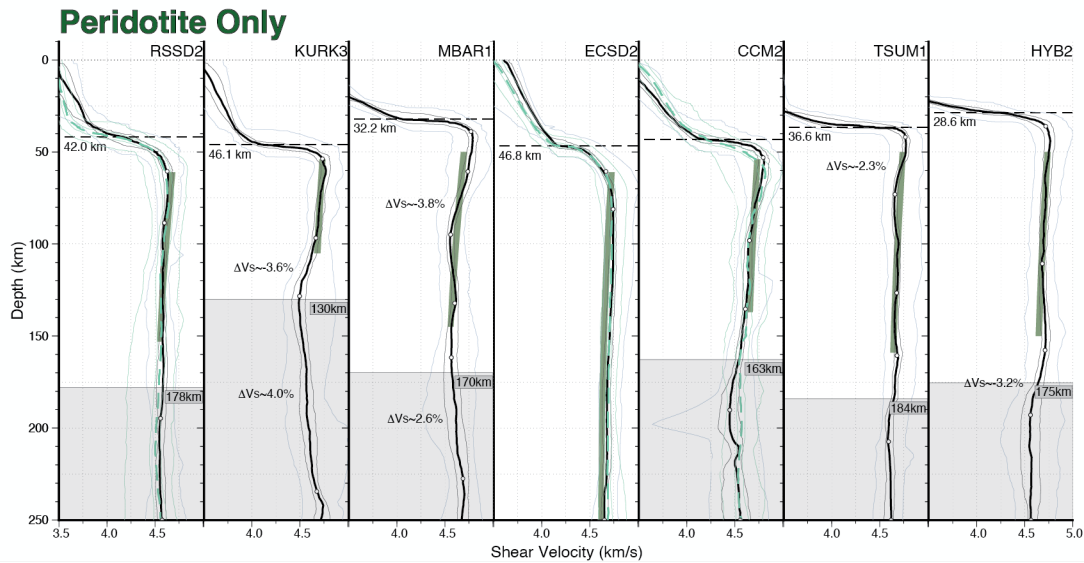


Figure 2-4. Shear velocity profiles fit with peridotite compositions

Shear velocity profiles (bold black line) plotted with depth resulting from joint inversions with the global Rayleigh phase velocity data set GDM52 (Ekstrom, 2011) and Sp RF stacks [Krueger et al., 2021]. One- and two-sigma bounds on shear velocity profiles are shown with thin black and gray lines, respectively. Thick colored lines indicate models for lithosphere composition that overlap with the one-sigma bounds for 95% of their range. Compositional models represent ranges in shear velocity for peridotite mantle compositions. Additionally, Vs profiles resulting from inversions using regional Rayleigh phase velocities are shown with a bold, dashed aquamarine line with surrounding thin lines indicating one- and two-sigma bounds. Changes in velocity gradient are indicated with white circles and velocity gradients over 2% are annotated. Moho depth estimates are indicated with a horizontal dashed line and annotated. LAB depth estimates are annotated in a gray box and indicated by a horizontal black line, while the underlying asthenosphere is shaded.

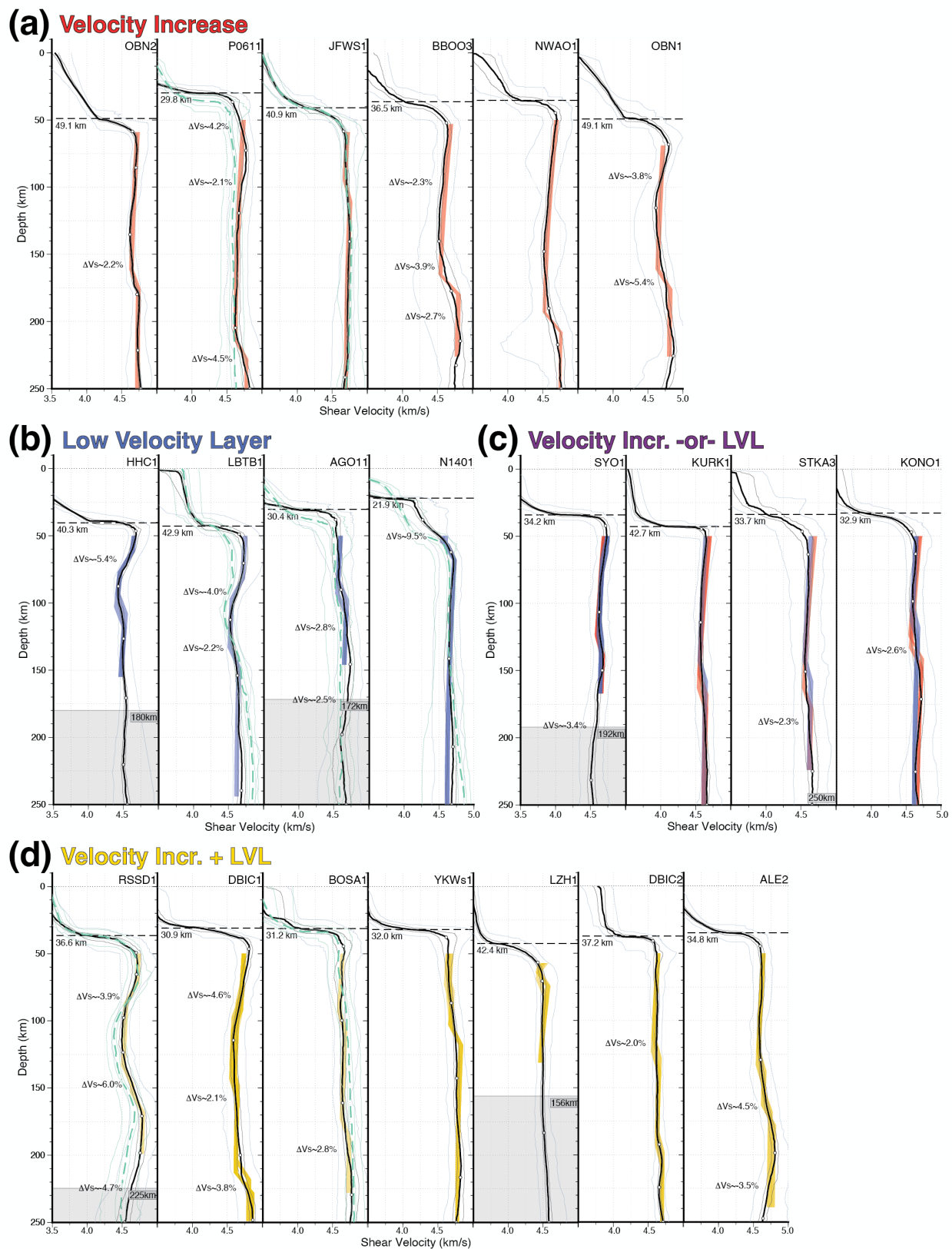


Figure 2-5. Vs profiles fit with layered compositions

Vs profiles (thick black line) plotted with corresponding compositional models (thick, colored lines) that represent ranges in shear velocity for **(a, red)** a Vs increase due to eclogite or diamond, and **(b, blue)** a low velocity layer in peridotite. **(c)** In some cases, a Vs profile can be fit with either a velocity increase or a low velocity layer. **(d, yellow)** Some Vs profiles require both a velocity increase and low velocity layer in the lithosphere. Shear velocity profiles plotted as in Figure 2-4.

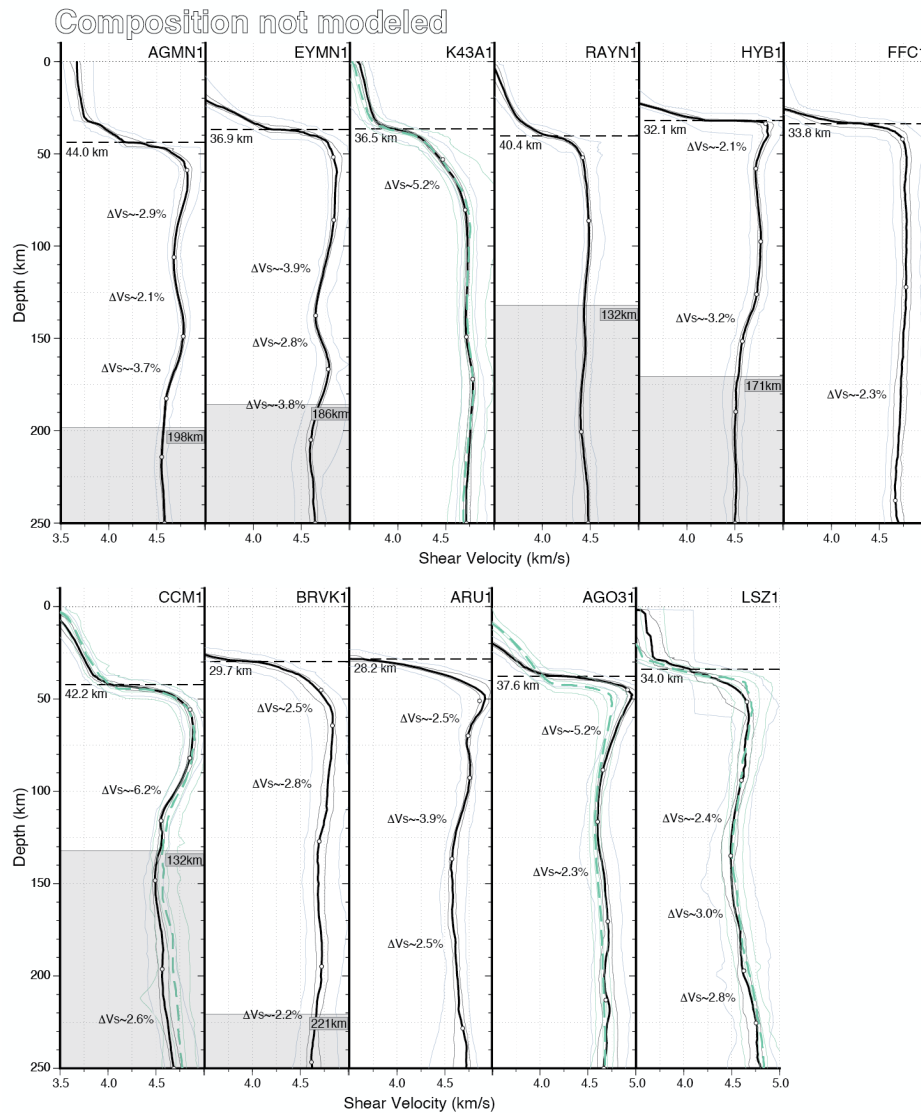


Figure 2-6. Vs profiles not fit with compositions

Vs profiles (thick black line) that are not fit with the range of compositions, layering structures, and geotherms we consider. Shear velocity profiles plotted as in Figure 2-4.

3.1.2 Vs profiles lacking cratonic lithosphere

Noncratonic Vs profiles TIA1 and LYN1 exhibit structure that is commonly observed in

noncratonic regions: (1) sharp NVGs above 100 km depth indicating a shallow (<100 km depth)

LAB; (2) deeper PVGs that define the lower boundary of a low velocity layer within the asthenosphere. TIA and LYN are located in the Eastern Block and Trans North China Orogen of the North China craton, respectively, where prior studies have found evidence for delamination of the cratonic lithosphere (Gao et al., 2002). Additionally, these profiles are located slightly outside of cratonic bounds provided by Woolley and Kjarsgaard (2008) (Fig. 2-1, outlines).

While the V_s profile for RAYN2 does have generally low upper mantle absolute velocities, it lacks a distinct large NVG to indicate a sharp LAB. Instead, the uppermost mantle exhibits an extended 2% NVG from the Moho to 170 km, followed by increasing shear velocities. It is possible this structure is due to the lack of fine-scale lateral resolution in the global model for Rayleigh phase velocities (Ekstrom, 2011), given that RAYN lies within a strong gradient in upper mantle velocities. Interestingly, we find that, despite using different criteria for craton classification, these profiles are also categorized as noncratonic in Krueger et al. (2021).

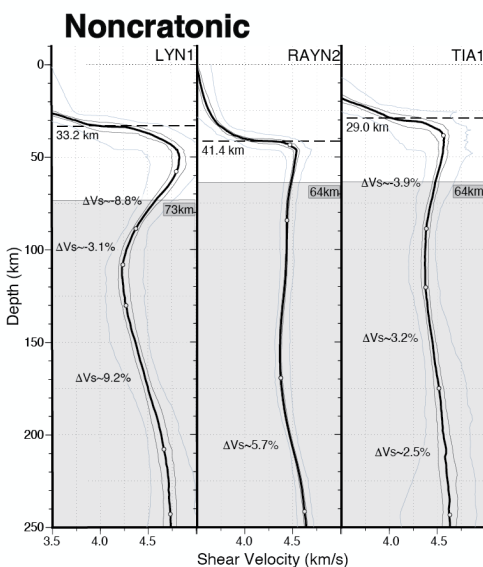


Figure 2-7. V_s profiles with noncratonic lithosphere

Examples of V_s profiles that are located within or adjacent to cratonic regions (Artemieva, 2006), but likely do not represent cratonic mantle due to their low upper mantle shear velocities. Shear velocity profiles plotted as in Figure 2-4.

3.2 Lateral variation in Vs profiles

Vs profiles inverted for data measured at the same station, but calculated for the 100 km conversion point location of different event backazimuths provide insight into local lateral variations in mantle structure (i.e. DBIC1 and DBIC2; Fig. 2-3). We see significant variation in mantle structure in all Vs profiles that represent data from a single station. We have two stacks representing nearby structure at the following stations: RSSD, DBIC, RAYN, KURK, CCM, OBN. In all cases, the velocity gradient structure (Fig. 2-4 to 2-6, white circles and annotations) varies significantly between pairs of profiles related to the same station. Additionally, the compositional structure required to model the Vs profile differs between profiles at a given station, except DBIC and OBN (Fig. 2-5, color-shaded profiles).

The presence of significant lateral variations in velocity structure raise the question of whether variations in true structure are adequately resolved in the GDM52 global phase velocities (Ekstrom, 2011). To evaluate this, we also present shear velocity profiles inverted with the same Sp RFs but using Rayleigh wave phase velocity data from regional surface wave studies (Supplementary Material section 2.1). Qualitatively, the velocity profile shapes appear remarkably similar between global and regional phase velocity cases (Fig. 2-4 to 2-6, aquamarine profiles). Quantitatively, the two-sigma uncertainties of the Vs profiles for global versus regional phase velocities overlap over almost all of the mantle at most stations, with LBTB1 and RSSD1 exhibiting the largest differences between global and regional cases. 11 of the 14 regional inversion profiles overlap the inversion profiles with one-sigma bounds with ≤ 50 km of depths not overlapping. However, more detailed comparisons between global and regional Vs profiles serve as a cautionary tale (Figures in supplement). In some profiles, the VGs and Moho/LAB depths are consistent (i.e. RSSD2, JFWS1, ECSD2, K43A1), but in many profiles

the calculated magnitude of velocity gradients vary significantly (i.e. LBTB1, BOSA1, AGO11, AGO31). We also see variations in LAB and Moho depths in some cases. Because of these differences in Vs profiles using global and regional surface wave data, we avoid interpreting specific VG magnitudes. Differences between regional and global inversions are discussed further in Supplementary Material section 2.1.

4 Discussion

4.1 Assessing bounds on the composition of the Cratonic Lithosphere

Our inversion results constrain the discontinuity and the absolute velocity structure of the cratonic upper mantle. The discontinuity structure from Sp RFs allows us to understand the depths of velocity gradients within the lithosphere but alone leaves ambiguity for interpretations of mineralogy. With the additional constraints on absolute velocity from Rayleigh phase velocities, we are able to assess the implications for mantle composition.

Specifically, we fit 28 of our 39 cratonic inversion profiles with simple compositional structures consisting of peridotite alone (Fig. 2-4), or peridotite plus a single velocity increase based on eclogite and/or diamond (Fig. 2-5a,c; Garber et al., 2018), and/or a single low velocity layer in the shallow lithosphere (Fig. 2-5b-d). Details of compositional fits are provided in Table S3. To represent peridotite and high velocity compositions, we use Vs values from the supplementary data of Garber et al. (2018) for peridotite as well as combinations of peridotite, eclogite (0 or 20%), and diamond (1-4%) for three geothermal gradients, 35, 40, and 45 mW/m². Garber et al. (2018) provides a range of Vs values for peridotite and peridotite-diamond compositions. To create a Vs range for peridotite-eclogite compositions, we use the range of peridotite for the corresponding geotherm around the peridotite-eclogite average Vs profile. To

investigate low velocity layers, we consider 1-6% velocity decreases to the peridotite Vs values from Garber et al. (2018).

To create layered structures, we calculate Vs profiles with low velocity layers and velocity increases at varying depths within a peridotite lithospheric mantle. We vary the depth at which a deeper layer containing graphite-diamond and/or eclogite begins, as well as the top and lower boundaries of a low velocity layer within the upper lithosphere. The depth of the top of the layer containing eclogite and/or graphite-diamond is a variable parameter, but in models containing graphite and diamond, the graphite-diamond transition determines the depth where diamond contributes to high velocities. Low velocity layers are limited to have lower boundaries above 150 km. If a profile has a low velocity layer and a velocity increase, the top of the velocity increase must be below the lower boundary of the low velocity layer. We calculate profiles for all combinations of composition, geotherm and velocity increases/decreases, the latter with depth increments of 10 km.

We calculate the root-mean-square misfit between each compositional Vs profile and the Vs profiles from the inversions over cratonic mantle lithosphere depths. Many of our profiles exhibit a relatively gradual crust-mantle transition. To avoid this transition when evaluating misfit, we identify the first NVG following the Moho or the first VG that lies >25 km below the Moho (the shallowest white circle on the profiles, Fig. 2-4 to 2-6). If this depth is shallower than the compositional data allows, we use 50 km as the top of the misfit calculation. The maximum depth of our misfit calculations is either 25 km above the LAB or 250 km, whichever is shallowest.

To be conservative about complexity in mantle composition, we first evaluate all possible peridotite compositions for each inversion profile. We perform an F-test to determine all average

peridotite composition profiles that are within the 99% confidence limits of the best-fitting model based on misfit. Peridotite models that pass this F-test are used to compared with inversion profiles. If the peridotite model Vs range and the 1-sigma bounds of the inversion profile overlap for 95% of the compared depths, we consider the profile to be explained by a peridotite-only model (Fig. 2-4). We then make the same comparison while allowing only a single velocity increase or a single low velocity layer (Fig. 2-5a-c). For the inversion profiles that still cannot be explained, we compare compositional profiles that combine a low velocity layer with a high velocity layer (Fig. 2-5d). After this process, about a quarter of the inversion profiles remain unexplained (Fig. 2-6). To see an example of the progression of these fits for station RSSD, see Fig. SB-3.

There are several limitations to this simple compositional comparison. (1) Inversion profiles with larger 1-sigma bounds may be fit with simpler mantle structure than is indicated by the shape of the Vs median profile (i.e. MBAR1, Fig. 2-4). (2) Non-uniqueness is created by trade-offs between the effects of varying geotherms, composition, and imposed velocity decreases (Fig. 2-5c). (3) We explore a limited range of compositions and layer configurations, with the result that not all inversion profiles are well-modeled (Fig. 2-6). (4) Despite smoothing the layered compositional profiles before calculating misfit with a 15 km moving average, compositional profiles have relatively sharp boundaries (i.e. STKA3, Fig. 2-5c). It is apparent from synthetic inversions that our inversion methodology can smooth layered structure (Fig. 2-2). Difference in the shape of sharp layering in composition and smoothed layering in inversion profiles can lead to high misfits near a discontinuity (Fig. 2-2b). This may explain the profiles that were not fit by our modeling scheme. (5) The absolute velocities in our Vs profiles are largely controlled by the Rayleigh wave velocities, which are corrected for attenuation in the

PREM model (Dziewonski & Anderson, 1981) referenced to a 1 Hz frequency during the inversion, whereas the V_s models from Garber et al. (2018) are purely elastic. However, this approximation likely does not result in a significant mismatch. If true attenuation matched the PREM model, the attenuation correction would result in 1 Hz V_s values from the inversions that underestimate the elastic case. However, since attenuation in cratons is likely lower than in the PREM model, the V_s values may have been slightly over-corrected for attenuation, resulting in an increase in V_s in the direction of the elastic values. An alternative approach would be to correct the elastic V_s values from Garber et al. (2018) to nominal 1 Hz V_s values using experimentally determined parameterizations such as Jackson and Faul (2010). However, given the large uncertainties in such corrections due to imperfectly known parameters such as grain size, we chose the simpler approach of assuming that the 1 Hz V_s values from the inversions approximate elastic velocities.

Overall, fitting the observed V_s profiles with these simple layered models offers a conservative approach to interpreting mantle structure. Because of the limitations built into this strategy, we do not focus on the compositional models that fit individual profiles. Rather, we focus on ensembles of compositions as a tool for understanding the heterogeneity of the cratonic lithosphere.

4.1.1 Simple compositional models

Peridotite compositions provide acceptable fits to seven of the 39 V_s models from the inversions over 95% of the depths range of the mantle lithosphere (Fig. 2-4). Most (5/7) of the profiles are fit with by a range of peridotite V_s values at 35 mW/m², while the remaining profiles are fit with a 40 mW/m² geotherm. To conservatively allow for a broader range of possible compositions, we use the 1-sigma V_s bounds of each inversion profile to assess where a compositional model fits at

a given depth, range instead of comparing the compositionally predicted Vs to the median observed Vs profile. This approach may result in underfitting of structure, particularly for profiles with larger error bounds (i.e. MBAR1). With lower uncertainty ranges, the observed Vs profiles at MBAR1 and TSUM1 would be best fit with a single low velocity layer; HYB2 may exhibit a lower lithosphere velocity increase; and ECSD2 may require the addition of a high velocity composition (such as eclogite) throughout all depths. However, given our conservative approach, they are adequately fit by peridotite alone.

10 of the remaining 21 Vs models from the inversions can be fit over 95% of mantle lithosphere depths by compositional models that include peridotite at shallower depths with a transition to a composition that includes diamond or eclogite at greater depths (Fig. 2-5a,c). All of the 10 Vs profiles are fit by models including diamond, which causes a velocity increase at 100 km or deeper, due to the depth of the graphite-diamond transition, regardless of the top of the imposed layer. Some of these inversion profiles can also be satisfied with low velocity layers, as explained in the next section (Fig. 2-4c). Most of these stacks are well-fit, although OBN1 may be better fit with an additional low velocity layer centered at 125 km, and P0611 may be better fit with an additional high velocity layer at ~75 km.

Eight of the remaining 11 Vs models can be fit with a low velocity layer with a velocity decrease of 1-3% in the upper 150 km of the lithosphere (Fig. 2-5b,c). As previously mentioned, four of these inversion profiles can also be satisfied with a velocity increase (Fig. 2-5c). N1401 can also be explained with a 4% low velocity layer, but this velocity reduction is potentially related to a strange Moho signal (Section S3.1.6). Qualitatively, all stacks appear to fit inversion profiles well, although AGO11 and N1401 could be better fit by also including velocity increases around 125 and 250 km, respectively.

4.1.2 Complex lithospheric structures

Seven inversion profiles are fit for 95% of mantle lithosphere depths by a combination of a single low velocity zone overlying a velocity increase and could not be explained by simpler lithospheric structure (Fig. 2-5d). All low velocity zones can be explained by shear velocity decreases of 1-3%, excluding LZH1 (see below). All velocity increases involve 1-3% diamond (excluding LZH1), with some also requiring the presence of eclogite.

All of these fits appear excellent. However, LZH1 is an anomalous case. LZH1 has relatively low mantle shear velocities for cratonic mantle and has no significant velocity gradient. Among the models tested, it is best fit with a low velocity layer of 6% combined with a layer of 4% graphite-diamond starting at a depth of 66 km, plus a higher temperature geotherm (45 mW/m²). Due to the relatively shallow LAB at LZH1 (156 km) and warmer geotherm, this model contains graphite throughout the depth range considered for misfit, which adds an additional slight velocity decrease. These features combine to simply create overall velocity decreases in the profile, instead of a clear layered structure. While this structure is not impossible, it is non-unique and atypical of the other models. After this point, we do not consider LZH1 compositions for our discussion or analysis.

The compositional models considered here fail to match 11 of the Vs models from the inversions. In some cases, the models we test do not reach the necessary shear velocities in the uppermost mantle, as in the cases of the CCM1, BRVK1, ARU1, FFC 1, EYMN1, AGO31 Vs models, which all have exceptionally high shallow shear velocities. In some cases the layered structures we test are too simple, which is likely for the complicated layering obvious at ARU1. It is also possible that the sharp gradients we impose at layer boundaries introduce large amounts

of misfit at transitions between layers, which is likely for profiles with distributed velocity gradients, such as LSZ1.

4.2 Layering in the Global Cratonic Lithosphere

The Vs profiles we solve for indicate that the cratonic lithosphere is heterogeneous at all scales.

These findings coincide with prior work that indicates significant heterogeneity within the cratonic mantle lithosphere (e.g. Altoe et al., 2020; Hopper & Fischer, 2018; Liu & Shearer, 2021). Notably, only 11 of the 28 inversion profiles fit by the compositional models require a low velocity layer in the upper 150 km (Fig. 2-5b,d). This result indicates that the existence of a lithospheric velocity drop in the shallow lithosphere, commonly termed the mid-lithospheric discontinuity (MLD), is not ubiquitous in cratons globally. This evidence confirms a similar observation in Krueger et al. (2021) and coincides with a growing body of work that finds the MLD to be variable in occurrence and character (e.g. Altoe et al., 2020; Liu & Shearer, 2021). This observation has important implications for studies that evaluate the origin of lithospheric layering, for example studies that use MLD ubiquity as a test for plausible mechanisms (Fu et al., 2022; Yang et al., 2023).

For stations with acceptable compositional models with low velocity layers, velocity drops of 1-3% are observed, with 2-3% being the most common (Fig. 2-8ab). The range of Vs reductions we observe is significantly smaller than some prior estimates of up to 12% velocity reductions (Hopper & Fischer, 2018; Wolbern et al., 2012). Low velocity layer thicknesses are distributed from 10 to 100 km, but 60 km is the most common value (Fig 2-5c). The center depths of these low velocity layers vary from 55 to 125 km (Fig. 2-8d). The distributions of acceptable model parameters show no obvious trends with thermotectonic age.

Amongst the acceptable compositional models that require velocity increases, all involve diamond while only half involve eclogite. Velocity increases due to 1-2 vol% diamond introduced at 150-175 km are most common in the compositional models. Again, we do not observe any pattern in model parameters with thermotectonic age despite using similar Sp RFs as Krueger et al 2021 (Fig. 2-8, 2-9). This finding is consistent with Liu et al. (2021).

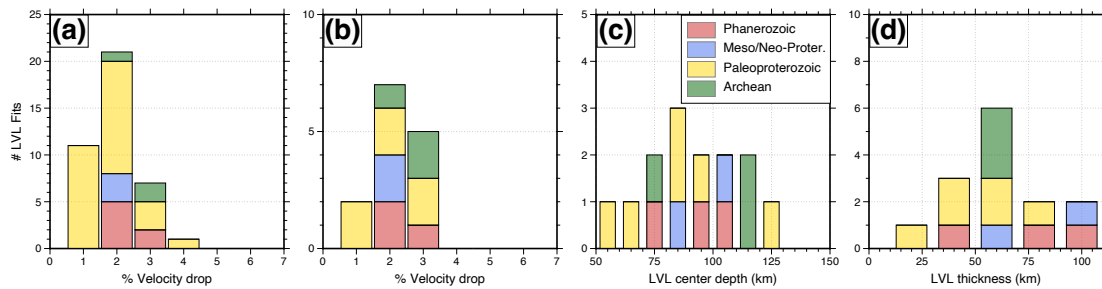


Figure 2-8. Low velocity layer characteristics

Summaries of acceptable compositional models that have low velocity layers (LVLs) in the upper 150 km of the lithosphere (Fig. 2-4b to 2-4d). **(a)** The distribution of velocity drops found from all models that fit Vs inversions, including multiple models that fit a single inversion. **(b)** The median velocity drop of the compositional models that fit each Vs model. **(c)** The average value of the central depth of the low velocity layers among acceptable models. **(d)** The thickness of low velocity layers for a single profile. In all panels, the data are binned and bars are plotted at the center of each bin. Model parameters for profile LZH1 are not included.

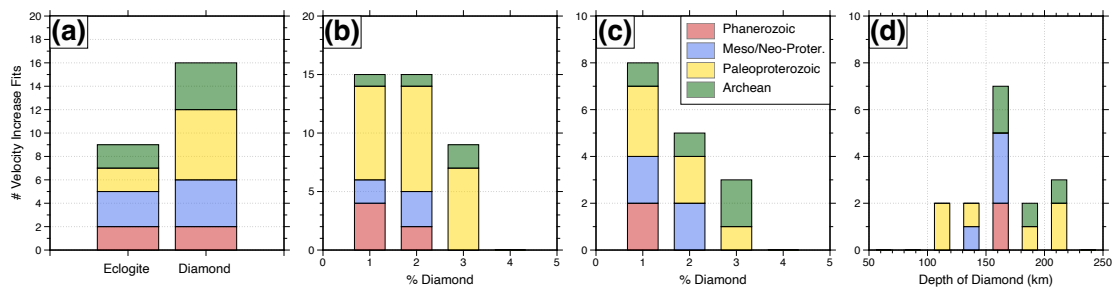


Figure 2-9. Velocity increase characteristics

Summaries of acceptable compositional models that have velocity increases in the lithosphere (Fig. 2-4a,c,d). **(a)** The number of velocity increases that involve eclogite and diamond. **(b)** The range of % vol. diamond in all acceptable models, including multiple models that fit a single Vs profile. **(c)** The median % vol. diamond of the compositional models that fit each Vs profile. **(d)** Average values for the depth of the velocity increase in the mantle among the models that fit each Vs profile. In all panels, the data are binned and bars are plotted at the center of each bin. Model parameters for profile LZH1 are not included.

4.2.1 Low velocities in the cratonic lithosphere

All profiles we model can be explained with layers with velocity reductions 3% or less in the upper 150 km of the mantle lithosphere. Since velocity decreases of 1-6% were tested, the 3% maximum velocity decrease from peridotite appears to be a robust upper bound.

The commonly proposed mechanisms for observed shear velocity reductions observed in the cratonic upper mantle are variation in layering in melt depletion, melt presence, elastically accommodated grain boundary sliding, anisotropy, and metasomatic minerals (e.g. Hansen et al., 2015; Karato et al., 2015; Rader et al., 2015; Saha et al., 2021; Selway et al., 2015). Here, we consider these mechanisms and outline evidence that metasomatic minerals within the cratonic upper mantle which have low seismic velocities and are anisotropic are the best explanation for the full range of observed velocity reductions.

Neither melt depletion nor elastically accommodated grain boundary sliding can explain all of our observations of low velocities in the cratonic lithosphere. Shear velocity increases due to melt depletion are estimated to be 1-1.7% (Afonso & Schutt, 2012). Melt depletion layering (e.g. a less depleted layer in between two more depleted layers) may explain low velocity layers with velocity drops at the lower end of the range we observe, such as the velocity drop at TSUM1 which is actually encompassed by the range of peridotite compositions at a single geotherm. However, layering in melt depletion cannot explain the V_s profiles that are fit by low velocity layers with up to 3% velocity drops. Attenuation due to elastically accommodated grain boundary sliding can explain velocity reductions of a few percent in the uppermost mantle (Karato, 2012; Karato & Park, 2018; Krueger et al., 2021). However, this mechanism predicts a velocity reduction in all locations in the depth range of 50-150 km, and we do not observe low velocity layers or velocity drops in the upper 150 km of approximately half of our inversion

profiles (Fig. 2-8), indicating that any mechanism requiring a ubiquitous mid-lithosphere feature is unlikely based on our results. Furthermore, although we cannot rule out the presence of small amounts of melt at the locations we observe, all of the velocities and velocity reductions we observe can be explained without melt presence (Saha et al., 2021).

If actual partial melt is present within the shallow mantle lithosphere of cratons, it would have potential to create velocity reductions of as much as 3%. However, while some studies have proposed partial melt layers in the shallow cratonic mantle (e.g. Thybo & Perchuc, 1997), cratonic geotherms are typically thought to be too cold to permit the presence of partial melt (e.g. Karato et al., 2015). Although elevated levels of CO₂ could potentially allow partial melt to be stable at the upper end of the global range of cratonic geotherms (Dasgupta, 2018), such geotherms are warmer than the 35-45 mW/m² geotherms used to model the Vs profiles, and would require more anomalous compositions (e.g. higher percentages of diamond) to explain the observed high velocities in the deep lithospheric mantle.

Volatile-rich minerals in the solid-state provide a simpler explanation for the low velocity layers in the shallow cratonic mantle. The most commonly proposed low velocity minerals in the cratonic lithosphere are the hydrous minerals amphibole and phlogopite (e.g. Altoe et al., 2020; Eeken et al., 2018; Rader et al., 2015; Saha et al., 2021; Selway et al., 2015). Phlogopite is often evoked because small amounts can cause significant velocity drops (~1% phlogopite decreases Vs by ~1%) and it is stable throughout all cratonic lithosphere depths (Peng & Mookherjee, 2020). Several studies note that amphibole stability in the cratonic mantle only reaches to approximately 90 km depth, which does not encompass the depth range of low velocity layers we observe (Peng & Mookherjee, 2020; Selway et al., 2015). Alternatively, (Saha et al., 2021) finds that the stability of amphibole is strongly dependent on mantle composition (bulk K₂O/H₂O) and

can exist throughout the lithospheric mantle under certain conditions, allowing amphibole presence to be a viable explanation for the low velocities we observe.

Based on the bulk peridotite compositions of Archean xenoliths and experimental data on phase equilibria in H₂O and CO₂ bearing peridotite systems, Saha et al. (2021) place upper limits on the amounts of amphibole and phlogopite that could be produced by metasomatism of the original Archean mantle: $\leq 10\%$ pargasitic amphibole and $\leq 2\%$ phlogopite. Saha et al. (2021) further conclude that these levels of amphibole and phlogopite can produce velocity reductions of up to 3% for geotherms comparable to those we employed to model the V_s profiles. The acceptable low velocity layers from our modeling all have $\leq 3\%$ velocity reductions, meaning that these velocity drops can be explained by the presence of hydrous minerals within the cratonic mantle, without resorting to compositions that deviate from metasomatism of typical bulk Archean mantle lithosphere.

The significant lateral variability we see in existence, magnitude, and depths of low velocity layers could be partially explained by small-scale variations or anisotropy of hydrous minerals. Additionally, hydrous minerals sourced from percolating melts and/or interaction with kimberlites would exist in veins or dikes and vary at local scales and global scales (e.g. Jackson & Gibson, 2023). This is evidenced by the lack of amphibole and phlogopite in xenoliths at all cratonic localities (Selway et al., 2015). Additionally, phlogopite exhibits significant seismic anisotropy at pressures corresponding to the cratonic lithosphere (Chheda et al., 2014). Amphibole is also seismically anisotropic (Peng & Mookherjee, 2020). If phlogopite or amphibole aligned at a single location, anisotropy could cause significant variation in V_s, creating apparent changes in the magnitude of velocity decreases within low velocity layers (Chheda et al., 2014; Peng & Mookherjee, 2020).

4.2.2 High velocities in the cratonic lithosphere

All profiles we successfully model can be explained by compositions with 0-3% diamond and/or 0-20% eclogite. The range of diamond content is consistent with the estimates from Garber et al. (2018) who found that diamond of 1-3% provided the best match to observed Vs in the cratonic mantle lithosphere, with a most likely bulk diamond of content of ~2%. In our compositional models, we only find eclogite in approximately half of our models with velocity increases, which is relatively consistent with the observation that diamond inclusions indicate that only about 30% of diamonds are eclogitic (Garber et al., 2018; Stachel & Harris, 2008). Furthermore, Garber et al. (2018) consider graphite/diamond to exist in equal amounts at all depths, but most proposed mechanisms for diamond emplacement involve melt infiltration from below cratonic keels (e.g. Pinter et al., 2022; Walsh et al., 2023; Weiss et al., 2021) and redox reactions that decrease with pressure within the cratonic lithosphere (Pinter et al., 2022). Because of this, we assume that carbon is not equally distributed throughout the cratonic mantle and consider the top of the high velocity layers we model as the boundary where diamonds begin to exist in the lower lithosphere.

While we explicitly explored cratonic mantle velocity increases due to diamond and eclogite, other potential mechanisms for velocity increases within the cratonic mantle also exist. Anisotropy within or at the base of the cratonic lithosphere (Gung et al., 2003) or the phase transition from spinel- to garnet-peridotite (Lebedev et al., 2009) could also result in shear velocity increases in the upper mantle. For example, we also observe some shallower velocity increases that are not captured by the compositions from Garber et al. (2018). These are at P0611 (Fig. 2-5a) as well as K43A1 and BRVK1 (Fig. 2-6). It is possible these gradients are related to

the transition from spinel- to garnet-peridotite (Yang et al., 2023), although layers of eclogite in the uppermost mantle (Altoe et al., 2020) are another alternative.

While a complete investigation of anisotropy as the source for a velocity increase in the cratonic mantle is beyond the scope of this work, we do not think anisotropy is the primary source of these high shear velocities. For stations OBN and DBIC, the presence and depth of velocity increases from separate stacks appear relatively consistent despite overlying low velocity layer structure showing variation between stacks (Fig. 2-5). Other stations with multiple stacks are difficult to compare (KURK and RSSD) due to significant lateral variability in the overall lithosphere.

Finally, the fact that we observe significant velocity increases in over half of the inversion profiles we can model with composition (Fig. 2-5) indicates that such velocity increases are a fairly common feature of the cratonic mantle. This result is consistent with the widespread observation of a velocity increase in the 150-300 km depth range below cratons in global models of shear velocity and attenuation (Dalton et al., 2017; Dalton & Faul, 2010; Garber et al., 2018), and may explain the velocity increase in this depth range identified as the Lehmann discontinuity (Lehmann, 1959). In some of the V_s profiles from the inversions in this study, it is ambiguous if this velocity increase is the top of a high velocity layer or bottom of a low velocity layer (Fig. 2-5c), but in most cases we are confident the velocity increase is the top of a high velocity layer.

4.3 A Common Origin for Hydrous Minerals and Diamond in the Lithosphere

In the previous section, we concluded that phlogopite and amphibole are the most likely cause of velocity reductions in the mid-lithospheric cratonic mantle. Multiple origin(s) for these hydrous

minerals have been proposed. Some studies believe that hydrous minerals were introduced into the cratonic lithosphere during craton formation (Liu et al., 2021), amalgamation (Lee et al., 2011) or early evolution (Yang et al., 2023). For example, Yang et al. (2023) propose that channel flow was triggered within the lithosphere of young cratons resulting in shear zones with strong anisotropy, grain size reductions, and accumulated now-crystallized metasomatic fluids. Others conclude that lithospheric hydration occurred after craton formation from underlying asthenospheric upwellings or subduction dehydration, although the scale of this hydration is unclear (e.g. Fu et al., 2022; Jackson & Gibson, 2023). For example, Fu et al. (2022) model long-term upward migration of fluid which collects in a mid-lithospheric layer. However, in both Yang et al. (2023) and Fu et al. (2022), the modelled processes result in a fairly globally ubiquitous seismic velocity feature in the cratonic upper mantle, which is inconsistent with the lateral variability in mid-lithospheric low velocities we observe in the V_s profiles in this study. To reconcile these models with our results, the formation processes would either have to be variable at shorter length-scales, or a subsequent process that disrupts the mid-lithospheric low velocity layer would also need to occur.

Alternatively, intermittent and heterogenous infiltration of carbon-rich fluids into cratonic lithosphere from kimberlites could lead to a more variable formation of hydrous minerals (Jackson & Gibson, 2023) that is consistent with the wide range of velocity structures we observe in cratonic mid-lithospheres. The varied compositions of kimberlites indicate that different styles of metasomatism may occur in the cratonic mantle as volatiles interact with surrounding peridotite. As these fluids evolve in their ascent, phlogopite and amphibole can form (Jackson & Gibson, 2023; Pinter et al., 2022). The addition of graphite in the upper lithosphere

from carbonatite fluids could also explain high-conductivity layers observed within the cratonic upper mantle (Chen et al., 2009; Dasgupta & Hirschmann, 2010).

Reactions between peridotite and carbonate fluids may also play a role in the shear velocity increases we observe in the lower cratonic lithosphere, in the sense that high velocity minerals are also produced by the interaction between carbonate melts and the lithospheric mantle. Most notably, diamond is deposited when oxidized carbonate melts first contact reduced lithosphere at high pressures (Dasgupta & Hirschmann, 2010). In this reaction eclogitic minerals can also sometimes be stabilized (Pinter et al., 2022). As the reaction front progresses into the lithospheric material, diamond continues to deposit. When the leading edge of the fluid migration solidifies, a hydrous assemblage rich in phlogopite, clinopyroxene, and amphibole forms. If the fluid is of a carbonate silicate composition, the reaction front deposits fewer hydrous minerals (Pinter et al., 2022). The reaction propagates when re-melting occurs due to continued upwellings or temperature changes. This process can cause upward propagating dikes and gradually oxidizes the roots of cratons (Pinter et al., 2022).

Kimberlite interactions are expected to be variable locally and globally (e.g. Eeken et al., 2018; Tomlinson & Kamber, 2021). Reaction fronts of carbonate fluid are expected to propagate upwards as dikes instead of altering the entirety of the surrounding lithosphere (Pinter et al., 2022). Furthermore, in xenoliths, the amount of eclogite and diamond is incredibly variable. Only a fraction of a percent of xenoliths constitutes diamonds, but some xenoliths exhibit “seams” of diamond (Garber et al., 2018 and references therein). This could be due to lateral variation in the reduced state of the ancient lower lithosphere causing varied potential for diamond growth (Tomlinson & Kamber, 2021; Walsh et al., 2023).

Most of the inversion profiles we model are consistent with the emplacement of high velocity minerals in the lower lithosphere by carbonate melt. In some cases, this melt may percolate upward to form hydrous low velocity minerals in the upper mantle. Although not all compositions of carbonate melt form the low-velocity mineral phlogopite and some observations may be obscured due to anisotropy. The intermittent observation of low velocity layers overlying velocity increases corresponds with this process.

Conversely, a low velocity layer without an underlying velocity increase is inconsistent with the process of carbonate melts percolating upward from the base of the lithosphere. However, two of the four profiles (LBTB1, N1401) fit only low velocity layers appear to be underfit and do exhibit velocity increases that are clearer in inversion profiles from regional SW data (Fig. 2-5b, aquamarine). Furthermore, the remaining two low velocity layer profiles (AGO11, HHC1) and six of the seven profiles fit with peridotite compositions, have thinned cratonic lithospheres ($LABs < 190$ km). The absent lower lithosphere could have had diamond and/or eclogite prior to thinning. It is also possible that the lower lithosphere at these locations was not adequately depleted for diamond-producing carbonate reactions. If the lithosphere was not adequately depleted, introduction of metasomatic fluid may have weakened the cratonic keel instead of crystallizing diamond (Walsh et al., 2023), promoting thinning.

Based on these observations, we speculate that velocity increases in the lower cratonic lithosphere due to diamond and/or eclogite presence may be a common feature in cratons with intact, deep lithospheres ($LAB > 250$ km).

5 Conclusions

To provide new insight on the question of layering in the cratonic mantle lithosphere, we combine Sp RFs and Rayleigh phase velocities in a Bayesian MCMC inversion. The

combination of these two data types allows for the resolution of discontinuities and absolute velocities within the uppermost mantle. Resolving both seismic characteristics in one profile provides a powerful tool for interpreting global patterns (or lack thereof) in mid-lithospheric layering and mantle compositions.

We resolve shear wave velocity structure for the upper 250 km of the mantle at 42 locations globally. Of these, 39 locations are underlain by cratonic lithosphere extending below 150 km depth. Within these profiles we observe a wide variety of complexity in lithospheric structure, such as layering, that varies on local and global scales. Simple modeling of a peridotite mantle with a single velocity increase and/or a single upper-lithosphere low velocity layer explains the structure of 28 of our cratonic profiles. Seven of our inversion profiles can be explained by peridotite-only compositions. Most of these peridotite-only profiles exhibit thinned lithospheres. Ten profiles can be matched by peridotite plus a single velocity increase within the lithosphere, although four of these could also be explained by peridotite plus a single low velocity layer. Four additional profiles indicate a low velocity layer in the upper lithosphere. Of these, two have thinned lithospheres and the other two appear to be underfit and may have velocity increases as well. Seven profiles are fit by a combination of a low velocity layer and a velocity increase.

Our models find that low velocity layers with velocity reductions of 1-3% in the upper lithosphere fit half of our profiles and are likely due to the presence of hydrous minerals such as phlogopite and amphibole. We also find that velocity increases due to 1-3% diamond fit 60% of our profiles. Of the profiles that exhibit low velocity layers, but not velocity increases, we observe either thinned lithosphere or the likely underfitting of lower lithosphere velocity

increases. This indicates that in these cases, either a velocity increase in the lower lithosphere is possible or the lower lithosphere has been removed.

The structures we observe support the common origin of carbonate fluids for both low velocity layers and velocity increases in the lithosphere. The infiltration of carbonate fluids from underlying asthenospheric upwellings into a depleted cratonic keel could lead to deposition of diamond, while further upward percolation of these fluids could result in the crystallization of low-velocity hydrous minerals. The lack of low velocity layers in half of the profiles could be explain by (1) limited infiltration of underlying fluids, or (2) silica-rich carbonate fluid that deposits diamond without introducing phlogopite (Pinter et al., 2022). Variability in the occurrence and composition of asthenospheric upwellings also explains the heterogeneity we observe in cratonic lithospheric structure.

6 Acknowledgements and Data

We thank Zachary Eilon, and Colleen A. Dalton for their useful advice regarding methodology and interpretations. We also thank Dr. Weisen Shen, Dr. Jordyn Babikoff, Dr. Andrew Nyblade, Dr. Xuewei Bao for sharing their regional phase velocities with us. This work was funded by the U.S. National Science Foundation's Graduate Research Fellowship Program and the National Science Foundation Award EAR-2044136. Generic Mapping Tools is used to plot figures in the manuscript and supplementary information (Wessel & Smith, 1991).

Seismic waveform data were acquired from the Integrated Research Institutions in Seismology Data Management Center (IRIS DMC). These data are from the following networks: the IRIS/USGS Global Seismic Network (II: <https://doi.org/10.7914/SN/II>, IU: <https://doi.org/10.7914/SN/IU>); the Global Telemetered Seismograph Network (GT: <https://doi.org/10.7914/SN/GT>); GEOSCOPE (G: <https://doi.org/10.18715/GEOSCOPE.G>); the

United States National Seismic Network (US: <https://doi.org/10.7914/SN/US>); the Canadian National Seismograph Network (CN: <https://doi.org/10.7914/SN/CN>); GEOFON (GE: <https://doi.org/10.14470/TR560404>); the Australian National Seismograph Network (AU); the Central and Eastern US Network (N4: <https://doi.org/10.7914/SN/N4>).

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Chapter 3 Investigating the Complex Tectonic History of Antarctica with Phase Velocity & Attenuation

Hannah E. Krueger^a, Colleen A. Dalton^a, Josh Russell^b, Karen M. Fischer^a

^a Department of Earth, Environmental, & Planetary Sciences, Brown University

^b Syracuse University

Key Points:

- Rayleigh wave observations of phase velocity and attenuation reflect the general dichotomy of cratonic East Antarctica and tectonically active West Antarctica
- Rayleigh wave phase velocities reveal complex mantle heterogeneity in cratonic and noncratonic regions of Antarctica
- Preliminary attenuation inversions reveal lower attenuation beneath the Gamburtsev Mountains in East Antarctica, and higher attenuation beneath West Antarctica

Abstract

Antarctic geologic structure is poorly understood due to its limited accessibility. In the past 20 years, various seismic deployments have created an excellent dataset to use for regional-scale seismic studies. We utilize broadband and low frequency vertical-component seismic data from four regions across Antarctica for Eikonal tomography. We calculate Rayleigh wave phase velocities for the Gamburtsev Subglacial Mountains in East Antarctica, the northern Transantarctic Mountains and adjacent Ross Embayment, as well as various tectonic provinces of West Antarctica, and the northernmost tip of the Antarctic Peninsula. Although the regional-scale phase velocities we observe are consistent with recent continent-scale studies (e.g. Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018), the detail we resolve is significantly improved. We also present preliminary surface wave attenuation results calculated using the innovative ASWMS-Q software from Russell and Dalton (2022). Although attenuation results are still being refined, certain key features are resolved, such as consistently low attenuation values for the cratonic Gamburtsev Subglacial Mountains and the northern Transantarctic Mountains as well as high attenuation values beneath a large swath of West Antarctica that includes part of the West Antarctic Rift System. Through their implications for mantle viscosity and heat flow, these attenuation results have the potential to aid models of ice sheet melting and glacial isostatic rebound in future work.

1 Introduction

Our understanding of the geologic and geophysical character of Antarctica is relatively rudimentary due to isolation, difficult field conditions and ice cover. In this setting, the use of geophysics is particularly important to probe below the surface of thick ice sheets. Furthermore, the modern challenge of sea level rise contextualizes the importance of the Antarctic mantle. Mantle temperature and viscosity are directly related to the rate of ice melting and the isostatic adjustment of land as ice melts, respectively (e.g. van der Wal et al., 2015). Recent geophysical studies have achieved resolution with methods that previously were impossible in Antarctica due to array density (e.g. Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018). In this study, we contribute to the understanding of Antarctic upper mantle characteristics, through calculation and interpretation of Rayleigh phase velocities and attenuation at a regional scale in Antarctica (Fig. 3-1, red outlines).

Antarctica has two primary tectonic regions, East and West Antarctica, which are split by the Transantarctic Mountains (Fig. 3-1). East Antarctica has remained relatively stable since the Proterozoic, while West Antarctica has experienced extensive Phanerozoic tectonic activity consisting of margin-wide subduction, supercontinent breakup, and multiple stages of extension (Jordan et al., 2020 & references therein). In order to understand the broad character of Antarctic upper mantle structure, we study regions from cratonic East Antarctica, rifted West Antarctica, the subduction zone at the northernmost Antarctic Peninsula, and the boundary between East and West Antarctica. The regions we investigate are: (A) The Gamburtsev Subglacial Mountains (GSM) in East Antarctica; (B) the northern Transantarctic Mountains (TAM), the Ross Embayment, and Terror Rift; (C) the West Antarctic Rift System (WARS), Marie Byrd Land (MBL) and the Ellsworth-Whitmore Mountains (EWM) in West Antarctica; and (D) the South

Shetland Islands. The geologic makeup of these regions is summarized in Figure 3-1A-D. In this investigation, we seek to holistically understand observations of Rayleigh wave phase velocity, and initial attenuation results in the context of tectonic processes and previous geophysical studies.

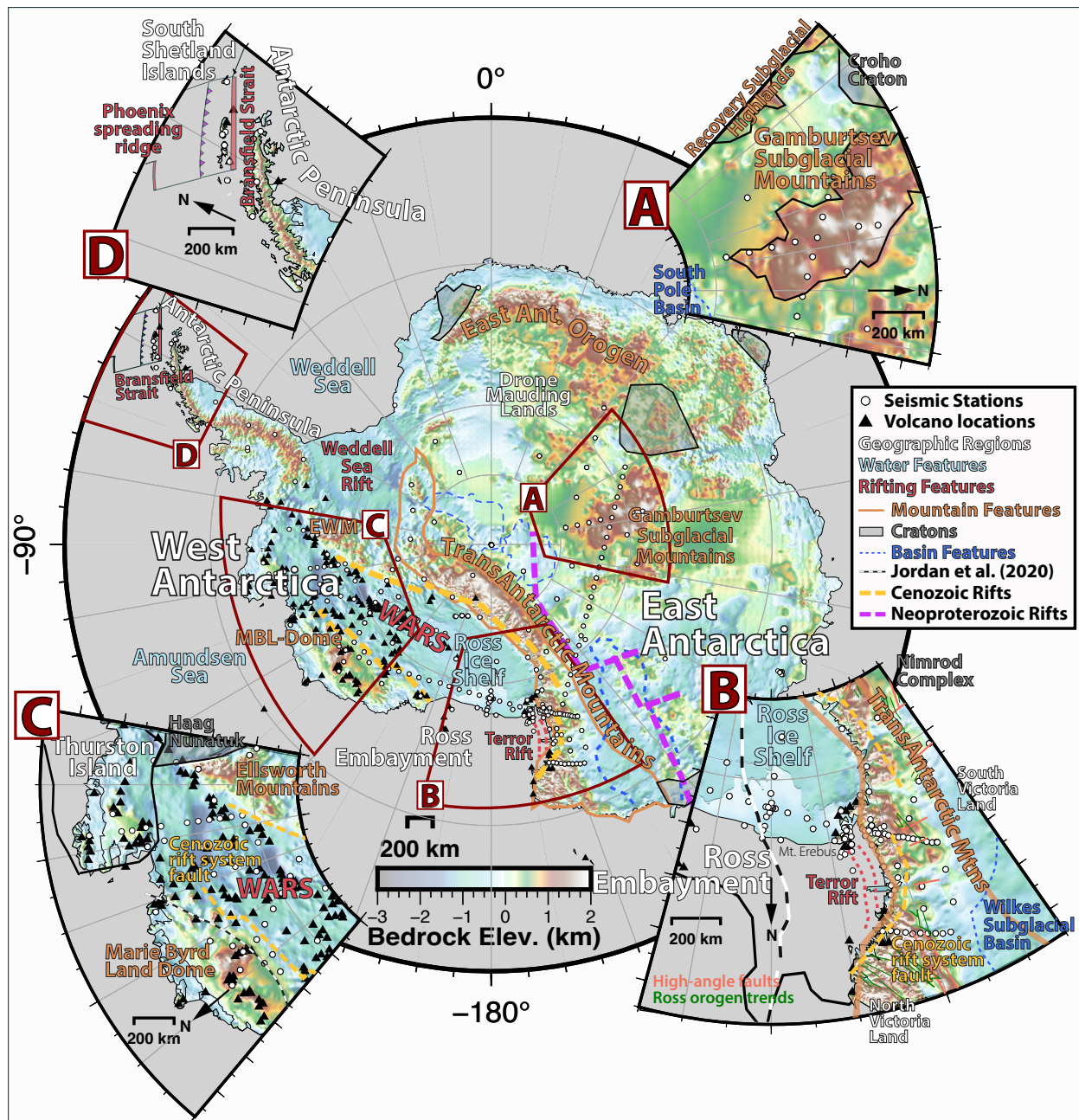


Figure 3-1. Map of Antarctica

Map of Antarctica with individual study regions (outlined in red) annotated with major geographic and geologic features and colored by elevation (Amante & Eakins, 2009). (A-D) magnified maps of the regions discussed in this

study. Text color indicates type of features: geographic regions are labeled in white text; water features labeled in light blue text; mountainous features labeled in brown; cratons gray; basin features dark blue; regions of rifting red. White and black dashed lines indicate the boundaries between the MBL, WARS, and Ross Ice Shelf sub-provinces from (Jordan et al., 2020). Shaded gray regions are considered cratonic and have geologic evidence of Archean or Paleoproterozoic material. Dashed lines indicate Cenozoic (orange) and Neoproterozoic (purple) rifts. White circles are seismic station locations and black inverted triangles are volcano locations. Geologic maps are modified from those provided in the cited articles (Goodge, 2007, 2020; Jordan et al., 2013; Jordan et al., 2020; Kleinschmidt, 2021; Paxman et al., 2019; Rilling et al., 2009; Robertson Maurice et al., 2003).

2 Methods

2.1 Data

The focus of our analysis is determination of Rayleigh wave phase velocity maps using the Eikonal equation and a preliminary investigation of Rayleigh wave attenuation. We first measure Rayleigh wave phase delays and amplitudes from all earthquakes and broadband stations that meet our selection criteria. The measurement and analysis approaches are described in Section 3.3. We utilize all broadband and low frequency vertical component seismic data measured from 1999 to 2020 available through the IRIS Data Management Center. The stations used in this study are plotted on (Fig. 3-1).

Close station spacing and concurrent deployment time windows are crucial for this study because we utilize individual station pairs to measure travel time delays for single events. We allow up to 600 km spacing for our calculations. To evaluate the gradient of travel time in each region for the Eikonal equation, stations must also have a 2D array geometry. This requirement limits the information provided by more common linear seismic arrays.

We use events that have a magnitude greater than or equal to M_w 5.5 and maximum depths of 100 km. Events are sourced from epicentral distances of 20°-120° to minimize contamination from overtone Rayleigh waves (Hariharan et al., 2022). To initially quality control our data, we require each event to have a signal-to-noise ratio of greater than or equal to three for at least two

stations in each region. Russell and Dalton (2022) requires five stations, but we loosen this restriction to achieve more complete data coverage.

2.2 Discretizing Polar Regions

We found that discretizing Antarctica in a grid using longitude and latitude introduced problems with inconsistent grid sizes as the difference between a degree in latitude and a degree in longitude is significant near the poles. To compensate for this problem, we translated our study region to a Pacific equatorial location using a coordinate rotation. We also transformed all event location data during initial processing. This transformation preserves any distance between two points and has no negative effect on the quality of our phase velocity and attenuation measurements. Following the completion of the inversions, we translate the locations of resulting grid points back to Antarctica for the purpose of displaying and interpreting our results.

We calculate average phase velocity and attenuation values in 0.25° cells at equatorial latitudes and longitudes. After translating results back to polar locations, we interpolate our results onto a gridded surface of $0.5^\circ \times 1^\circ$ polar longitudes and latitudes allowing interpolation within 1.5° of a data point to create relatively smooth maps (Wessel et al., 2013).

2.3 Eikonal Rayleigh Wave Tomography

We use the Automated Surface-Wave Measurement System (ASWMS) software from Jin and Gaherty (2015) updated by Russell and Dalton (2022) to measure Rayleigh wave phase and amplitude and calculate Rayleigh wave phase velocities. We consider 15 periods between 20s and 150 s.

To measure phase delays, ASWMS utilizes cross-correlation of waveforms at nearby stations, yielding a time lag between all eligible pairs of stations. Amplitudes are single-station

measurements that are determined from the square root of the autocorrelation function at each station. More details of the measurement procedure can be found in Russell and Dalton (2022).

The Helmholtz equation, or Helmholtz tomography, forms the basis for our approach to solving for phase velocity maps from the measured travel time $\tau(x, y)$ and amplitude $A(x, y)$ fields:

$$\frac{1}{c(x, y)^2} = \frac{1}{c'(x, y)^2} - \frac{\nabla^2 A(x, y)}{\omega^2 A(x, y)}$$

where x is longitude, y is latitude, ω is angular frequency, $c'(x, y) = |\nabla\tau(x, y)|^{-1}$ is the apparent, or Eikonal, phase velocity, and $c(x, y)$ is the structural, or Helmholtz, phase velocity (Russell & Dalton, 2022). When the second term on the right-hand side is small due to high frequency and/or a small amplitude Laplacian, the Helmholtz equation reduces to the Eikonal equation, and lateral variations in phase velocity can be determined from only the gradient in travel time.

In practice, we apply the Helmholtz equation to phase and amplitude measurements from each earthquake, resulting in a phase velocity map for each earthquake, and those individual maps are ultimately averaged to yield the composite phase velocity maps shown here. We consider six separate regions, which we consolidate to the final four regions we present in this paper. In all regions, except for the South Shetland Islands (D), we have at least 100 contributing events. In the region that encompasses the northern Transantarctic Mountains (region B), 814 events contribute phase velocity results. Region A has 304 events; region C has 400 total events; and region D has 9 events contributing results. We acknowledge concerns with region D having very few events contributing to phase velocity measurements, but we still provide results as they appear to have an interesting pattern.

In this study, we focus on presenting and interpreting the Eikonal phase velocity maps. The assumption that structural and apparent phase velocity are equivalent is certainly reasonable at short periods but likely less valid at our longest periods. However, calculating structural phase velocity requires a robust estimate of the Laplacian of the amplitude field. Amplitude is typically a noisier measurement than travel time, owing to inherent wavefield complexity as well as the sensitivity of amplitudes to errors in reported instrument gain factors. Numerically differentiating the amplitude field two times inevitably worsens the noise level. Thus, we delay an interpretation of the Helmholtz maps for future work when we can curate a high-quality amplitude data set. Interpretations of the Eikonal maps are themselves valuable. Previous studies (Lin & Ritzwoller, 2011) have shown that the differences between Eikonal and Helmholtz tomography are second-order, even at long periods, with Helmholtz tomography largely refining the features imaged by Eikonal tomography.

Eikonal and Helmholtz tomography work best when applied at seismic arrays that are truly two-dimensional in layout (i.e., similar dimension in latitude and longitude) with evenly spaced stations. Since those conditions are not met in all Antarctic seismic deployments, we adjust many quality control parameters from the ASWMS workflow in order to utilize as much of the available seismic data as possible. We adjust parameters related to the required ray coverage within grid spaces and permitted tomography inversion error. For example, we only require 10 rays to contribute to each grid cell. We also increase the largest variance of inversion error as well as the number of standard deviations above which to discard measurements from 2 to 3. Through testing, we also found that smoothing of the composite (stacked) Eikonal maps caused the loss of resolution of regional-scale features, so we chose to not apply smoothing of the Eikonal map, contrary to the method of Jin and Gaherty (2015).

2.4 Seismic Attenuation

Although our focus is phase velocity, we also calculate Rayleigh wave attenuation using the travel time and amplitude fields of Antarctica. The toolbox we use to do this was expanded by Russell and Dalton (2022) from the original ASWMS workflow (Jin & Gaherty, 2015) to incorporate the transport equation first derived by Lin et al. (2012). It is publicly available as the ASWMS-Q toolbox. The goal of this methodology is to isolate the effects of anelasticity on Rayleigh wave amplitude from source, receiver, and focusing effects (Russell & Dalton, 2022). The transport equation is

$$2\nabla\tau(x, y) \cdot \frac{\nabla\beta(x, y)}{\beta(x, y)} - \frac{2\alpha(x, y)}{c(x, y)} = 2\nabla\tau(x, y) \cdot \frac{\nabla A(x, y)}{A(x, y)} + \nabla^2\tau(x, y)$$

where $\alpha(x, y)$ is the anelastic attenuation coefficient and $\beta(x, y)$ is the local site amplification (Lin et al., 2012; Russell & Dalton, 2022). The attenuation coefficient is related to Rayleigh wave attenuation Q^{-1} , by $\alpha = \omega/2UQ$, where U is group velocity. The transport equation relates the unknown parameters, α and β , on the left-hand side to the spatial derivatives of the measured travel time and amplitude on the right-hand side.

Applying the transport equation to real data takes advantage of the fact that the first term on the left-hand side depends on propagation azimuth, but the second term does not. In practice, the quantity on the right-hand side is constructed for each earthquake, and therefore each earthquake contributes a data point corresponding to its propagation azimuth. If the study region has adequate azimuthal coverage by earthquakes, the individual data points together define a sinusoidal curve from which the amplification gradient ($\frac{\nabla\beta(x, y)}{\beta(x, y)}$) and the attenuation coefficient (α) can be estimated. Since this process requires adequate azimuthal coverage, binning of

adjacent pixels can create more robust results, but does decrease resolution from corresponding phase velocity maps (Russell & Dalton, 2022).

The challenges of imaging attenuation structure with Rayleigh waves, especially at the regional scale, are well known, and we found Antarctica to be no exception. The imaging challenges are largely due to the factors that influence amplitude in addition to attenuation, especially horizontal refraction (focusing/defocusing) during propagation, and to imperfect knowledge of the instrument response (Dalton & Ekström, 2006). Calculations done with the Antarctic data set sometimes yielded negative attenuation values, which could either be indicative of near-zero attenuation or a problem with the methodology. We took substantial effort to manipulate parameters in the ASWMS-Q toolbox to find attenuation values that are physically possible (non-negative) and provide reasonable geographic coverage. For these tests, we focused on the eastern portion of region B (the Transantarctic Mountains) as this region has the most complete station coverage, ideal array geometry, and concurrent station occurrence. Despite these data qualifications, we struggled with finding non-negative attenuation values, particularly in the period range 36s to 63s.

We found that varying parameters related to cross-correlation, phase velocity stacking, and attenuation measured decreased the number of negative attenuation values we would find in a region. In particular, during cross correlation, allowing a Gaussian filter width range of 0.06-0.1 instead of just 0.1, as well as increasing the number of cycles used for the Gaussian wavelet to four reduced the presence of negative attenuation. We also found that decreasing the minimum ratio allowed between a good and bad event during phase velocity stacking to 0.5 improved our data coverage. Directly related to the attenuation calculation, we found that decreasing the

minimum number of grid cells required to use an event to 10 also improved our attenuation results.

Additionally, many of our regions are adjacent to coastlines, so it is possible that focusing and defocusing affects are contributing significantly to our concerning results (Russell & Dalton, 2022; Russell & Gaherty, 2021).

3 Results Overview

3.1 Phase Velocity Maps

We calculate Rayleigh phase velocities with the Eikonal equation in four regions of Antarctica chosen for adequate seismic array coverage with concurrent timing (Fig. 3-1). These regions consist of:

- (A) The Gamburtsev Subglacial Mountains (GSM) in East Antarctica (Section 4.1).
- (B) The northern Transantarctic Mountains (TAM) with the adjacent Ross Ice Shelf and Terror Rift (Section 4.2).
- (C) A portion of West Antarctica that includes the West Antarctic Rift System (WARS), Marie Byrd Land (MLB) and the Ellsworth-Whitmore Mountains (EWM) (Section 4.3).
- (D) The northernmost Antarctic Peninsula (AP) in the proximity of the South Shetland Islands (Section 4.4).

These regions all exhibit interesting lateral and depth variation in phase velocity which we analyze in-depth in section four. We also directly compare our results with the phase velocities used by Shen, Wiens, Anandakrishnan, et al. (2018) that include results from Heeszel et al. (2016). Figures of these comparisons are provided in the supplementary material (Fig. SC-1c to

S3-15c). We reference the standard deviations of individual measurements at each grid point to avoid overinterpretation (Fig. SC-1b to SC-15b).

3.2 Preliminary Attenuation Maps

We resolve seismic attenuation maps for all regions presented here except for region D, the South Shetland Islands. To discuss broad regional trends in attenuation, we present regionally-averaged attenuation values for the GSM, TAM, and West Antarctica (Fig. 3-2). In general, the GSM exhibit attenuation significantly lower than PREM at periods greater than 40 s. The northern TAM has low attenuation (relative to PREM) from 36 s to 73 s, but has attenuation significantly higher than PREM greater than 97 s. In West Antarctica, the regions of the WARS and MBL manifest a large variation in attenuation values. Below 97 s, this region exhibits attenuation values within one standard deviation of PREM. At periods 97 s and above, this region exhibits very high attenuation values ≥ 0.025 . Although relatively high attenuation values are expected in a recently tectonically active region like West Antarctica, attenuation values that significantly exceed 0.025 for a majority of the region is more extreme than attenuation found in regions with similar tectonic histories, such as the western United States (e.g. Bao et al., 2016). We believe these values may be unreliable due to poor resolution in this region. In section five, we discuss regional attenuation results in the context of the regional phase velocities and geologic setting.

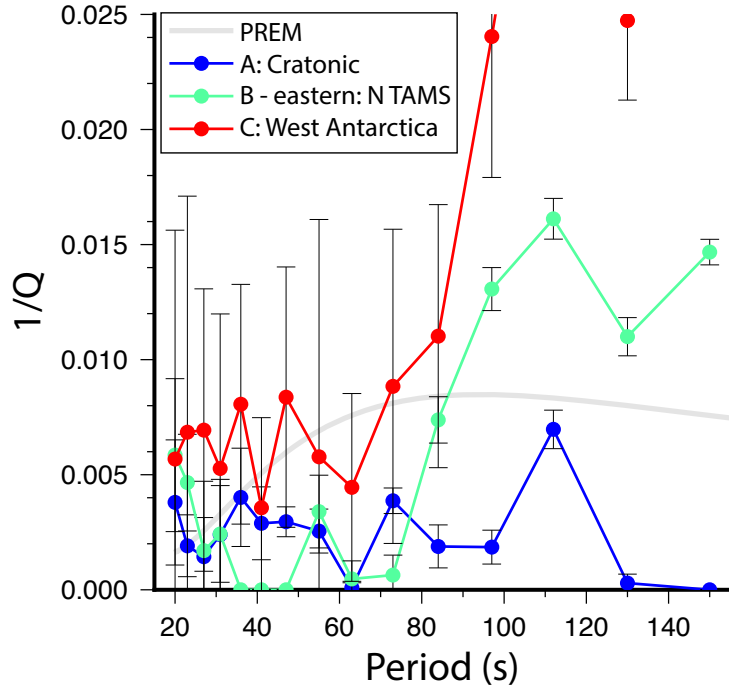


Figure 3-2. Regionally averaged attenuation values

Regionally averaged attenuation values from three regions (A,B,C). For the regions A (blue) and C (red), we average all attenuation values in the regions to represent cratonic and West Antarctica, respectively. Region B samples the TAM and the Ross Ice Sheet. Instead of averaging this entire region, we average only the TAM attenuation values (green).

Despite significant effort in parameter tuning, negative attenuation values persisted in some regions, predominately in the GSM and TAM at intermediate periods (41-73 s). Based on tectonic setting, these regions, especially cratonic GSM, are expected to have lower attenuation than West Antarctica. For this study, we assume that all negative attenuation measurements indicate near-zero true seismic attenuation. We present preliminary results for all regions of negative attenuation as zero attenuation.

4 Regional Phase Velocities

Throughout Antarctica, we observe seismic structures that are broadly consistent with previous work (e.g. Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018; Wiens et al., 2023). In the Gamburtsev Subglacial Mountains (GSM) (Fig. 3-3,4) and northern Transantarctic

Mountains (TAM), we observe high seismic velocities (Fig. 3-5,6). In contrast, we observe relatively slow phase velocities in West Antarctica (Fig. 3-7,8), although significant heterogeneity exists laterally and with depth in this region. We also observe subduction features at the northernmost tip of the Antarctic Peninsula (Fig. 3-9).

While our results are consistent with previous seismic studies, we resolve more detailed features than previous work (e.g. Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018). In this section, we individually address each region we analyze (Fig. 3-1A-D) by summarizing the geologic setting and presenting results of Rayleigh phase velocity and attenuation in the context of prior seismic and tectonic knowledge.

4.1 East Antarctica: Gamburtsev Subglacial Mountains (A)

4.1.1 Geologic Setting & Previous Work

The East Antarctic upper mantle exhibits fast seismic velocities with variable lithospheric thickness that is 200 km thick on average, indicating that the upper mantle is also relatively cold and depleted (Wiens et al., 2023 & references therein). Geologic sampling of East Antarctica is sparse, but it is assumed to represent a Precambrian shield based on its cratonic crustal and lithospheric thicknesses (Wiens et al., 2023 & references therein). Geologic sampling from the coasts of Antarctica indicates Archean and Proterozoic age rocks, but it is unclear how the evolution of the Antarctic interior relates to other ancient continents (Lloyd et al., 2020 & references therein). Some previous work infers that East Antarctica was formed during the amalgamation and breakup of Gondwana, indicating that East Antarctica consists of an Archean nucleus that predates the Rodinia supercontinent (Lloyd et al., 2020 & references therein). However, the timing of the final assembly of East Antarctica is debated (Lloyd et al., 2020 & references therein).

East Antarctica also exhibits high elevations (Lloyd et al., 2020 & references therein) that may be supported by thick crust (Wiens et al., 2023 & references therein) and persist due to low erosion rates from ice cover (Wiens et al., 2023 & references therein). The East Antarctic crust likely first thickened during Proterozoic orogenic events (Wiens et al., 2023 & references therein). Previous work observed exceptionally thick crust in the GSM, with thicknesses up to 55 km, which may support its high elevations (Fig. 3-1) (Shen, Wiens, Anandakrishnan, et al., 2018; Wiens et al., 2023 & references therein). However, seismic and gravity estimates of GSM crustal thickness do conflict, indicating either a dense deep crustal root or a shallow crust-mantle boundary underlain by anomalously high-density upper mantle (Shen, Wiens, Anandakrishnan, et al., 2018 & references therein).

The Antarctic upper mantle exhibits some seismic velocity variations, but these are generally confined to the upper 350 km (Lloyd et al., 2020). Throughout East Antarctica, fast shear velocities extend to up to 350 km in the interior (Lloyd et al., 2020). Previous seismic studies find that the East Antarctic lithosphere is fastest and thickest beneath the Gamburtsev Subglacial Mountains (GSM), with exceptionally fast velocities extending to 300 km (Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018). This craton-like structure infers that the GSM has remained untouched by Meso- and Cenozoic Antarctic tectonic activity (Lloyd et al., 2020; Wiens et al., 2023 & references therein).

Previous work finds that lateral variations within the GSM are typically relatively small (Wiens et al., 2023 & references therein). However, Shen, Wiens, Anandakrishnan, et al. (2018) finds evidence of a 2% shear velocity reduction in the uppermost mantle (~80-120 km depth) underlying the highest elevations of the GSM. This region may represent the presence of anomalous mantle compositions due to changes in temperature, variations in mantle depletion, or

variations in mantle mineralogy. Based on the tectonic setting of the GSM and discrepancies between observed seismic structure and high gravity observations in the GSM, Shen, Wiens, Anandakrishnan, et al. (2018) propose that this low velocity zone could be due to mantle eclogization remnant of ancient continental collision (Shen, Wiens, Anandakrishnan, et al., 2018 & references therein).

4.1 2 Results & Interpretation

Compared to phase velocity maps from West Antarctica, the GSM exhibits relatively homogenous phase velocities at periods above 23 s, although we do observe consistent lateral heterogeneity in the crust and mantle (Fig. 3-3). Our phase velocity results are generally consistent with those of Shen, Wiens, Anandakrishnan, et al. (2018) (Fig. SC-1c to SC-15c), although our results provide more detail and, consequently, indicate slightly different spatial patterns.

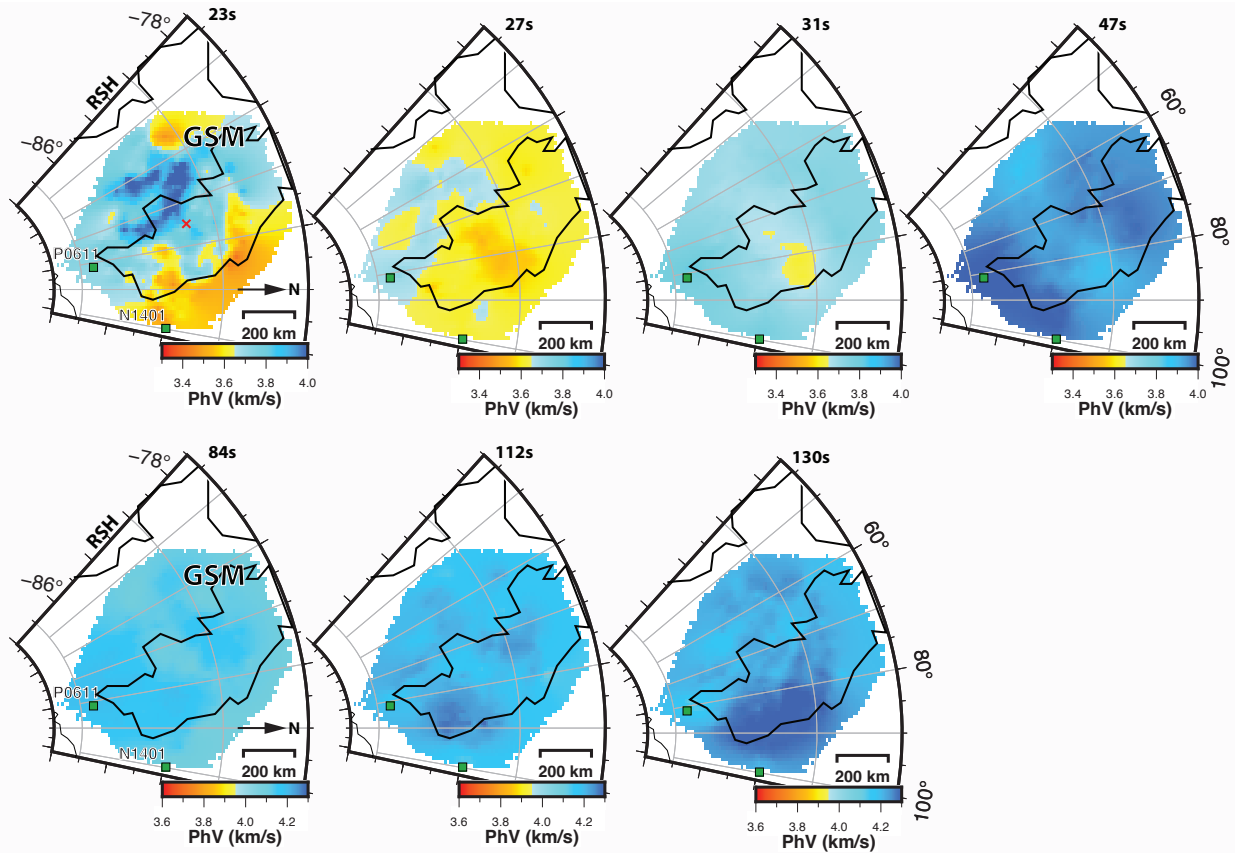


Figure 3-3. Rayleigh wave phase velocity results for region A: the Gamburtsev Subglacial Mountains (GSM)
 Rayleigh wave phase velocity results for region A: the Gamburtsev Subglacial Mountains (GSM) in East Antarctica at selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A. Note a shift in phase velocity color scale at 55 s. Locations from Krueger et al. (in prep.) shown as green squares and locations of sampled phase velocities for Figure 3-4 are indicated by red x's.

In general, we find that the GSM exhibits high, but spatially variable, phase velocities at periods of 20-23 s (Fig. 3-3). This is consistent with high shallow crust shear velocities observed by Shen, Wiens, Anandakrishnan, et al. (2018). Beginning at 27 s, there is a consistent increase in phase velocity with period. In all maps of intermediate periods (27 – 47 s) there are clearly lower velocities in the center of the GSM, but this is most obvious in the 31 s map due to the color scale used. The simplest interpretation of this observation is thicker crust in the center of the GSM. Previous work estimates extremely thick crust in the GSM (~55 km) (Shen, Wiens, Anandakrishnan, et al., 2018).

We continue to see the pattern of slightly lower velocities in the center of the GSM with increasing period up to 112 s, although the velocity difference is less extreme. The presence of this anomaly to higher periods indicates that this signal could possibly have a mantle source. Interestingly, Shen, Wiens, Anandakrishnan, et al. (2018) argues for a low velocity zone with a 2% velocity reduction from 80-120 km within GSM. At even higher periods (130 s and 150 s), we observe higher velocities towards the south of the region. The general lithospheric heterogeneity we observe is consistent with global observations of local heterogeneity in the cratonic lithosphere (e.g. Krueger et al., in prep.).

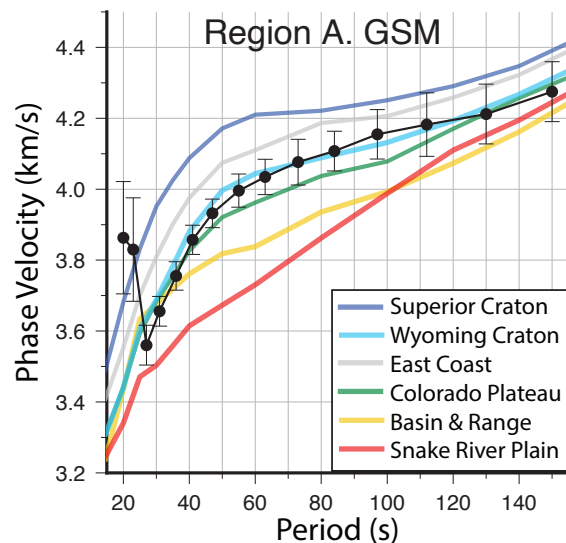


Figure 3-4. GSM average phase velocities

Phase velocities averaged within a 50 km radius of red x's in Figure 3-3 are plotted in black. Representative phase velocities for regions of the US are plotted in the background for comparison (Babikoff & Dalton, 2019). Error bars represent one standard deviation of averaged phase velocities.

The high velocities we see at all periods beneath the GSM are consistent with estimates of lithospheric thickness of approximately 190 km (Lloyd et al., 2020) and observations of relatively high velocities underlying the GSM extending to 350 km depth (Lloyd et al., 2020) (Fig. 3-4). By plotting a sample of the GSM phase velocities, we see that high velocities from 36 s to 150 s are indicative of lithospheric mantle comparable to the Wyoming craton or the

Colorado Plateau in the US (Fig. 3-4). This result is surprising as Krueger et al. (in prep.) finds higher upper mantle velocities at locations P0611 and N1401 than in the Wyoming craton. It is possible this inconsistency is because in Fig. 3-4 we sample the center of the GSM, where velocities appear slightly lower at mid-lithospheric depths.

4.2 East/West Boundary: Transantarctic Mountains and Ross Ice Shelf (B)

4.2.1 Geologic Setting & Previous Work

The formation of the TAM is one of the primary tectonic events impacting East Antarctica (Lloyd et al., 2020 & references therein; Shen, Wiens, Anandakrishnan, et al., 2018). Geologic samples from the Transantarctic Mountains (TAM) indicate Proterozoic terranes (Lloyd et al., 2020 & references therein). Various studies have proposed Cenozoic uplift mechanisms for these terranes. Among these mechanisms are thermal mantle support, isostatic support through crustal thickness or other density variations, plateau collapse, and flexural uplift (Shen, Wiens, Anandakrishnan, et al., 2018 & references therein; Wiens et al., 2023 & references therein).

This study focuses on the northern TAM (Fig. 3-1B) which represents the high elevations of Victoria Land (Wiens et al., 2023 & references therein). Previous work finds the crust of TAM in region B to be 40 km thick in the northernmost TAM and up to 50 km thick towards the south of region B (Shen, Wiens, Anandakrishnan, et al., 2018). The TAM lithosphere in region B varies in thickness from approximately 180 km inland to ~60 km at the coastline (Lloyd et al., 2020). The ANT-20 models shows that high shear velocities persist to over 250 km depths throughout TAM in most of region B except for in the northernmost TAM (Lloyd et al., 2020). Notably, near the coastline of northern Victoria Land, previous studies observe slow seismic velocities in the mantle above 150 km (Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018; Wiens et al., 2023 & references therein). This observation prompts Shen, Wiens,

Anandakrishnan, et al. (2018) to consider lithospheric modification in the northernmost TAM due to lithospheric foundering or lateral interactions with the West Antarctic Rift System (WARS). It also appears that moderate velocities may encroach from the Ross Ice Sheet region into the TAM towards the top (south) of region B (Lloyd et al., 2020).

Furthermore, the TAM exhibit significant along-strike variations in surface and LAB topography that indicate a complex geologic history (Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018; Wiens et al., 2023). While we only study the northern TAM, other regions of the TAM exhibit particularly interesting features. For example, the southern TAM shows strong evidence of lithospheric foundering (Shen, Wiens, Anandakrishnan, et al., 2018).

The TAM serve as a boundary between East and West Antarctica (e.g. Lloyd et al., 2020) where the cratonic lithosphere of the TAM abuts the warm upper asthenospheric mantle of West Antarctica (Wiens et al., 2023 & references therein). At the western side of the boundary, there is a strong shear velocity gradient between West Antarctica and the TAM. Near the Ross Embayment, this gradient is strongest where a low shear velocity anomaly extends adjacent to the TAM and recent extension has occurred on the Terror Rift (Lloyd et al., 2020). There is also a significant gradient in crustal thickness at the boundary of the TAM and the Ross Embayment with crustal thickness decreasing from ~45 km to 20 km below the Ross Ice Sheet (Shen, Wiens, Anandakrishnan, et al., 2018).

The strong gradients in seismic characteristics at the boundary of the TAM with the Ross Embayment is emphasized by a broad connected band of low velocities that originates in the Macquarie Triple Junction (Lloyd et al., 2020; Wiens et al., 2023 & references therein). The Terror Rift is a key feature of this low velocity band with slow velocities likely related to warm mantle temperatures and decompression melting from recent extension (Shen, Wiens,

Anandakrishnan, et al., 2018). The Terror Rift is bounded to the north and south by active volcanic provinces (Lloyd et al., 2020; Wiens et al., 2023). To the south, lies Ross Island which contains the active volcano Mt. Erebus. Mt Erebus is proposed to be sourced from a mantle plume but this is debated (Wiens et al., 2023 & references therein). In addition to a potential plume in the Terror Rift region, the extensive history of subduction along the West Antarctic margin may have left remnant metasomatism in the upper mantle that could contribute to low seismic upper mantle velocities (Wiens et al., 2023 & references therein).

Overall, the Ross Embayment has crust as thin as 15-30 km. Thin crust in the Ross Embayment is likely a result of extensive Cretaceous backarc extension (Jordan et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018). Extension in the Ross Embayment stopped approximately 11 Ma (Lloyd et al., 2020 & references therein) and there is evidence of volcanic centers spread throughout the eastern Ross Embayment (Jordan et al., 2020 & references therein). Despite shared extension, the West Antarctic Rift System, Marie Byrd Land, and the Ross Embayment all exhibit different magnetic signatures indicating compositional differences (Jordan et al., 2020 & references therein).

The lithosphere in the Ross Embayment is relatively thin, ranging from approximately 60 km near the northern TAM and 80 km towards the WARS (Lloyd et al., 2020). Previous work observes generally slow mantle velocities below the Ross Ice Sheet (Lloyd et al., 2020). These velocities are slowest within the band of low velocities that continue south of the Terror Rift (Lloyd et al., 2020), resulting in a general velocity increase eastward across the Ross Embayment (Lloyd et al., 2020; Wiens et al., 2023). Despite a shallow LAB, Shen, Wiens, Anandakrishnan, et al. (2018) observes as high velocity mantle lid in the uppermost mantle in the Ross

Embayment and West Antarctic Rift System that indicates upper mantle cooling since the most recent rifting episodes in those regions.

4.2.2 Results & Interpretation

Region B exhibits significant variation in phase velocity laterally and with depth (Fig. 3-5). Compared to the phase velocities used by Shen, Wiens, Anandakrishnan, et al. (2018), we have relatively consistent results at periods below 55 s, although we do resolve more detail. At periods of 55 s and above we see more complexity and extremity of velocities than Shen, Wiens, Anandakrishnan, et al. (2018). While very broad spatial and depth trends are relatively consistent between the two phase velocity data sets, some periods are inconsistent with spatial patterns from Shen, Wiens, Anandakrishnan, et al. (2018) (Fig. SC-1c to SC-15c). While we do believe our results represent Earth structure in a detailed manner, some of the inconsistency with Shen, Wiens, Anandakrishnan, et al. (2018) may be due to uncertainties in our data (i.e. the very high velocities in the TAM; e.g. Fig S3-5).

At periods 20-23 s, the Ross Ice Shelf exhibits high phase velocities relative to the TAM. This is consistent with the thin crust (~20 km) and “high velocity lid” that Shen, Wiens, Anandakrishnan, et al. (2018) observe below the Ross Ice Shelf and West Antarctic Rift System. These relatively high velocities in the Ross Ice Shelf are observed below 55 s and are more extreme on the eastern side of the RIS. This is consistent with ~60 km lithospheric thickness reported in the region with slightly thinner lithosphere at the boundary with TAM (Lloyd et al., 2020). Phase velocities in the Ross Ice Shelf and TAM generally equilibrate from 27-36 s. At periods of 41 s and greater, the northernmost TAM exhibits notably high velocities that are maintained for all greater periods.

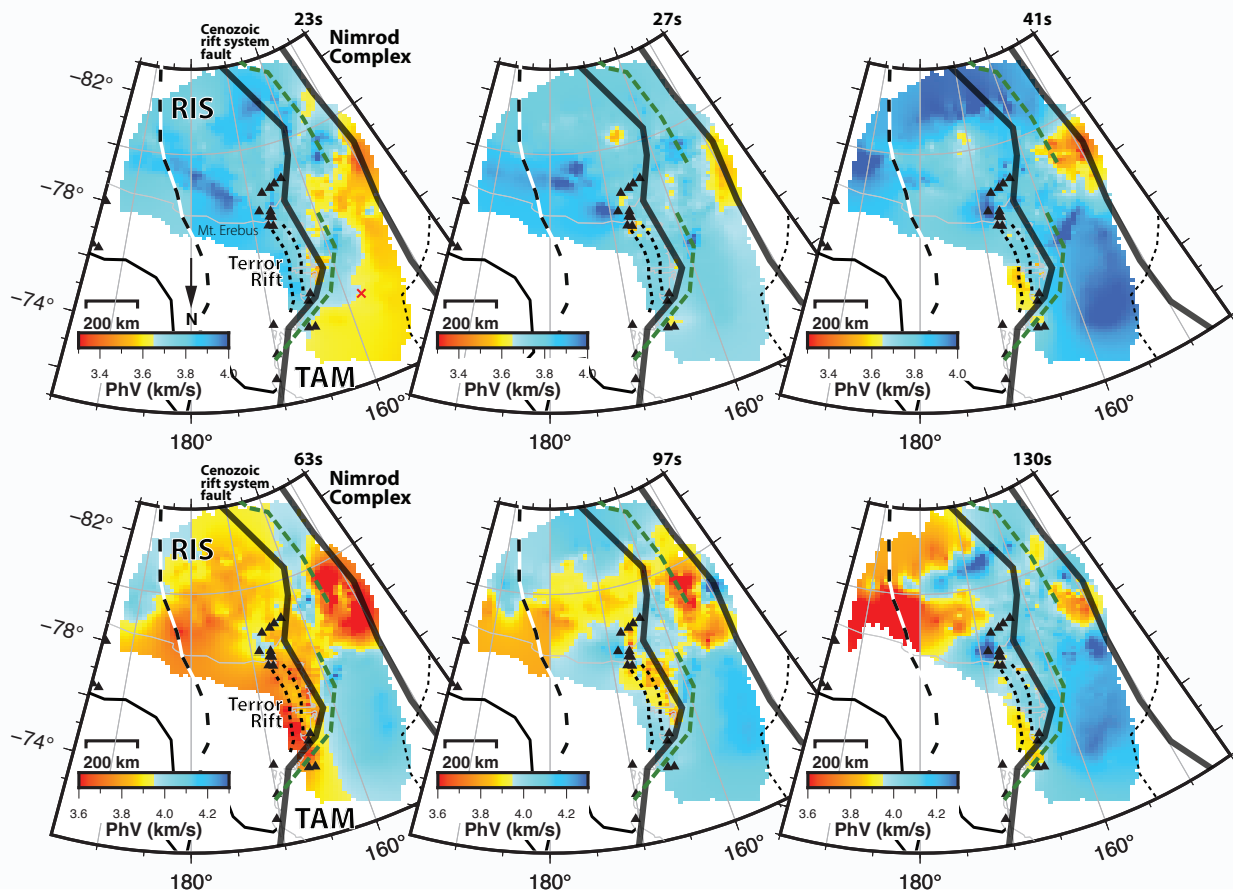


Figure 3-5. Rayleigh wave phase velocity results for region B: the northern Transantarctic Mountains
 Rayleigh wave phase velocity results for region B: the northern Transantarctic Mountains (TAM) and adjacent Ross Embayment at selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 1A. Note a shift in phase velocity color scale at 55 s. Locations of sampled phase velocities for Figure 3-6 are indicated by red x's.

In the TAM adjacent to the Ross Ice Shelf our maps show a consistent, exceptionally low velocity region that appears to coincide with the locations of Cenozoic rifts (green). It is tempting to interpret this observation, but this region has relatively high standard deviations in phase velocity values (Fig. SC-1b to SC-15b). While we are not confident in interpreting this feature, it is a location of interest for future work as data quality and quantity improves.

We also see an interesting pattern below the Ross Ice Shelf (RIS). At intermediate periods (36 s) there are relatively high velocities in the center of the RIS. At higher periods, the center of the RIS appears to have low seismic velocities, which become more extreme with

increasing period. These low velocities appear to be encroaching from the east, although the exact pattern of phase velocities within the RIS is inconsistent between periods. Additionally, while we do observe slow velocities at the RIS/TAM boundary at some periods, we do not consistently observe the low velocity “band” reported by Lloyd et al. (2020). These inconsistencies could be due to heterogeneity in structure with depth or simply attributed to uncertainty in our results. We also observe substantially slower phase velocities at high periods (130-150 s) than Shen, Wiens, Anandakrishnan, et al. (2018).

We find the local seismic structure underlying Mt. Erebus (on Ross Island) and the Terror Rift particularly fascinating. We are confident in variations in these regions as it is well instrumented, especially Ross Island (Fig. 3-1). At low periods (20-23 s) we see no indication of velocity anomalies at Mt. Erebus or the Terror Rift. At 27 s, we see isolated regions of lower phase velocity in the volcanic provinces north and south of the Terror Rift. In the period range from 31 s to 63 s we see lower phase velocities in the northern Terror Rift and adjacent volcanic regions. At these periods, phase velocities near Mt. Erebus appear relatively high. At 73 s, this pattern reverses and we see relatively high phase velocities in the northern Terror Rift and low phase velocities in the southern Terror Rift. Interestingly, the Mt. Erebus region maintains significant variation with some higher velocities. Consistent with our observations from 73 s to 97 s, Shen, Wiens, Anandakrishnan, et al. (2018) also see lower velocities in the Terror Rift Region, although our map is significantly more complex.

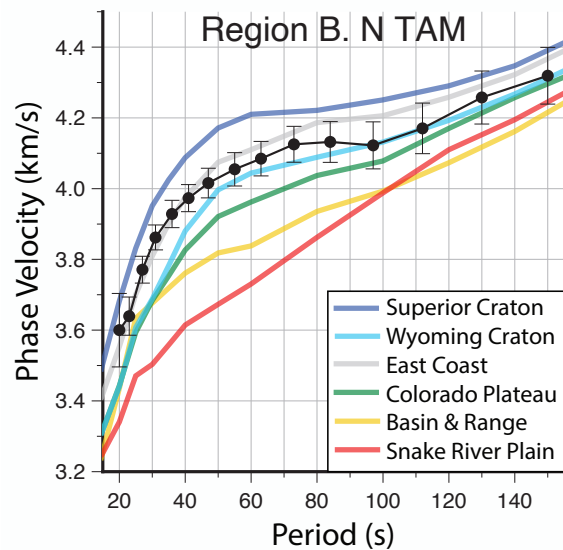


Figure 3-6. TAM average phase velocities

Phase velocities averaged within a 50 km radius of red x's in Figure 3-5 are plotted in black. Representative phase velocities for regions of the US are plotted in the background for comparison (Babikoff & Dalton, 2019). Error bars represent one standard deviation of averaged phase velocities.

We isolate a region of the northern TAM where we have high quality results to compare with tectonic regions in the US (Fig. 3-6). Phase velocities in the TAM coincide clearly with the East Coast until 84 s. Above 84 s, phase velocities are lower although they are within one standard deviation of the East Coast. In these periods, TAM phase velocities are closer to Wyoming Craton phase velocities. This connection is intriguing because lithospheric foundering has been proposed in both the Wyoming craton (Carlson et al., 2004; Eilon et al., 2018) and the northernmost TAM (Shen, Wiens, Anandakrishnan, et al., 2018).

4.3 West Antarctic Rift System, Ellsworth-Whitmore Mountains, and Marie Byrd Land (C)

4.3.1 Geologic Setting & Previous Work

Seismic characteristics of West Antarctica are similar to those of other regions with extensive Meso- and Cenozoic rifting such as the western United States (US) (Wiens et al., 2023 &

references therein). These characteristics include thin lithosphere, slow upper mantle, and thin crust (Shen, Wiens, Anandakrishnan, et al., 2018) and are indicated in our results as well (Fig. 3-7). Similar to the United States, West Antarctica shows evidence of complex subduction and extensive extension in the Phanerozoic (Lloyd et al., 2020 & references therein). Because of these similarities, we compare the region we study in Antarctica with the velocity structure of select regions in the well-understood United States (e.g. Fig. 3-8). The primary tectonic events in West Antarctica that influenced the locations we study are:

- (1) Subduction along the entire West Antarctic margin that gradually shrunk in extent from Marie Byrd Land to Thurston Island and southward along the Antarctic Peninsula (AP) (Wiens et al., 2023 & references therein).
- (2) Rotation and overall translation of the Ellsworth-Whitmore Mountains (EWM) during the separation of New Zealand from West Antarctica (Wiens et al., 2023 & references therein).
- (3) Extensive Cenozoic extension in the West Antarctic Rift System (Wiens et al., 2023 & references therein).

West Antarctic Rift System

The West Antarctic Rift System (WARS) is an extensive rift system that dominates central West Antarctica. Extension in WARS is thought to be related both to the separation of New Zealand and subduction backarc spreading (Jordan et al., 2020). This extension also resulted in widespread regional volcanism (Jordan et al., 2020). The WARS exhibits little-to-no modern seismicity, extension, or volcanism, indicating that the rift system is dormant (Lloyd et al., 2020 & references therein). The crust in WARS varies from 30 to 45 km thick and is underlain by relatively high seismic velocities in the uppermost mantle, indicating the development of mantle

lithosphere following Cenozoic extension to approximately 60 km depth (Heeszel et al., 2016; Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018; Shen, Wiens, Stern, et al., 2018; Wiens et al., 2023 & references therein). This high velocity lithosphere also extends into the Ross Embayment (Shen, Wiens, Anandakrishnan, et al., 2018). Below the LAB in WARS, slow velocities are observed (e.g. Lloyd et al., 2020).

Ellsworth-Whitmore Mountains

The Ellsworth-Whitmore Mountains (EWM) represent the most stable and oldest tectonic region of West Antarctic (Jordan et al., 2020). The EWM topographically appear as a continuation of high TAM elevations at the boundary of West and East Antarctica (Fig. 3-1). The EWM region also contains the Haag Nunataks (Fig. 3-1) which have Mesoproterozoic age geologic samples and may extend over 100 km² (Jordan et al., 2020) and show little evidence for Phanerozoic deformation (Jordan et al., 2020). The Ellsworth Mountains are extensively folded sediment and contain the highest elevation peak in Antarctica (5 km) (Jordan et al., 2020). The EWM also contain granites emplaced in the Jurassic that may be related to Large Igneous Province magmatism (Jordan et al., 2020). It has been proposed that these mountains have thermal support for their high elevations (Wiens et al., 2023 & references therein).

The crust of the EWM has variable crustal thickness, from 30 to 45 km thick, with indications of a slow mid-crust possibly due to felsic compositions (Shen, Wiens, Anandakrishnan, et al., 2018). The crust between separate mountain ranges in the EWM is thinner than the crust below the highest elevations (Shen, Wiens, Anandakrishnan, et al., 2018). Despite high elevations in the EWM, Shen, Wiens, Anandakrishnan, et al. (2018) observe slow uppermost mantle velocities in the EWM that extends to approximately 120 km depth. They evoke lithospheric modification below EWM from rift-related processes (Shen, Wiens,

Anandakrishnan, et al., 2018). Furthermore, the ANT-20 model also interprets a thin lithosphere of approximately 60 km below EWM (Lloyd et al., 2020).

Marie Byrd Land

Marie Byrd Land (MBL) was also influenced by both extension and subduction processes (Jordan et al., 2020). MBL is a heterogeneous region consisting of sediment deposits, convergent margin arcs, and lava flows. The basement of the MBL is likely metamorphosed Precambrian rock (Jordan et al., 2020). The crust is observed to be approximately 30 to 35 km thick in MBL with a shallow LAB at approximately 60 km depth (Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018).

High elevations in Marie Byrd Land, referred to as the Marie Byrd Land Dome (MBL-Dome), are also related to subduction along the West Antarctic margin. The MBL-Dome is associated with volcanoes that may be of mantle plume origin, which would provide thermal support for high elevations despite thin crust (Jordan et al., 2020 & references therein; Shen, Wiens, Anandakrishnan, et al., 2018 & references therein; Wiens et al., 2023 & references therein). Although, some studies argue that volcanism in the MBL-Dome was sourced from volatile interactions of the historic subducting slab (Jordan et al., 2020 & references therein; Wiens et al., 2023 & references therein). Recent seismic studies are consistent with a mantle plume underlying MBL-dome. They observe slow seismic velocities within the upper mantle below the MBL-Dome (Shen, Wiens, Anandakrishnan, et al., 2018; Wiens et al., 2023 & references therein) with some studies observing low velocities that extend through the mantle transition zone (Lloyd et al., 2020; Wiens et al., 2023 & references therein). The strongest West Antarctic low velocity anomaly in ANT-20 is beneath MBL and may connect to low velocities in

the Amundsen Sea Embayment and along the Antarctic Peninsula (Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018).

Thurston Island

Thurston Island originated at the convergence of the Pacific with Gondwana and shares overlapping history and structure with the Antarctic Peninsula and MBL (Jordan et al., 2020 & references therein; Lloyd et al., 2020 & references therein). Similar to the history of MBL, the slab window resulting from subduction cessation as the Phoenix spreading ridge was subducted caused modification of the mantle underlying Thurston Island (Wiens et al., 2023 & references therein). This modification is seen today in slow seismic velocities in the upper mantle of Thurston Island that appear to connect to similar anomalies underlying the MBL and Antarctic Peninsula (e.g. Lloyd et al., 2020). Understanding the crust and mantle structure of Thurston Island is particularly important for constraining ice sheet melting and isostatic rebound as the adjacent Admundsen Sea Embayment contains 25% of the West Antarctic Ice Sheet (Shen, Wiens, Anandakrishnan, et al., 2018).

4.3.2 Results & Interpretation

The phase velocities we resolve in Region C are complex (Fig. 3-7). Region C spans multiple tectonic provinces within West Antarctica that represent various iterations of compression and extension. It is also likely that underlying formation of a slab window has caused significant modification of the lithosphere from below West Antarctica. Furthermore, compared to regions A and B, resolution in this region is relatively poor (Fig. SC-1b to SC-15b). It is possible that some of the complexity we observe is due to a poorly a constrained travel time field.

At the lowest periods we analyze (20 s and 23 s), we see velocities that are slow in MBL and EWM relative to the WARS. This is likely related to slightly thicker crust in MBL and

EWM than WARS. At periods 20 s to 41 s we see moderate-to-high phase velocities in West Antarctica, likely indicating an underlying lithospheric mantle. This observation is consistent with the “high velocity lid” identified by Shen, Wiens, Anandakrishnan, et al. (2018) across WARS, which we also identify in the Ross Embayment.

Beginning at 47 s and 55 s, we see low phase velocities in the WARS, MBL, EWM, and Thurston Island, although there is a small region of moderate phase velocities near the boundary of WARS and MBL. This is reasonably consistent with lithospheric thickness estimates of 60 km in WARS and MBL (Lloyd et al., 2020).

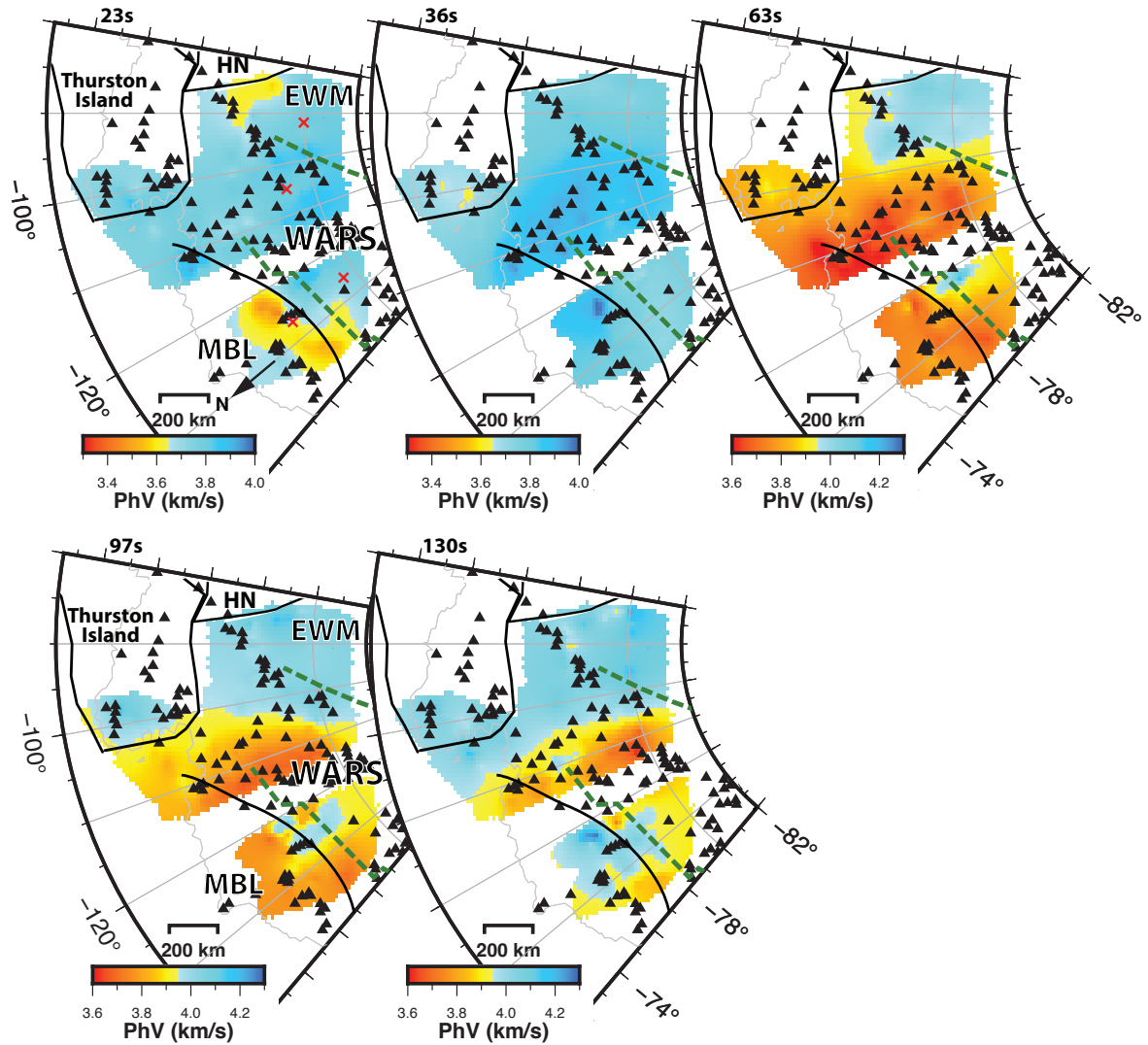


Figure 3-7. Rayleigh wave phase velocity results for region C that includes the West Antarctica

Rayleigh wave phase velocity results for region C that includes the West Antarctic Rift System (WARS), Marie Byrd Land (MBL), Ellsworth-Whitmore Mountains (EWM), and Thurston Island. We provide only selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A. Note a shift in phase velocity color scale at 55 s. Locations of sampled phase velocities for Figure 3-8 are indicated by red x's.

West Antarctic Rift System

Slow phase velocities in the WARS persist from 47 s to 130 s despite significantly variable phase velocities at its boundaries (Fig. 3-7). Although the regional trend of low velocities in the WARS is consistent with phase velocities used by Shen, Wiens, Anandakrishnan, et al. (2018) (Fig. SC-

1c to S3-15c) the magnitude of phase velocities we observe in the WARS are lower than phase velocities from Shen, Wiens, Anandakrishnan, et al. (2018) at all periods.

In the western WARS, we see phase velocities that are nearly constant with period and substantially slower than phase velocities observed in the Basin and Range or Snake River Plain of the US above 84 s. We do note the large standard deviation observed in this region. The eastern WARS exhibits lower variation within the region we averaged phase velocity and has a profile that is more consistent with observations from the US. Shallow phase velocities are consistent with the high phase velocities of the Superior Craton, although phase velocities level off at intermediate periods with the overall trend we observe being most similar to the Snake River Plain or Basin and Range of the US.

Ellsworth-Whitmore Mountains

Interestingly, velocities in the EWM remain low-to-moderate until 73 s, then increase and remain moderately fast until 130 s (Fig. 3-7). This observation may be related to proposed lithospheric modification below the EWM by Shen, Wiens, Anandakrishnan, et al. (2018). It is possible that metasomatism from the adjacent rifting in the WARS modified the EWM lithosphere leading to either a thick layer of metasomatic fluid and minerals in the lithosphere or foundering of the EWM lithospheric keel at this weakened boundary. It is possible we are seeing the bottom of the lower lithosphere below EWM when velocities increase again at 150 s. At all periods, the magnitude of the velocities we observe are similar to phase velocities used by Shen, Wiens, Anandakrishnan, et al. (2018) (Fig. SC-1c to SC-15c) although we resolve more detailed structure. This supports that our observations are congruent with their interpretation of modified lithosphere below the EWM.

Moderate-to-high velocities within the EWM appear to expand laterally, beginning at 63 s. As period increases the expansion of moderate velocities continues to Thurston Island and past the boundary of the EWM with the WARS. After 55 s, the moderate velocity zone at the boundary of the WARS and MBL grows laterally and migrates north towards the highest elevation regions of the MBL-dome. At all periods 55 s and above, the center of the WARS remains low velocity.

The lateral growth of moderate velocities from the EWM towards the WARS and Thurston Island may indicate the presence of complicated lithosphere structure where heat or melt from the interior of the WARS has encroached at mid-lithospheric depths (Krueger et al., 2021).

The moderate-to-high phase velocities we see below the EWM may indicate either a thicker mantle lithosphere as there is no evidence of direct extension within the EWM (Jordan et al., 2020). The standard deviation of phase velocities in EWM is high, but the average of the phase velocities we sample in this region appears similar to the Basin and Range in the US (Fig. 3-8). The phase velocity we measure at 150s in EWM is not like any we compare to in the US, so it is possible our results at these higher periods are not reliable.

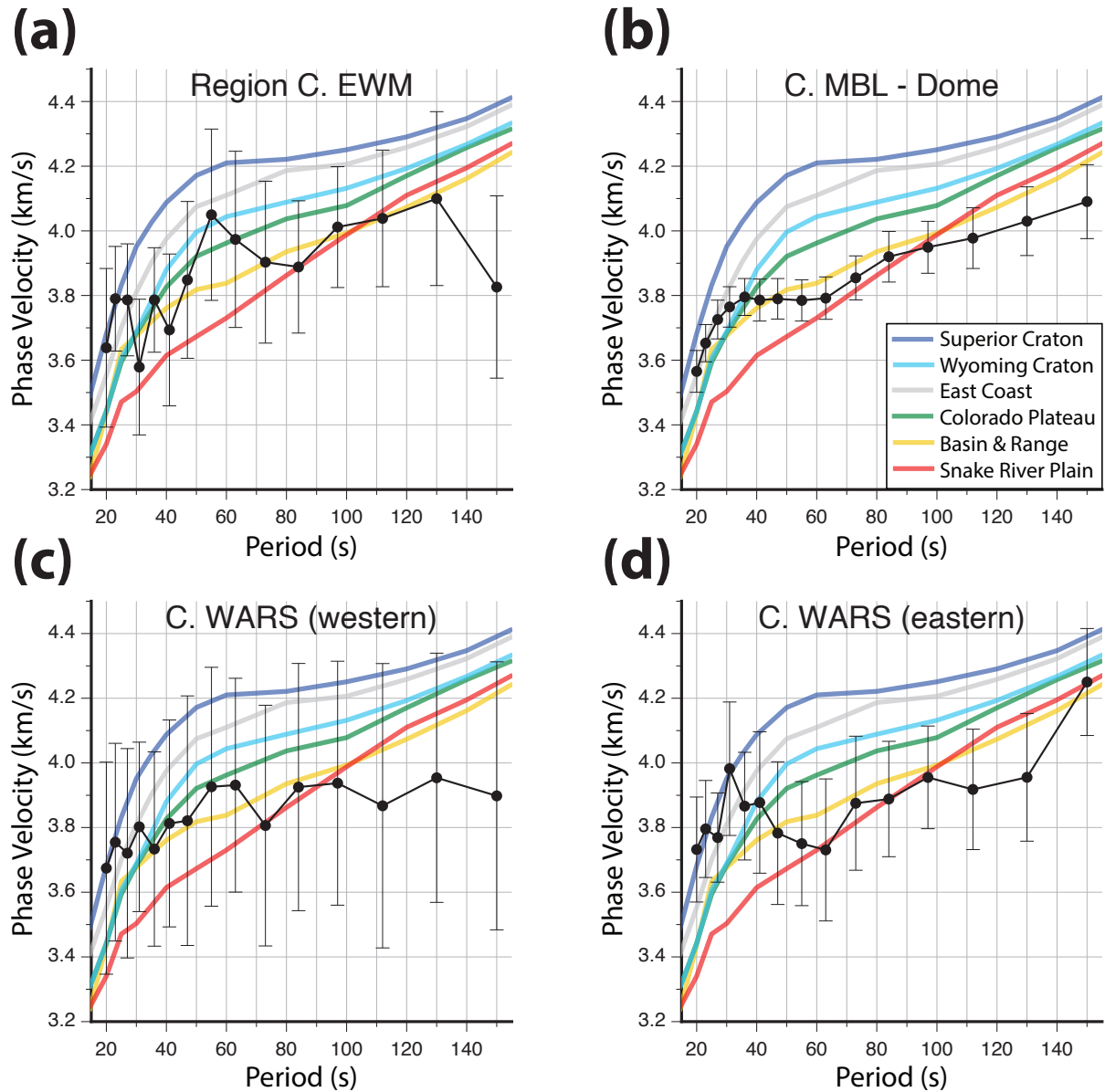


Figure 3-8. West Antarctica average phase velocities

Phase velocities averaged within a 50 km radius of red x's in Figure 3-7 are plotted in black. Representative phase velocities for regions of the US are plotted in the background for comparison (Babikoff & Dalton, 2019). Error bars represent one standard deviation of averaged phase velocities.

Marie Byrd Land

The structure of the MBL-dome is also intriguing. At periods below 97 s, the MBL is low velocity and generally indistinguishable from the WARS with similar phase velocities as Shen, Wiens, Anandakrishnan, et al. (2018) (Fig. 3-7). At 97 s and above, a small region of moderate

velocities migrates towards the highest elevations in the MBL-dome and grows in lateral extent. Similar to the EWM, it is possible that below the Moho, we see indications of heat, hydrous minerals, or melt decreasing seismic velocities in the uppermost mantle, which have not affected deeper lithospheric depths as strongly.

Phase velocities in the MBL dome are relatively high at periods below 41 s, similar to the East Coast of the US. From 41 s to 97 s, these phase velocities remain constant and are relatively low compared to US phase velocities. At these periods, MBL dome phase velocities appear similar to the Southern Rocky Mountains. Above 97 s, we see low phase velocities that are not like any of the locations we sample in the US. It is possible this is due to high mantle temperatures from recent tectonic activity or an artefact of poor resolution at higher periods. These low velocities could also coincide with the theory that the high elevations of the MBL-dome are thermally supported by underlying plume activity (Jordan et al., 2020; Lloyd et al., 2020; Shen, Wiens, Anandakrishnan, et al., 2018).

Thurston Island

Although the region-wide pattern we observe is likely robust, we are skeptical of the moderate-to-high velocities we observed in Thurston Island above 84 s (Fig. 3-7). We are skeptical because this pattern of slightly higher phase velocities in Thurston Island appears to conflict with previous work identifying slow velocities in Thurston Island and this location is at the edge of our resolved region (Lloyd et al., 2020). The magnitude of the velocities we observe are consistent with Shen, Wiens, Anandakrishnan, et al. (2018) above 84 s and below 55 s, but in the period range 55 to 73 s, we observe much lower velocities than Shen, Wiens, Anandakrishnan, et al. (2018).

4.4 Antarctic Peninsula: South Shetland Islands (D)

4.4.1 Geologic Setting & Previous Work

The Antarctic Peninsula (AP) consists of igneous terranes accreted onto Gondwana and hosted subduction arc magmatism until approximately 20 Ma (Jordan et al., 2020). Subduction along the West Antarctica coastline ceased gradually, with the extent of the subducting margin migrating towards the South Shetland Islands (SSI). The subduction of a spreading center led to the formation of a slab window that aided the recent destruction of AP lithosphere (Lloyd et al., 2020 & references therein).

Subduction is ongoing beneath the SSI and has led to rifting along the Bransfield Strait (Fig. 3-1), but has not drastically deformed the AP (Jordan et al., 2020). In general, the AP has slow mantle shear velocities that are underlain by fast shear velocities, but at the northern end of the Peninsula, in the region of the SSI, high velocities are relatively shallow (~200 km) (Lloyd et al., 2020). In the ANT-20 model, velocities at the SSI appear to shift from very low velocities of the AP to higher velocities at depths as shallow as 75-150 km. High velocities in the mantle below the AP likely indicate the Phoenix Plate (Lloyd et al., 2020).

4.4.2 Results & Interpretation

We calculate phase velocities for the northernmost Antarctic Peninsula (AP) in the region of the South Shetland Islands. At most periods, we see evidence of sharp gradients in phase velocity within this region that appear to coincide with the location of the Bransfield Strait (BS) (Fig. 3-9). Although this region is poorly constrained, we are encouraged by features in the phase velocity maps that correspond with tectonic features in the region. This is particularly interesting as regional seismic studies of this region are scarce. It is possible the sharp gradients in phase velocity we see indicates remnant subduction.

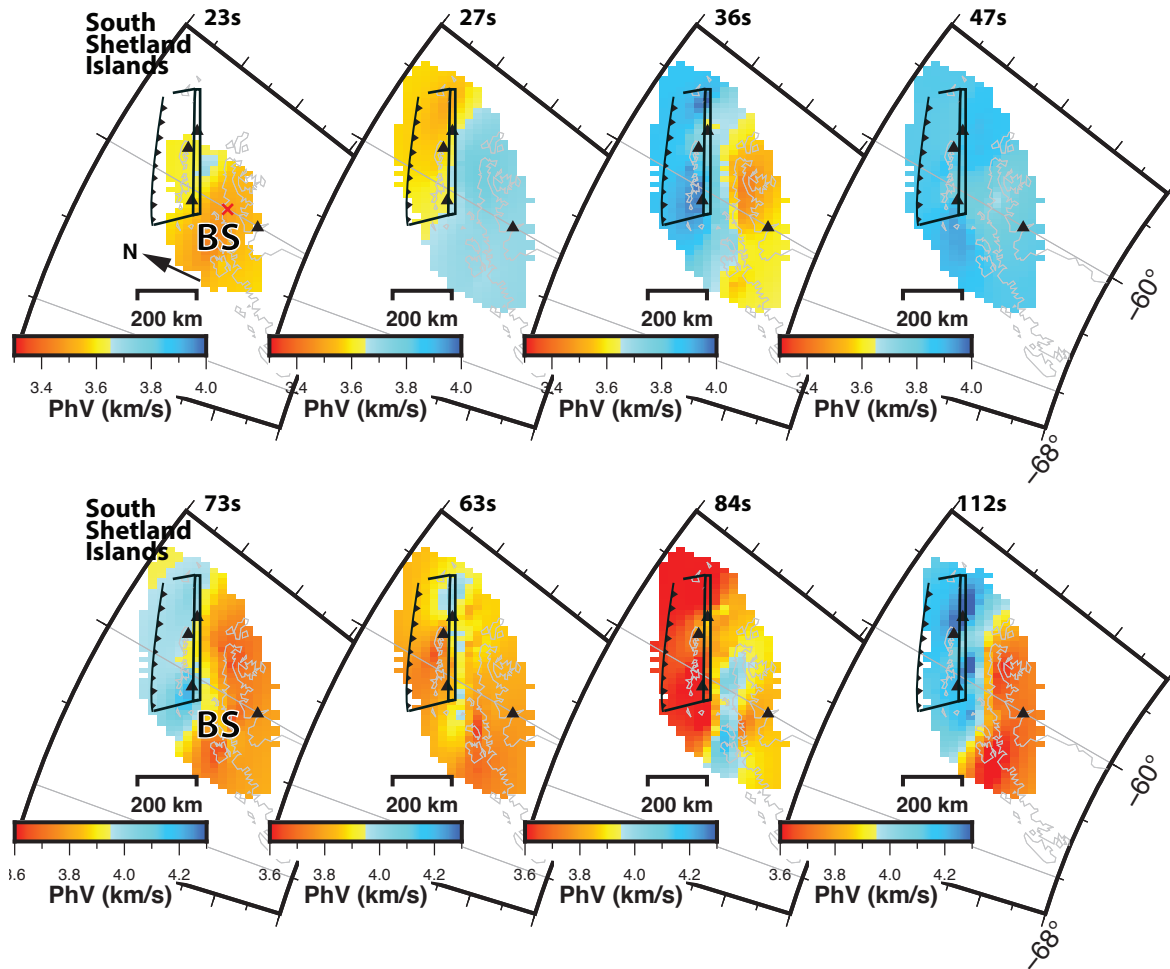


Figure 3-9. Rayleigh wave phase velocity results for region D: the northernmost Antarctic Peninsula
 Rayleigh wave phase velocity results for region D: the northernmost Antarctic Peninsula at selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A. Note a shift in phase velocity color scale at 55s. Locations of sampled phase velocities for Figure 3-10 are indicated by red x's.

We observe moderate velocities at 23 s indicating crustal structure. At 27 s and 31 s, we see a dichotomy of moderately low phase velocities to the north of the BS but moderately fast velocities south of the BS that could indicate thinned crust to the south from extension. At 36 s we see a drastic swap of this velocity pattern, where we may be observing the juxtaposition of the intact subducting oceanic lithosphere to the north and the metasomatized mantle wedge to the south.

At a period of 47 s, we see velocities that are more uniform in this region. At increasing periods we observe a growing high velocity anomaly in the center of the region that may indicate the subducting lithosphere surrounded by hot, hydrated asthenosphere.

Previous geophysical studies of Antarctica often neglect this region as it is on the periphery of most Antarctic continental structures (e.g. Shen, Wiens, Anandakrishnan, et al., 2018). ANT-20 does cover this region, clearly observing a possible remnant slab under the northern AP at approximately 200 km depth (Lloyd et al., 2020). Lloyd et al. (2020) does appear to resolve slightly higher velocities in the South Shetland Islands region at approximately 75 km, which could correspond with the slivers of moderately high velocities we see appearing at 63 s.

5 Preliminary Attenuation Maps

The phase velocity maps provide crucial information for inferring the physical and chemical state of the upper mantle beneath Antarctica, but open questions remain. In the near future, robust attenuation values for Antarctica will greatly complement our phase velocity maps in determining the structures and processes at work underneath the Antarctic surface. We will be able to constrain temperature, composition, the possibility of melt presence, and viscosity in the Antarctic upper mantle by jointly interpreting shear velocity and attenuation (Havlin et al., 2021). These two data sets are complementary because their sensitivity to temperature, composition, melt, and volatiles are different (Havlin et al., 2021).

Here, we present preliminary Rayleigh wave attenuation maps for several regions in Antarctica: (A) the Gamburtsev Subglacial Mountains (GSM) in East Antarctica; (B) the northern Transantarctic Mountains and westernmost Ross Embayment; and (C) a region of West Antarctica that includes the Ellsworth-Whitmore Mountains (EWM), the eastern West Antarctic

Rift System, the Marie Byrd Land (MBL) dome, and easternmost Thurston Island. These results and their interpretations are preliminary.

5.1 East Antarctica: Gamburtsev Subglacial Mountains (A)

Similar to phase velocities in the GSM (Fig. 3-3), attenuation is relatively homogeneous in the GSM compared to other Antarctic regions (Fig. 3-10). Attenuation is low, as is expected for upper mantle that may represent thick, cool cratonic lithosphere (e.g. Bao et al., 2016). We see a slight increase in attenuation between 23 s and 36 s.

Interestingly, we see the same spatial patterns of attenuation variation across the GSM as for phase velocity, but in different period ranges. At 23 s, attenuation is slightly higher in the center of the GSM, similar to the pattern of lower phase velocities in the center of GSM phase velocity maps for periods up to 112 s. At periods of 36 s and greater, we see attenuation patterns across the GSM that indicate slightly lower attenuation towards the south. This spatial pattern is similar to the higher phase velocities we see at periods of 84 s and greater (Fig. 3-3).

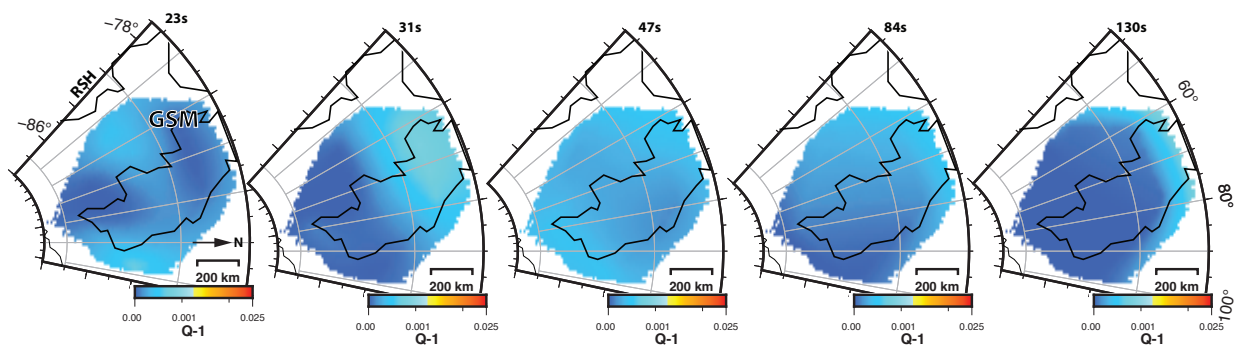


Figure 3-10. Rayleigh wave attenuation results for region A: the Gamburtsev Subglacial Mountains
Rayleigh wave attenuation results for region A: the Gamburtsev Subglacial Mountains (GSM) in East Antarctica at selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A.

5.2 East/West Boundary: Transantarctic Mountains and Ross Ice Shelf (B)

The anomalies in the attenuation maps we calculate in the Transantarctic Mountains and the westernmost Ross Embayment (Fig. 3-11) reflect interesting features also seen in the phase velocity maps of this region (Fig. 3-5). The clearest anomalies we see in attenuation at periods from 20 s to 41 s are Mt. Erebus and the Terror Rift (Fig. 3-11). At periods of 20 and 23s, the Mt. Erebus volcanic region is clear and exhibits relatively high attenuation compared to the surrounding region. From 27 s to 41 s, the Mt. Erebus and southern Terror Rift regions have slightly higher attenuation than the TAM or RIS. At 47 s, we resolve no anomalous attenuation from Mt. Erebus or the Terror Rift.

We observe predominately low attenuation in the northern TAM. There is no obvious indication in attenuation of large-scale lithospheric modification or foundering in the northernmost TAM, as proposed by Shen, Wiens, Anandakrishnan, et al. (2018). This corresponds with observations from phase velocity. We do see evidence of potential local modification by the adjacent Terror Rift in phase velocity, which also appears in attenuation maps at some periods such as 31 s.

At the highest periods we observe, there is a region-wide gradient of higher attenuation in the northernmost TAMs to lower attenuation southward, although we are less confident in results from these higher periods. In robust attenuation results, the observation of moderate attenuation in northernmost TAM could indicate elevated temperatures and would contribute to evidence of lithospheric foundering in the northernmost TAM (Shen, Wiens, Anandakrishnan, et al., 2018). Unfortunately, we are skeptical of our results, especially at higher periods, and are hesitant to draw conclusions from this observation.

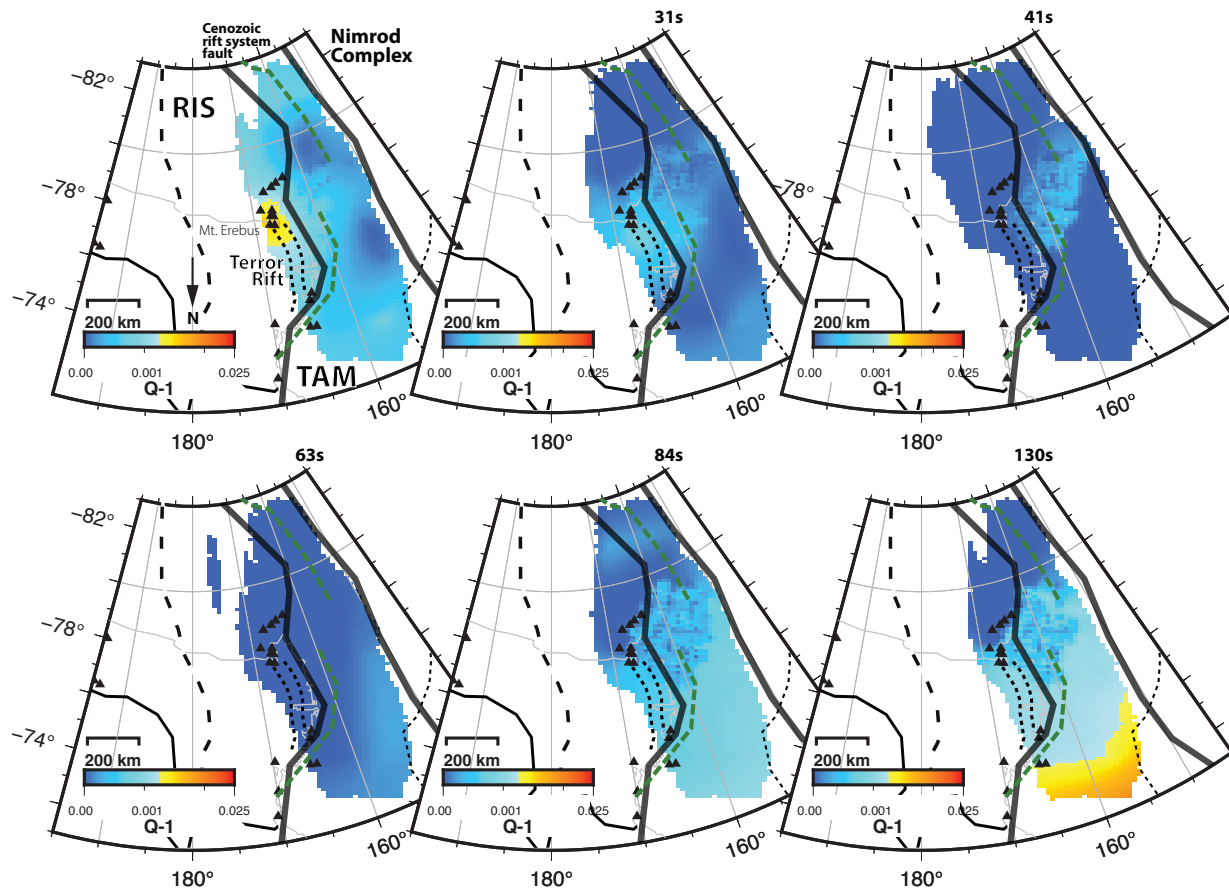


Figure 3-11. Rayleigh wave attenuation results for region B: the northern Transantarctic Mountains
Rayleigh wave attenuation results for region B: the northern Transantarctic Mountains (TAM) and adjacent Ross Embayment at selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A.

5.3 West Antarctic Rift System, Ellsworth-Whitmore Mountains, and Marie Byrd Land (C)

Our attenuation maps for the region of West Antarctica that includes the Marie Byrd Land (MBL) dome, eastern West Antarctic Rift System (WARS), Ellsworth-Whitmore Mountains (EWM), and western Thurston Island indicates substantial complexity (Fig. 3-12). We do see significant variability in attenuation values with period in this region, indicating that we should be skeptical of our results (Fig. 3-2). Despite being skeptical of our results in this region, we discuss them below.

Over the entire region (Fig. 3-12), we see the average attenuation increase greatly with period. At low periods (≤ 27 s), we see low-to-moderate attenuation in the EWM, WARS, and MBL with moderate-to-high attenuation in Thurston Island. From 31 s to 36 s, we see low attenuation in EWM and MBL with moderate attenuation in the WARS and Thurston Island. At periods of 41 s to 73 s, we observe moderate to low attenuation throughout the region, but there are unexpected attenuation patterns at 47 s and 55 s, where we see moderate-to-high attenuation beneath the EWM but low attenuation in the WARS. At 47 s the MBL-dome has attenuation that is slightly higher than the adjacent WARS. This elevated attenuation at the MBL-dome continues to 84 s. Above 84 s, we see very high attenuation values throughout the region that we believe are unreliable. As previously mentioned, this region has low resolution relative to the GSM and TAMs, thus we are particularly skeptical of these attenuation results.

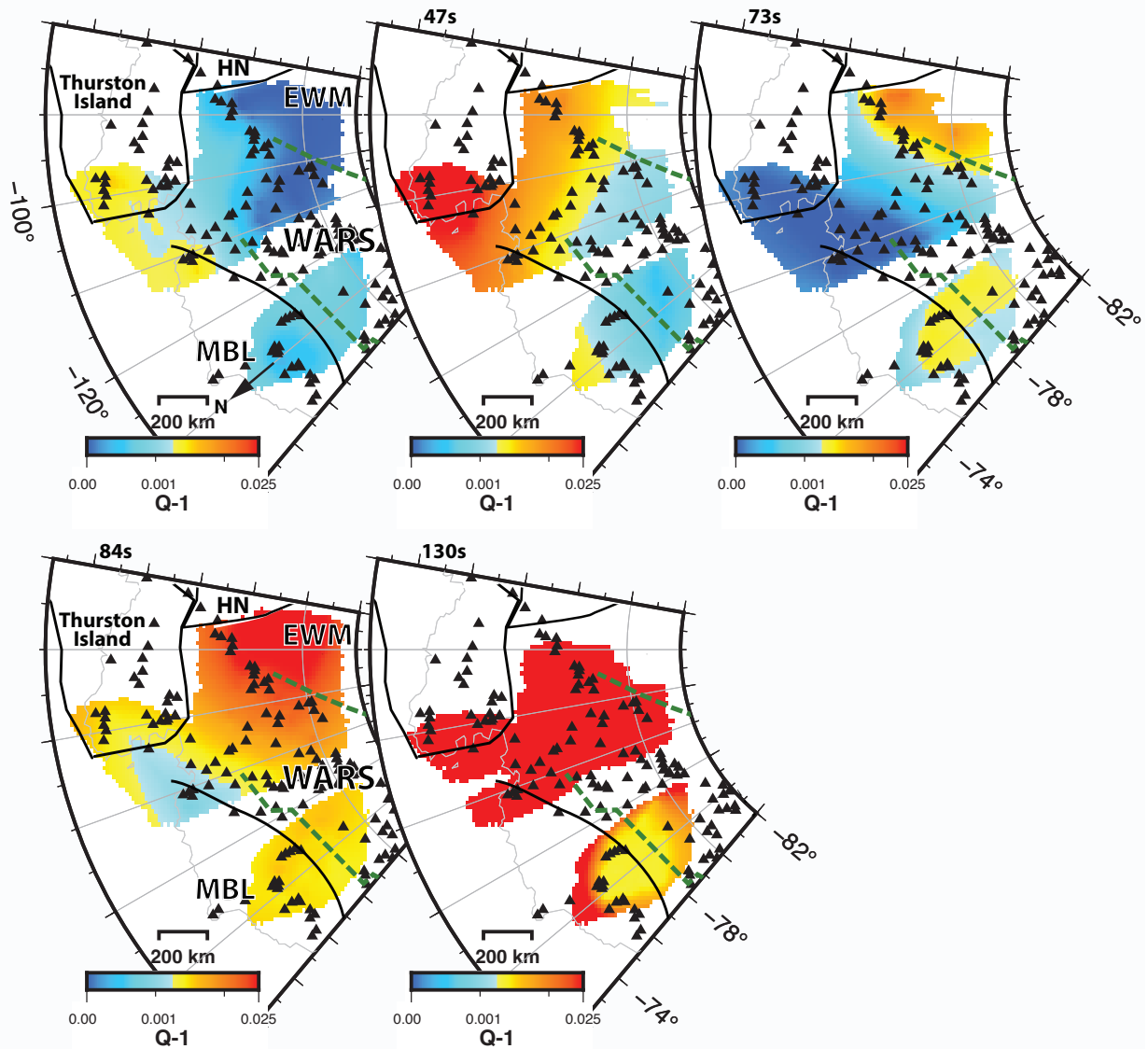


Figure 3-12. Rayleigh wave attenuation results for region C that includes the West Antarctica
 Rayleigh wave attenuation results for region C that includes the West Antarctic Rift System (WARS), Marie Byrd Land (MBL), Ellsworth-Whitmore Mountains (EWM), and Thurston Island. We provide only selected periods (indicated at top of individual maps). Maps are overlain by a simplified version of the map from Fig. 3-1A.

7 Summary & Conclusions

Our incomplete understanding of Antarctic structure and history is a result of limited geologic sampling and seismic instrumentation hindered by extensive ice-cover, inhospitable field conditions, and inaccessibility. Despite these limitations, it is clear that Antarctica exhibits remnants of a complex tectonic past that includes the breakup of Gondwana, Phanerozoic rifting

and subduction, and potential plume interactions. Recent continent-wide seismic models of the crust and upper mantle of Antarctica also reflect this complexity. We utilized the ASWMS toolset (Jin & Gaherty, 2015; Russell & Dalton, 2022) to investigate phase velocity and attenuation throughout Antarctica with regional-scale resolution. With this methodology, we were challenged to find regions with appropriate instrumentation for the estimation of travel time fields for phase velocity calculations. We also struggled to resolve attenuation with realistic values throughout our regions. The process of resolving robust attenuation values has not been completed, so in this thesis we present only preliminary attenuation maps for Antarctica.

We present results of Eikonal tomography Rayleigh phase velocities for four regions in Antarctica. In the GSM, we observe generally high phase velocities indicating a cool upper mantle lithosphere that is similar to the Wyoming craton in the US. We observe some heterogeneity in the upper mantle of the GSM that is consistent with the analysis of Shen, Wiens, Anandakrishnan, et al. (2018). In the northern TAM, we observe high phase velocities. Directly east of the TAM we observe slow but spatially variable seismic velocities that appear to be related to the ongoing Terror Rift and adjacent volcanism. In the Ross Sea Embayment, we observe thin, fast velocity lithospheric mantle underlain by heterogeneous phase velocities related to both historic rifting in the WARS and modern rifting in the Terror Rift. We also observe a similar fast uppermost mantle underlain by heterogeneous low phase velocities in the WARS of region C. Further east in West Antarctica, we observe phase velocities consistent with Shen, Wiens, Anandakrishnan, et al. (2018)'s proposed modification of thick lithosphere below the EWM as well as complicated variations in velocity beneath MBL. Finally, under the South Shetland Islands we observe clear indications of subduction structure.

We also present preliminary attenuation results for all of these regions except the South Shetland Islands. In general we find that the GSM in East Antarctica have low seismic attenuation at all periods, similar to attenuation expected in a cratonic region. The northern TAM exhibit attenuation that is lower than PREM at periods 20-84 s and higher than PREM at periods greater than 84 s. Notably, we observe elevated attenuation below Ross Island and the Terror Rift. The attenuation values we measure for the WARS are within one standard deviation of PREM up to periods of 84s, but exceed PREM at higher periods.

Ultimately, our Rayleigh surface wave results for phase velocity and attenuation reflect the complexity of Antarctica's geologic history. Further investigation of Antarctic mantle attenuation will provide valuable insight into mantle temperature and viscosity structure that will improve estimates of ice sheet melting and glacial isostatic rebound.

8 Acknowledgements

We thank Anant Hariharan for advice for interpretations. This work was funded by the U.S. National Science Foundation's Graduate Research Fellowship Program and National Science Foundation Award OPP-2201129. Generic Mapping Tools is used to plot figures in the manuscript and supplementary information (Wessel & Smith, 1991). Seismic waveform data were acquired from the Integrated Research Institutions in Seismology Data Management Center (IRIS DMC).

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Chapter 4 A comprehensive model for Inclusivity in the Geosciences: Reflecting on three years of AGU Fall Meeting panel sessions

Co-equal authors: Hannah E. Krueger¹, Sloane Garelick²

Contributing co-conveners: Kierra Wilk¹, Meredith Parish¹, Isabella Gama¹

Contributing panelists: Kuheli Dutt, Michael Hubenthal, Lorena Medina Luna, Darryl Reano, Marta Toran, Josh Bearman, Hendratta Ali, Allison Mattheis, Joseph Tumber- Davila, Kaatje van der Hoeven Kraft, Susan Meabh Kelly, Vashan Wright, Padi Boyd, Sarah Fortner, Cheryl Manning, Bryan Dewsbury, Aisha R. Morris

¹ Department of Earth, Environmental, and Planetary Sciences, Brown University

² Museum of Science, Boston, MA

Key Points:

- A Venn diagram model investigates the three primary foci of inclusivity work within a geoscience institution: local engagement, learning and teaching, and community atmosphere.
- At the intersection of these three primary approaches to inclusivity are: outreach initiatives, place-based and interdisciplinary education, and supporting diverse scientific identities
- In the center of the Venn diagram is relationships and accountability. Improving inclusivity is possible with authentic relationships that incorporate accountability at multiple levels

1 Introduction

In March 2020, we proposed a panel session at the American Geophysical Union (AGU) Fall Meeting titled “Inclusivity in the Geosciences: A conversation between scientists and educators.” We were inspired by the K-12 teachers we interacted with during outreach in local public schools, where we witnessed teachers include and empower diverse groups of students throughout their learning process.

A key aspect of geosciences’ diversity problem (Bernard & Cooperdock, 2018; Dutt, 2020b) is exclusion, both explicitly and implicitly. We realized that there are many experts working to create inclusive learning environments, but that these conversations and efforts to improve diversity, equity, and inclusion (DEI) in science education seemed to be relatively isolated and were seldom discussed amongst geoscientists in more general settings like AGU outside of the AGU education section. We wanted to create a venue for experts to share inclusive practices with the more general scientific audience through a panel discussion in hopes of improving the quality of geoscience classrooms, department atmosphere, and public engagement.

Months after submitting our session proposal, the tragic murder of George Floyd occurred and a large public outcry for social justice through the Black Lives Matter protests ensued. In this tumultuous time, the public awareness and motivation for improving inclusivity grew. This growth opened doors to be more explicit about the importance of improving the culture of inclusivity in the geosciences and made the goal of having a “general scientific audience” for our session that much more relevant, important, and feasible.

When determining panelists for our session, we strove to gather educators with various backgrounds working with diverse groups of geoscientists. We worked to bring together K-12 teachers, college-level educators, academic experts in geoscience education and DEI, outreach

coordinators, leaders in DEI efforts, and program directors, among others. These panelists, listed in Table 1, provided holistic perspectives of challenges to and potential actions for improving inclusivity in the geosciences in classrooms, outreach, research, and programmatic levels. The purpose of this paper is to consolidate conversations from our discussions amongst panelists during the formal AGU session as well as meetings between the conveners and individual panelists in order to share actionable approaches the broader geoscience community can take to work towards making the field of geoscience more inclusive, diverse, equitable and accessible.

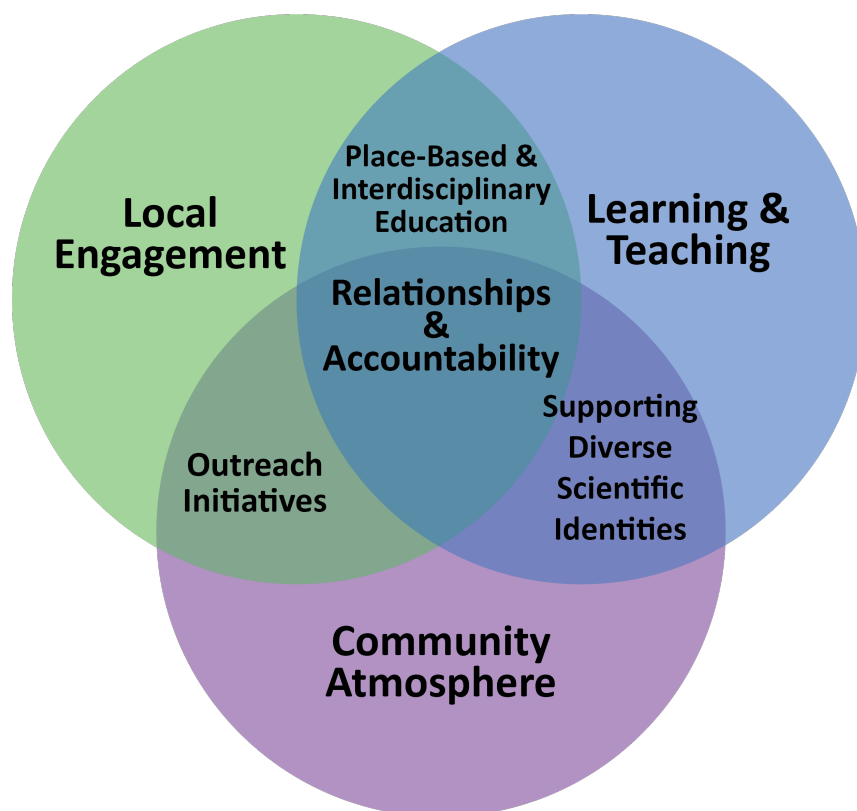


Figure 4-1. A Venn Diagram model for inclusivity in the geosciences

The key components to improving inclusivity within geoscience spaces can be represented by a simple Venn diagram that explores the intersections of local engagement, learning and teaching, and community atmosphere. This model is based on takeaways from conversations with and between panelists in the AGU session, Inclusivity in the Geosciences: A conversation between scientists and educators.

Table 1. Panelists and presentations for this panel session during the AGU Fall Meeting of 2020, 2021 and 2022. Gray text indicates panelists that were not able to attend the live session.

AGU Fall Meeting 2020	
Kuheli Dutt, PhD	Promoting Diversity, Equity, and Inclusion in the Geosciences
Michael Hubenthal, MS	Intentionality, effort, data and time are key ingredients for change: Fostering inclusive research opportunities in seismology and geophysics
Lorena Medina Luna, PhD	Creating an inclusive environment in the Geosciences through science education programs
Darryl Reano, PhD	Using Critical Reflexivity to Build Relational Accountability in the Geosciences
Marta Toran, MS	Maximizing the Potential of K12 Outreach as a Launchpad for Inclusivity In the Geosciences
Josh Bearman, MEd, MS	The Importance of Providing Common Experience and Personal Relevance in Engaging Students In Earth Science Concepts
AGU Fall Meeting 2021	
Hendratta Ali, PhD	Inclusion in Geoscience Education: Intentionally Braiding the Essential
Allison Mattheis, PhD	Educating for Intersectional Justice and Equity in the Geosciences
Joseph Tumber-Davila, PhD	Approaches to make the Geosciences more Inclusive: Highlighting Student-Led initiatives
Kaatje van der Hoeven Kraft, PhD	Developing a Geoscience Interest and Identity for a More Diverse Array of Students in Introductory Courses
Susan Meabh Kelly	Research- and Practice- Informed Pathways towards an Inclusive Geoscience Ecosystem
Vashan Wright, PhD	We are scientists
AGU Fall Meeting 2022	
Padi Boyd, PhD	NASA's Science Mission Directorate Bridge Program—A New Initiative to Boost Diversity, Equity, Inclusion and Accessibility in STEM
Sarah Fortner, PhD	The Need to Advance Geoscience Collaboration for Environmental Justice
Cheryl Manning, MS	Doing my Homework: How an Inclusive Geoscience Classroom and Community Begins with Me
Bryan Dewsbury, PhD	Beyond Inclusion - The Inclusive Classroom as Preparation for a Socially Just Society
Aisha R. Morris, PhD	Developing an Inclusive and Equitable Geoscience Community: Perspectives from a Black Geoscientist's Personal and Professional Journey

2 A Venn Diagram Model for Inclusivity in the Geosciences

In this commentary, we reflect on the lessons we learned about how to grow as an inclusive geoscience community from our session, *Inclusivity in the Geosciences: A conversation between scientists and educators*. Our reflection is based on conversations between panelists during the AGU panel session as well as individual conversations between the conveners and panelists in preparation for the session. In these conversations, common themes emerged. We use a Venn diagram to organize these themes and present key takeaways (Fig. 4-1). The three main themes are (1) local engagement, (2) learning and teaching, (3) community atmosphere. We identify the overlaps of these main themes as (1,2) place-based and interdisciplinary education, (1,3) outreach initiatives, (2,3) supporting diverse scientific identities. All of these aspects of an inclusive science community require authentic relationships and accountability, which exists at the intersection of all of these themes in the Venn diagram. We hope that this organizational scheme will provide a clear, holistic, and actionable approach for the geoscience community to promote inclusivity.

2.1. Three Main Themes: Local Engagement, Learning & Teaching, and Community Atmosphere

2.1.1 *Local Engagement*

Community interactions that prioritize community needs and build long-term relationships increase science accessibility while addressing local environmental injustice and inequity.

The first step in an authentic community partnership is determining community needs. Oftentimes, scientists assume or ignore community needs when implementing outreach activities or other broader impacts (Boyd, 2022; Fortner, 2022; Tumber-Davila, 2021). Open communication and development of a clear plan in collaboration with the community are

necessary to address community needs (Fortner, 2022). Community need-based research directions, often referred to as undone science, can illuminate inequities within the community that are crucial for scientists to address in their work (Fortner, 2022; Frickel et al., 2010). For example, a starting point for identifying socially relevant research areas in your local community is to consider the impacts of historic redlining & systematic racism (Fortner, 2022; Fortner et al., 2022), using resources such as the opportunity atlas (Chetty et al., 2018), or the EPA's social justice map. Once these areas are identified, inquiry can be made to community partners from these areas about the research that can best serve them.

It is also important to identify ways to make science more accessible to the community. For example, investigating the language diversity of the local community can indicate needs for multilingual materials and communication (Medina Luna, 2020; Medina Luna et al., 2019; Toran, 2020a). Providing bilingual materials in outreach and community interactions is a key way to significantly increase accessibility to science in community partnerships (Medina Luna, 2020; Toran, 2020a). Furthermore, communities may also have generational barriers to language. Often students will be confident English speakers, but will have parents that need communication in an additional language (Medina Luna, 2020).

In addition to open communication, a commitment of time is key for a respectful and sustained community partnership. Scientists should not propose a project that does not commit the time and resources required to bring benefits to the community, rather than just the academy (Mattheis, 2021). The time commitment required to grow relationships and do science that truly benefits the community in the long-term often conflicts with the limited time windows of grant funding (Hubenthal, 2020b; Toran, 2020b). Therefore, before beginning community interactions, scientists should construct strategies for continuing their work after grant funding ends.

In summary, partnerships between communities and scientists should be symbiotic and build trust. These collaborations should provide social relevance and broader impacts to science while benefiting the partnering community with accessible science that addresses inequities.

2.1.2 Learning and Teaching

Pedagogy that centers student identity, inquiry, and engagement will inherently create a more inclusive learning environment. These teaching approaches are commonly used in K-12 and community college classrooms, but less frequently applied to teaching in higher education settings (Mattheis, 2021; van der Hoeven Kraft, 2021). It is important to train educators at all levels of geoscience education in inclusive and evidence-based pedagogical approaches (Mattheis, 2021; Tumber-Davila, 2021).

2.1.2a Student Identity

Traditionally, classroom environments reward students that perform well on assessments, provide correct answers during lectures, and are adept at associating classroom material with their previous experiences. These are all traits rooted in how *prepared* or *ready* a student is prior to taking a course. National racial and socioeconomic inequities in the outdoors, travel, and K-12 education quality can be directly reflected in student *readiness* for geoscience material (Dewsbury, 2022; Morris, 2022b). To create a more inclusive classroom it is essential to create a space that values a students' *passion* and *potential* aside from their preparedness (Boyd, 2022; Morris, 2022a).

Evaluating passion and potential requires a commitment to valuing the individual identities of students. Taking the time to get to know students and their cultural background is essential for teaching well and inclusively (Dewsbury, 2022; Wright, 2021). This pedagogical approach is commonly referred to as culturally responsive teaching. With this knowledge,

instructors can orient teaching and learning towards students' interests and identities. (Kelly, 2021). Geoscience is unique in that it can be equally field intensive, laboratory focused, and/or technology heavy (Ali, 2021), and involves interdisciplinary topics from chemistry, physics, biology, and more. By learning about students' interests, educators can more directly draw connections to the interdisciplinary topics being taught, therefore making the learning more personal and relevant to student lives (Ali, 2021). There is active geoscience education research that works to create resources for culturally responsive teaching. For example, the GeoConnections project focused on creating culturally relevant modules for Indigenous students (Reano, 2020b). Furthermore, value placed on student identity allows for social growth within the classroom as well, allowing for both the intellectual and social growth of students (Dewsbury, 2022).

2.1.2b Student Inquiry

Inquiry-based learning is a student-centered teaching approach that incorporates active participation of students by encouraging them to ask questions and relate concepts to their lived experiences (e.g. Apedoe et al., 2006). One of the obstacles to implementing inquiry-based learning in university classrooms is the lecture-lab structure that separates content from practice (Manning, 2022b). However, incorporating inquiry-based learning into the classroom is a valuable approach for inclusive geoscience education as it allows for more engaging, student-driven learning.

Using Bybee's 5Es lesson structure is one approach to inquiry-based learning that's often used in K-12 classrooms. The five steps of Bybee's 5E's are Engage, Explore, Explain, Elaborate, Evaluate (Bybee et al., 2006). Students initially engage with a question, phenomenon or challenge that is thought-provoking and encourages further questioning. Students then explore

the question through investigations and activities to challenge their prior knowledge and encourage new ideas. After exploration, students draw conclusions and explain the phenomenon with their newly gained knowledge. Students then elaborate upon their learning and apply their knowledge to a new situation. Finally, students reflect on their learning process through an assessment. The 5E model can be used to emphasize societal significance, sense of place, and personal connection in Earth science material (Bearman, 2020b).

An additional approach to inquiry-based learning is creating a common experience for students (Bearman, 2020a). Creating a common experience allows all students to begin with the same information, regardless of their lived experiences and educational backgrounds, therefore giving all students equal opportunity to learn and succeed. For example, beginning a class by going on a field trip, watching a video, discussing an interesting phenomenon, etc., will create a common experience for students to then explore and learn from. This approach prioritizes student passion over readiness, by including all students in the learning process regardless of their past experiences or knowledge. This is contrary to the traditional lecture-lab structure where students listen to a lecture about material, then participate in a lab exercise that reiterates lecture material. To effectively implement inquiry-based learning at the university level, the traditional lecture-lab structure must be reconsidered.

2.1.2c Student Engagement

Engagement in the classroom is a crucial component of encouraging student passion, making learning accessible and inclusive, and ultimately diversifying the geosciences. Oftentimes students with prior knowledge of a subject will appear more engaged regardless of the effort put forth by the instructor to engage students. Without intentional engagement of all students in the classroom, the excitement and motivation students have towards a subject may just reflect their

preparedness and *readiness* rather than their *potential* and *passion*. Engagement can be particularly challenging for students from resource-poor, lower socio-economic status households who are underrepresented in STEM. This disconnect is often due to a lack of context or relevance provided to the students (Bearman, 2020b). Student identity and inquiry (and place-based learning, discussed in Section 2.2.1) are clear approaches to promote engagement. Providing students with common experiences and relevant context related to geoscience material promotes engagement for students who may lack experiences with the natural world (Bearman, 2020b). Utilizing these methods will inherently create a more engaged and inclusive environment. Once students are initially engaged, further engagement will be internally driven by the student (Michaels et al., 2007; van der Hoeven Kraft, 2021).

2.1.3 Community Atmosphere

Intentional and continuous work to equitably support people of all identities and backgrounds is crucial for an inclusive geoscience community.

Geoscience is one of the least diverse STEM fields. Low diversity often leads to the prevalence of stereotypes and implicit bias within a field, often making the geosciences a hostile environment for marginalized groups (Dutt, 2020a). The culture within STEM, particularly geoscience, can be considered a “hostile obstacle course” for historically marginalized and underrepresented people in geosciences (Berhe et al., 2022). Simply increasing diversity and visibility of historically underrepresented scientists within geoscience programs and leadership does not remove these obstacles, but retains a hostile environment that an increasing number of scientists are being asked to navigate (Berhe et al., 2022; Hubenthal, 2020a). In order to create an inclusive geoscience culture that is equitable, supportive, and safe, intentional steps must be taken at multiple levels within geoscience organizations to remove barriers to the success of

BIPOC, women, and members of the LGBTQIA+ community (Berhe et al., 2022). We must account for both the history of exclusion and consider the many paths people engage with the geosciences when addressing issues of diversity (Ali, 2021). When students have positive and inclusive experiences within geoscience spaces, this information will permeate and encourage other students to pursue geoscience higher education (Hubenthal, 2020b).

When working towards these changes, it is important to first identify goals in improving diversity and inclusion. An important question to ask before implementing change is “How do we define success in diversifying geosciences?”. Answering this question can lead to clear metrics to track and evaluate the impacts of changes (Hubenthal, 2020a). For example, the IRIS undergraduate internship program redefined the success of their program as supporting historically underrepresented students in geoscience. They tracked the number of students that pursue geoscience careers upon completing the internship in order to evaluate program success (Hubenthal, 2020a). Key datasets the program used to evaluate their success were applications, intern/mentor perceptions, and longitudinal tracking of their participants using a yearly census of program alumni (Hubenthal & Taber, 2020). One change the IRIS internship program made to increase diversity within their program was to change their application prompt to “What should we know about you?” instead placing as much value on awards and grades. This allowed students to use their own cultural perspective to articulate their value, provide more information on the value system of the student than a traditional application, and explain their transcripts in the context of their other responsibilities. This approach provides a space for applicants to demonstrate their ability to persevere regardless of their actual grades. Additionally, one metric of program success the NASA Bridge program uses is positive student experience, which places emphasis on the quality of training received by student mentors. It is crucial that any new goals

are clear and measurable (Boyd, 2022; Hubenthal, 2020a). Furthermore, narratives of engagement can be used to determine the direction of changes necessary within a program and aid with general feelings of inclusion (Reano, 2020a). Narratives of engagement are open and holistic reflections on past experiences that go beyond the narrow “cheery narratives” of research and education that are often shared (Reano, 2020a). For example, instead of simply celebrating a paper’s publication, sharing a full story of initial rejection from another journal or the significant revisions that were required before its acceptance. Sharing these narratives allows scientists to understand the complexities of the scientific process and not feel inadequate when they experience similar processes (Reano, 2020a). Open reflection on complicated and/or negative experiences within programs can illuminate the hostile obstacles that exist and can provide direction for necessary changes to improve inclusivity within geoscience programs.

Additionally, geoscience programs should consider changing their recruitment tactics in order to improve diversity within their departments (Boyd, 2022; Hubenthal, 2020b). Undervalued spaces to recruit Earth science students are community colleges and primarily undergraduate institutions (PUIs). Furthermore, Earth science programs often struggle to recruit at minority-serving institutions including historically Black colleges and universities (HBCUs) and Tribal Colleges and Universities (TCUs) (Boyd, 2022; Hubenthal, 2020a). Many HBCUs and TCUs do not have geoscience programs, consequently targeting internship or PhD program advertising towards primarily earth science undergraduate students (i.e. through listservs) neglects many diverse sets of students. In universities without geoscience programs, connections in physics & chemistry departments and relationships with faculty are important (Hubenthal, 2020a; Page, 2017).

In diversity and inclusion work, programs often rely on people from marginalized groups to do a majority of the labor. Instead, programs and individuals should seek out resources (e.g. experts, publications) and solicit volunteers for inclusion work (Morris, 2022a). We can also prepare the future generation of scientists and educators (e.g., students, post-docs, trainees) for this work through support of student-led DEI efforts, training, and courses (Tumber-Davila, 2021).

2.2. Overlapping Themes: Place-Based Learning & Interdisciplinary Collaboration, Outreach Programs, and Sense of Community

2.2.1 Place-Based & Interdisciplinary Education

A more engaging and inclusive classroom can be fostered by exploring topics through the local environment, perspectives from the local community, and collaborations with non-geoscience or nonSTEM academic fields.

In place-based learning, students have the opportunity to develop a personal connection with their community and environment, while also learning about important local issues such as climate change impacts, environmental injustice and natural hazards. However, there is a lack of place-based learning in higher education classrooms: only 30% of faculty teach about local environmental problems, 25% teach about issues of environmental justice, and only 8% incorporate community-based projects into their classes (Fortner, 2022). Incorporating place-based topics into the classroom can foster a more engaging and inclusive classroom by incorporating topics relevant to the students' lives, creating a common experience for all students, and encouraging discussions about important environmental issues.

For example, Josh Bearman created an activity to teach the concept of albedo in which middle school students explored differences in surface temperature throughout Richmond, VA

and compared this to tree cover relative to demographics and income within their city, to assess instances of climate injustice and the need for mitigation efforts (Bearman, 2020a). By exploring issues directly affecting the communities in which they live, students were able to create a more personal connection to their learning. Place-based learning can be used to further drive inquiry and engagement by allowing students to propose investigations of their own community concerns. For example, when Cheryl Manning was a middle school teacher in Utah, her students brought up the issue of increased dust in the air due to a lack of vegetation on nearby salt flats, which was disproportionately impacting low-income communities. Cheryl and her students wrote and received a grant to re-establish the population of willow trees in the area to reduce the dust in these communities. Allowing students to propose community issues to discuss and learn about is a powerful way to engage students and relate their learning directly to their lives (e.g. DeFelice et al., 2014; Gruenewald & Smith, 2014; Manning, 2022b).

Incorporating perspectives from community members can further foster engagement and inclusivity within the classroom (**Team: et al., 2022**). Consulting with experts from the local community is an excellent way to create a sense of place in the classroom. Students connect with local issues that may be relevant to their lives, while emphasizing areas of the community where investment and attention is needed (Fortner, 2022). Such interactions allow students to directly learn about community issues and work together to identify the intersection between the classroom and the community. One example of a program that fosters classroom and community engagement is the Redlining Education and Change Coalition in Springfield, Ohio, which aims to raise awareness of the structures and practices, such as redlining, that have resulted in racial inequities and injustices in Springfield, Ohio. Key components of their approach is to partner with community members to address inequities, engage with local students, and incorporate

racial injustices into curricula (Fortner et al., 2020). Community collaborations such as this can create open conversations to learn about and discuss community issues that allow for diverse perspectives, thus allowing for a more complete understanding of the problem and greater potential for developing solutions.

Another valuable perspective to have in the classroom are academics from outside of the geosciences, who can provide important insight for effective community engagement and place-based learning. For example, coordinating with experts in communication can help to establish relationships with community leaders (Fortner, 2022) and collaborating with education experts can help to create authentic and effective place-based learning experiences (Reano, 2020a). To establish these collaborations, geoscience departments can work directly with a geoscience education researcher, even adding faculty with this skill set to their department (Reano, 2020a), and professors can invite guest speakers from other departments. Larger-scale, discipline-wide efforts to foster collaboration outside of the geosciences can also aid effective community engagement and place-based learning experiences. For example, the Science plus Education Collaboratory (collaboratory, 2021) is an interdisciplinary partnership of scientists and educators that aims to leverage expertise beyond institutional and disciplinary boundaries to accomplish a broader impact and address overlapping national initiatives and issues (Kelly & Thompson, 2018). Another program that aims to foster interdisciplinary collaborations in geoscience education is InTeGrate (Fortner, 2022). InTeGrate is a collaboration between faculty in the sciences and other disciplines, educational specialists, and evaluation experts to develop materials for and promote interdisciplinary earth science education (Fortner et al., 2016).

2.2.2 Outreach Initiatives

Creating and valuing effective outreach programs improves accessibility to geoscience education, cultivates a more inclusive academic community and demonstrates a commitment to accessible geoscience education.

Outreach initiatives allow geoscience programs to interact and engage with broader communities and can provide pathways into the geosciences through mentorship and training for the next generation of scientists (Medina Luna, 2020; Toran, 2020a; Tumber-Davila, 2021). Geoscience is a globally relevant discipline but often is not introduced in pre-college education (Ali, 2021). This highlights the need to bring geoscience into the K-12 classrooms and broader public spaces. For example, the *Pertenecer/YouBelong* program is an outreach program at Stanford University introduces K-12 students to academic and environmental science careers, exposes students to college, and demonstrates what it means to be an earth scientist (Tumber-Davila, 2021). Another example is *Con-Ciencia En Las Américas*, which is a Spanish-language seminar series that aims to highlight Latin American Scholars, create a broader community of scholars across the Americas, and provide Spanish language science content for a broad audience (Tumber-Davila, 2021). Additionally, the PreK-12 Education and Outreach program in the Department of Geological and Environmental Sciences at Appalachian State (GES at AppState) has many initiatives that seek to expand access to geoscience education in and beyond the local community. The outreach opportunities offered by this program include classroom kits and professional development workshops for teachers, K-12 visits to the on-campus geology teaching museum, and outreach to rural communities through a mobile classroom (Toran, 2020a). Such initiatives expose students to earth science topics and make learning about these topics more accessible.

Outreach initiatives also have the potential to cultivate a more inclusive academic community within a department. In general, science outreach provides students and faculty with opportunities to develop science communication skills (Toran, 2020a). For example, engaging in outreach with K-12 students can expose higher education faculty and students to new ways of teaching (Kelly, 2021; van der Hoeven Kraft, 2021). When working with younger and more diverse groups of students, the need for inclusive teaching strategies, such as the 5 E's and inquiry-driven learning, are more apparent. Additionally, K-12 teachers are experts in these strategies, so collaborations between academic faculty and K-12 instructors can provide mentorship and feedback in these teaching methods. Engaging with students outside of academia and collaborating with K-12 teachers allows scientists to develop their individual inclusive pedagogies, which can be applied within higher education to create more inclusive learning experiences and improve the culture of inclusion in academic programs (Kelly, 2021).

Finally, treating outreach as a priority in a geoscience program and developing genuine relationships with the local community through outreach improves the atmosphere of inclusivity in a program and demonstrates a commitment to diversity, equity and inclusion (Toran, 2020a). Creating funding- and reward-structures to support outreach efforts aimed at making the geosciences more inclusive indicates the importance of outreach in a department. For example, in the GES at AppState, undergraduate students can receive paid outreach internships for a semester of outreach work (Toran, 2020a). Programs can also allocate significant funds toward outreach programs and encourage participation in outreach efforts (Toran, 2020a), particularly those led by early-career students (Tumber-Davila, 2021). Finally, considering outreach and DEI work in the tenure process and placing value on written products of outreach work, such as

publications or chapters of student dissertations, also demonstrates a department's commitment to supporting these programs (Dewsbury, 2022; Morris, 2022a; Wright, 2021).

2.2.3 Supporting Diverse Scientific Identities

The development of science identity continues outside of the classroom, so it is important that intentional inclusion continues at a program level. Applying strategies of inclusive pedagogy at a program level can create a culture that encourages change and growth that reflects the diversity of the scientific community. Two approaches to support a sense of belonging for diverse scientific identities are culturally responsive teaching and affective learning.

At a program level, scientific identity is not just defined by scientific interests but also by cultural background and lived experiences can help to foster a sense of belonging. One example of a program that works to prioritize cultural identity in the geosciences is GeoLatinas.

GeoLatinas is a cohort of Latina geoscientists that aims to empower the work and research of Latina geoscientists and close the cultural and linguistic gap that exists between Latin Americans and opportunities in geosciences (Medina Luna, 2020). Another approach to place value on a holistic scientific identity is to develop community guidelines that consider the intersection of multiple identities (e.g., scientific, cultural, social) in STEM spaces (Manning, 2022b; Mattheis, 2021). Reflecting on previous experiences and incorporating narratives of engagement is a valuable way to inform these guidelines and incorporate cultural and social identities (Manning, 2022b; Reano, 2020a).

Affective learning is another pedagogical approach that can extend beyond the classroom and be applied at the program level to further foster a sense of belonging. The affective domain of learning describes the manner in which we deal with things emotionally, such as feelings, values and attitudes (Krathwohl et al., 1973). Openly prioritizing the feelings and emotional

growth of scientists and students, both in the classroom and at the program level is crucial (Dewsbury, 2022). For example, emphasizing the importance of and providing resources for mental health and well-being can demonstrate the value of feelings and emotions in scientific identity. Additionally, developing specific and intentional programs within the department that address inclusion can provide a community space to work towards DEI goals and share feelings, values and attitudes related to the issues being addressed. For example, Unlearning Racism in Geoscience (URGE) provides a space within programs to draw on existing literature, expert opinions, and personal experiences to develop anti-racist policies and resources (Wright, 2021). Implementing actions at the program level that are culturally-responsive and prioritize the affective domain can foster identity growth and create a sense of belonging. The overall respect shown towards science community members when they are provided with an intentional sense of belonging gives rise to better morale overall and a more inclusive space (Ali, 2021).

2.3 Central Theme: Relationships and Accountability

The common thread amongst these many approaches to creating a more inclusive geoscience community is the importance of authentic relationships, which require consistent accountability ranging from the individual to program level.

Geoscience education traditionally relies on educator knowledge and resource availability, but it's also essential to consider the relationships between educators, learners and the communities served (Ali, 2021). Within these relationships, it's important to acknowledge the interests of everyone involved and address any biases because relationships often relate to resources provided to individuals and programs such as presentation or research opportunities, funding, and curriculum (Ali, 2021). Authentic relationships within scientific spaces exist in many forms, many of which we have indirectly discussed thus far in this commentary. In this

section, we explicitly address the following relationships, although we understand this list is not exhaustive: Local community and scientific programs/departments; student and teacher; mentor and mentee; relationships with yourself. In order to develop and sustain these authentic relationships, there are important values and approaches to consider. These values consist of approaching relationships with an asset-based mindset and developing trust through accountability.

An asset-based mindset in relationships allows for full participation, creative exploration and inclusive learning. An asset-based mindset in the classroom can create a space where students know that their “experiences, knowledge, contributions, abilities, and capacity to learn are respected and valued” as opposed to simply valuing the preparedness each student has for the material (Manning, 2022a). Demonstrating an asset-based mindset can start as early as student applications or faculty interviews (Hubenthal, 2020a; Morris, 2022a). Oftentimes, programs consider how applicants “fit” into their current culture, which can indicate exclusivity and resistance to change within a program’s culture (Morris, 2022a). Instead, programs can work more intentionally to expand their preexisting values and accept applicants with unique experiences (Hubenthal, 2020a; Morris, 2022a).

Fostering authentic relationships requires trust built through consistent and intentional accountability. As a scientific community, we must (1) hold ourselves accountable, (2) hold others accountable, and (3) make changes intentionally.

(1) We can hold ourselves accountable by assessing personal biases and traumas so that we do not perpetuate these within our fields, and by developing a practice of self-reflection (Manning, 2022b). Assessing how the current culture has affected our ways of acting, thinking, and speaking allows us to embrace different perspectives and cultural change (Manning, 2022b).

We must acknowledge the privilege, bias, and trauma experienced and perpetuated within the geoscience community (Manning, 2022b). It is common to perpetuate trauma that you have dealt with, so intentional self-reflection is necessary to stop this negative cycle (Manning, 2022b). Furthermore, maintaining emotional health and taking breaks for mental health are necessary to maintain healthy and productive relationships with others. You must be self-aware of when you need rest and have systems in place to cancel obligations and/or hold classes or meetings remotely (Manning, 2022b). “Acting out,” lacking patience with others, or feeling defensive can be signs that rest is needed for mental health. When this happens, it is crucial to take this rest, but also to take responsibility for this “acting out” with self-reflection (Manning, 2022b). These steps are particularly important for educators, as it is an incredibly emotionally demanding job that has a direct impact on students (Manning, 2022c). Cheryl Manning provides some starting points for reflecting on and assessing bias, personal trauma, and emotional health: regularly completing the Harvard implicit bias test (e.g. Ratliff & Smith, 2021); journaling, especially to identify when you feel defensive; and practicing vulnerability during DEI trainings (Manning, 2022b).

(2) We also must encourage honest feedback and be willing to accept feedback from others. Building trusting relationships is a learning process that requires continual growth. A willingness to listen and holistic reflection on experiences of others can overcome apparent barriers to relationships due to differences (i.e. age, race, gender) (Morris, 2022a; Reano, 2020a). In this process, it is also important that relationships are reciprocal (Kelly, 2021). The communication and reflection in relationships should not primarily be put on those who are historically excluded within science and/or those who have less power in the relationship dynamic (e.g. student-faculty) (Morris, 2022a). It is possible for programs to create clear

expectations for relationships between students, faculty, and the community and provide clear avenues for constructive feedback (Boyd, 2022).

(3) Finally, we must be intentional in accountability by tracking change, evaluating progress, and making improvements within scientific programs (Hubenthal, 2020a). Changes within a program must be intentionally made and evaluated based on clear expectations and continual reflection (Boyd, 2022; Hubenthal, 2020a; Manning, 2022b; Reano, 2020a). Collaboration with social scientists can improve how we implement changes based on feedback (Morris, 2022a). While the prospect of change can be intimidating, even small consistent steps that clearly align with thoughtful goals can build broader change (Hubenthal, 2020a).

3 Summary

In summary, a Venn diagram model organizes key components of improving inclusivity within the geoscience community. Local engagement, learning and teaching, and community atmosphere can often appear as distinct approaches to improving inclusiveness in Earth science, but within their connections emerge holistic approaches towards a more inclusive geoscience field. Broadening science identities to value culture, connections with the local community, life experiences, emotional health, and capacity to learn can increase engagement and sense of belonging, while promoting positive and authentic relationships between peers, mentors, and the community.

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Appendix A Supplementary Materials: Global Patterns in Cratonic Mid-Lithospheric Discontinuities from Sp Receiver Functions

Introduction

This supporting information provides additional detail regarding data processing, further evaluation of specific geographic regions, alternate analyses of data, and the individual receiver function results for each backazimuthal swath we present (Fig. SA16-100). Generic Mapping Tools is used to plot figures in the supplementary information (Wessel & Smith, 1991).

Text S1. Supplementary Methodology

S1.1 Data qualifications & initial processing

We consider events with magnitudes between 5.5 and 9, occurring at depths above 100 km, and within an epicentral distance range of 65° to 80° . In particular, this epicentral distance range ensures that primary S-wave ray paths do not reach the outer core (upper bound) and travel at a relatively steep angle through the upper mantle (lower bound) in order to sample local structure and prevent critical refraction of S-to-P converted waves (Farra & Vinnik, 2000). We also remove seismogram data in time ranges where converted P waves have likely refracted at the critical angle and are no longer traveling upward (e.g. Ford et al., 2010).

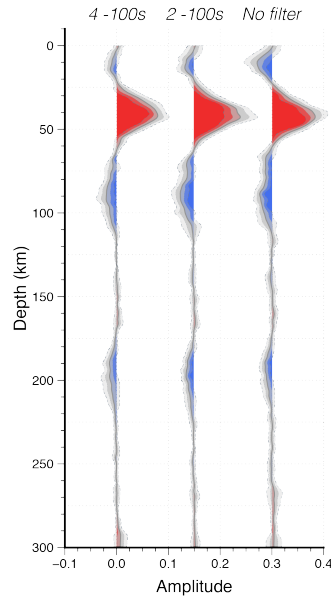


Figure S A-1. Filter comparison

Real data Sp RFs calculated for RSSD processed with different filters applied to the seismograms prior to deconvolution and stacking. The traces and weights used in the stack are those that are represented in Fig. 1-5. The RFs traces are filtered from (left) 4-100s, (middle) 2-100s, (right) no filter. Sp phases are similar regardless of filter used, but the shape of the MLD phase when no filter is used (right) appears to be more influenced by a Moho side-lobe. To include the most information possible (least filtered) while avoiding processing artifacts and noise, we choose the 2-100s filter (middle). The similar widths of the phases for the different filters are in part due to use of the same Gaussian shaping filter in the deconvolution operation. The Sp RF polarity has been flipped to match the Ps convention.

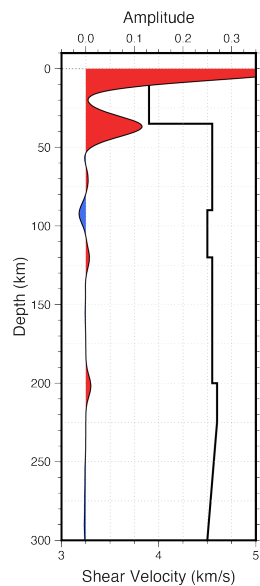


Figure S A-2. Rotation comparison

Synthetic Sp RF calculated for same structure as Fig. 1-2 processed with the same methodology except for data is processed with the vertical-radial components of the seismograms, instead of P-SV components. Note the large energy at 0 km depth in the unrotated RF as well as instability below the Moho and failure to resolve the deep, gradual NVG. The Sp RF polarity has been flipped to match the Ps convention.

In addition to the methodology description in the manuscript, such as filtering from 2-100s (Fig. SA-1), the following steps were taken to process data. We detrend and rotate seismograms from vertical-north-east components to vertical-radial-transverse components. Spikes in amplitude are determined to exist where an amplitude gradient is greater than thirty times the mean of the root-mean-squared gradients for the entire seismogram. Seismograms with unrealistically large spikes in amplitude are removed.

Vertical-radial-transverse component seismograms are converted to P - SV components with a Free Surface Transform (FST, Equation 1, Fig. SA-2; Kennett, 1991). A FST requires seismic velocities directly below the seismometer to be known. For both iterations in our data culling procedure, we search for free surface velocities that best convert the entire waveform with 7 s around the parent arrival (Fig. SA-2; Hua et al., 2020). This is performed on all Ps and Sp seismograms that remain at that point in the data culling procedure.

$$\begin{bmatrix} P \\ SV \end{bmatrix} = \begin{bmatrix} \frac{\beta^2 p^2 - \frac{1}{2}}{\alpha q_\alpha} & \frac{p \beta^2}{\alpha} \\ p \beta & \frac{\frac{1}{2} - \beta^2 p^2}{\beta q_\beta} \end{bmatrix} \begin{bmatrix} Z \\ R \end{bmatrix} \quad \text{Equation 1}$$

The signal-to-noise ratio (SNR) is determined by using parent (S-wave) data component that has been filtered from 2-100s. To calculate the SNR, the Hilbert transform is taken across the seismogram. The absolute value of this transform is averaged in noise (26 s) and signal (8 s) windows from -30s to 30s around the expected parent S-wave arrival on the seismogram. When the ratio of this average for a signal window and its preceding noise window is the greatest, this value is taken as the SNR of the trace. To determine the apparent parent phase onset, we perform the same SNR methodology as previously explained using data filtered at 0.5 - 100s. If the

apparent phase onset for a seismogram is within 15s of the expected arrival time calculated using the TauP toolkit (Crotwell et al., 1999) with the iasp91 model (Kennett & Engdahl, 1991), we keep this seismogram in our processing. Seismograms with apparent phase onset >15s of the expected arrival and SNR<1.5 removed from processing (Fig. SA-3).

Seismograms are migrated in time to the average ray parameter of all events using the SEMum2 model at the location of the station (French et al., 2013).

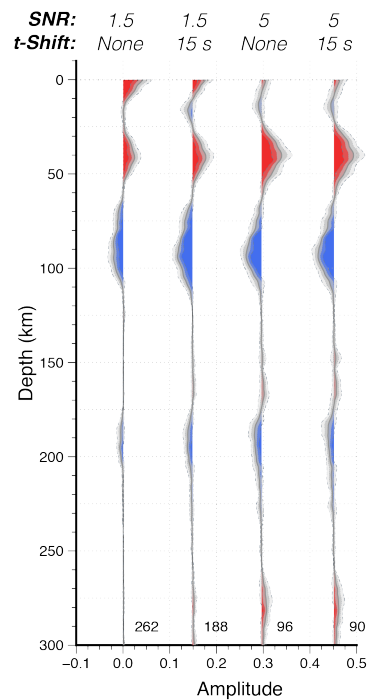


Figure S A-3. SNR criteria

Observed Sp RFs calculated for station RSSD processed with different signal-to-noise criteria for the traces in each stack. The stacks are processed with the choices indicated in the main text, but without k-means selection, so all traces that persist through SNR and time-offset criteria are represented and equally weighted in the RF stack. The number of traces in each stack is annotated at the bottom right of the stack. The RFs traces are required to have (**left-to-right**) SNR ≥ 1.5 with no time-offset criteria; SNR ≥ 1.5 with the maximum time-offset allowed being 15 s; SNR ≥ 5 with no time-offset criteria; SNR ≥ 5 with the maximum time-offset allowed being 15 s. Note how the addition of the time-offset criteria greatly increases the resolution of mantle phases for SNR ≥ 1.5 without decreasing the number of traces as significantly as a higher SNR requirement, which resolves the same mantle phases with or without the time-offset requirement. The Moho signal in stacks with SNR ≥ 1.5 is not as clear as with SNR ≥ 5 , but in the actual data processing used in the main text, the SNR threshold is ultimately aided by the k-means culling and weighting procedure. The Sp RF polarity has been flipped to match the Ps convention.

Table S A-1. Location, number of events, and quality of each RF stack.

STA.	REGION	BAZ	VS @ 150KM	EV #	RAT ING
TUC 2	W. USA	145	4.302	87	143
TIA 1	E. NCC	128	4.408	44	59
LYN 1	COB, China	124	4.468	67	95
RAYN 2	Arabian Shield	063	4.483	46	88
RAYN 1	Arabian Shield	094	4.486	57	119
MBAR 1	Tanzania Craton	092	4.507	34	45
KONO 1	E. Eur. Craton	032	4.528	52	104
LZH 1	W. NCC	118	4.533	52	84
HHC 1	W. NCC	125	4.534	40	83
SYO 1	D.M. Land, Antarctica	147	4.565	37	75
KURK 3	Kazakh Uplands	140	4.586	39	86
KURK 1	Kazakh Uplands	118	4.587	119	213
RSSD 1	Wy. Craton, NA	313	4.601	68	107
RSSD 2	Wy. Craton, NA	148	4.601	35	53
ALE 2	Ngn., NA	209	4.603	65	112
HYB 1	Indian Craton	100	4.602	62	127
STKA 3	S. Aus. Craton	001	4.582	50	84
NWAO 1	W. Aus. Craton	020	4.621	38	60
CCM 2	Maz., NA	162	4.630	56	113
HYB 2	Indian Craton	043	4.631	40	51
TSUM 1	Congo Craton	239	4.636	41	74
ARU 1	E. Eur. Craton/Urals	115	4.650	144	189
CCM 1	Maz., NA	322	4.659	90	130
LSZ 1	Congo/Kalahari	083	4.674	120	147
BOSA 1	Kaapvaal, Kalahari	081	4.679	41	84
BBOO 3	S. Aus. Craton	005	4.683	40	91
FFC 1	THO, NA	312	4.684	99	133
DBIC 1	W. African Craton	238	4.6880	126	201
DBIC 2	W. African Craton	273	4.699	51	101

BRVK 1	Kazakh Uplands	119	4.718	212	329
YKWS 1	Slave craton, NA	299	4.720	47	68
ECSD 2	Superior, NA	157	4.721	60	123
EYMN 1	Superior, NA	320	4.733	59	99
OBN 1	SA, E. Eur.	054	4.737	164	202
JFWS 1	NA Craton	163	4.740	63	128
OBN 2	SA, E. Eur.	110	4.743	73	91
K43A 1	NA Craton	165	4.745	37	82
AGMN 1	Superior, NA	318	4.767	58	96
LBTB 1	Kaapvaal, Kalahari	80	4.769	71	137
AGO1 1	E. Antarctica	46	4.771	78	151
N140 1	E. Antarctica	78	4.785	165	272
P061 1	E. Antarctica	099	4.786	58	100
AGO3 1	E. Antarctica	148	4.853	59	96

S1.2 Receiver function processing & data selection

Directly following deconvolution

(described in the main text), we remove RFs with unrealistically large amplitudes in the mantle.

We define “unrealistically large amplitudes” as absolute amplitudes exceeding 0.25 of the parent component from 1 to -30 s.

S1.3 Further description of the k -means data culling procedure

Whenever we enact k-means, we perform the algorithm twice, each time with a different method for comparing RFs. The two methods we use in the k-means algorithm are the city-block and cosine distancing measurements. We qualitatively find that these distancing methods are best at separating good- and poor-quality seismograms compared to other distancing methods available for use within the k-means algorithm. City-block distancing determines clusters by absolute value of the difference between seismogram amplitudes, s , and amplitudes of the cluster’s centroid, c (Eq. 2; Fig. 1-3(a,left)). Cosine distancing compares the angle between the points, s , on a trace and the mean of points within an existing cluster, c' (Eq. 2; Fig. 1-3(a,right)).

$$distance(s, c) = \sum_{j=1}^p |s_j - c_j|$$

Equation 2

$$distance(s, c) = 1 - \frac{sc'}{\sqrt{(ss')(cc')}}$$

Equation 3

The user picks the clusters produced from each distancing method that exhibit low noise, low 0 s energy, and prominent Moho arrivals (Fig. 1-3).

After we complete the five iterations of k-means during the last step of data-culling, we weight seismograms in a median bootstrap stack. We weight seismograms picked 1-5 times by 0.1, 0.3, 0.75, 1, and 2.5, respectively. To evaluate the quality of RF stacks, we consider both the total number of traces stacked as well as the amount those traces are weighted. If a station has many independent traces that make it through our selection algorithm, we assume that the RF stack is relatively high quality. As an additional consideration, we sum the weights of the traces stacked. Trace weight corresponds to the number of times a trace was chosen in our k-means picking algorithm, making the sum an indication of data self-consistency. We only consider data with a minimum of 30 events and 50 rating but feel most confident in RF stacks with a minimum of 40 events and 70 rating. The event number and rating of all RF stacks is provided in Table S1.

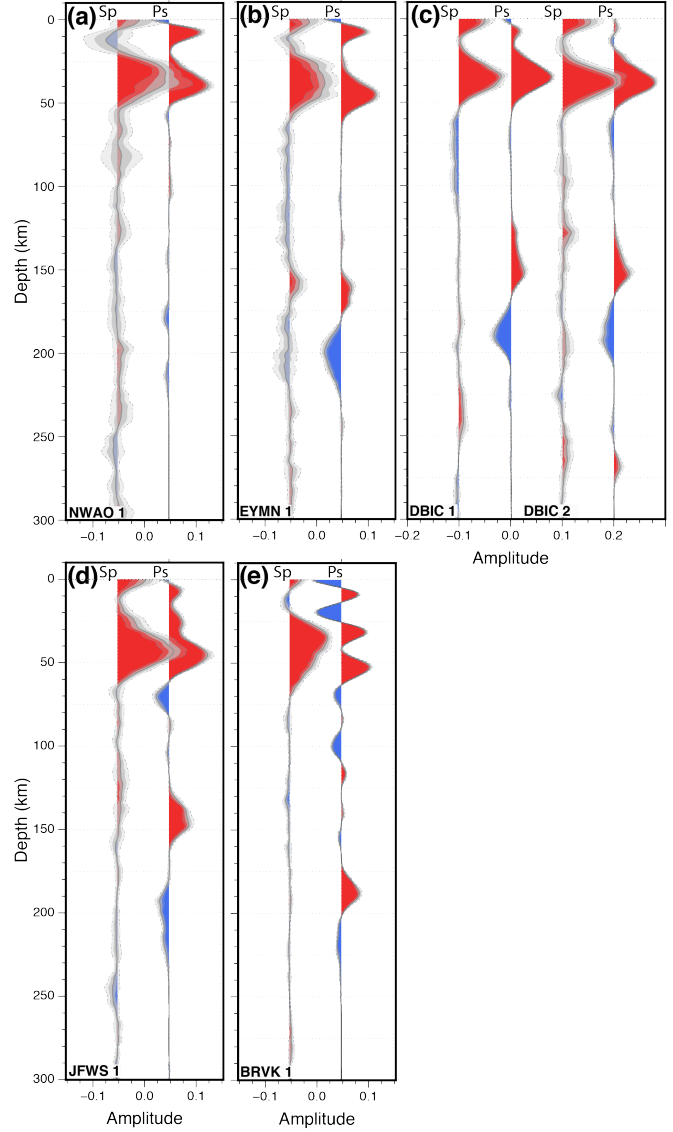


Figure S A-4. Sp and Ps comparisons

Comparisons of Sp RF stacks (left) with Ps RF stacks (right) from the same backazimuthal bin. **(a)** The phases observed in the Ps RF stack for NWA0 1 are remarkably similar to the Sp RF stacks and seems to lack any obvious crustal multiples. **(b)** The Ps RF stack for EYMN 1 corresponds well with the Sp RF stack with the exception of a set of simple crustal multiples at ~160 and 200 km depth. **(c)** Station DBIC shows lateral variation sampled by converted phases arriving at a single station. A low velocity MLD phase is observed below the Moho in the Sp stack of DBIC 1 and the Ps stack of DBIC 2 while no MLD is observed in the Sp stack of DBIC 2 and Ps stack of DBIC 1. **(d)** The Ps stack of JFWS 1 demonstrates the importance of using Sp RF stacks to interpret MLD phases. In the Ps stack, JFWS 1 has a strong negative phase below the Moho, which does not exist out of error in the Sp stack. It is likely this phase has been amplified in the Ps stack by interference with crustal multiples. **(e)** The Ps and Sp stacks of BRVK 1 are vastly different. These differences are likely due to crustal multiples in the Ps stack as well as lateral variation that can occur in the mantle sampled close the station (Ps) and ~50 km away from the station (Sp).

Text S2. Extended Results

Calculations of conversion point locations are made with the TauP toolkit (Crotwell et al., 1999).

S2.1 Comparisons with select Ps RF stacks

S-to-P conversions in the upper mantle occur 50-100 km away (map-view) from the station at which they are recorded. Because of this, S-to-P phases recorded at a seismometer from events of different backazimuths can be over 100 km apart. To isolate lateral variations that may exist around a seismic station, we group events by backazimuth in 30° bins before stacking. We also calculate RFs for P-to-S (Ps) converted phases (Fig. SA58-100). Ps phases arrive at the seismometer at a steeper angle, causing conversion points in the mantle to occur close to the seismometer. We generally do not interpret Ps RFs for mantle structure because interference with crustal multiples of P-to-S converted phases arriving at similar times as mantle phases is possible, but they provide interesting insights when compared to Sp RF stacks (Fig. SA-4). When evaluating Ps and Sp RF stacks from similar backazimuthal bins, we see a range of similarities. Some Ps and Sp RF stacks are virtually identical, with no apparent crustal multiples in the Ps stack (i.e. NWA0 1, Fig. SA-4(a)). Some Ps RF stacks have obvious crustal multiples intertwined with energy that we are confident is from the upper mantle because it is also in the Sp RF stack (i.e. EYMN 1, Fig. SA-4(b)).

At long-running stations with high quality data, we are able to compare multiple backazimuth binned RF stacks to gain information on local lateral variations in mantle structure (i.e. OBN 1 & 2, DBIC 1 & 2, CCM 1 & 2, Fig. 1-6). We observe a particularly interesting instance of lateral variation in mantle structure when comparing Ps and Sp RF stacks for DBIC 1 and 2 (Fig. SA-4(c)). DBIC 1 and 2 sample mantle to the west and southwest of the seismometer, respectively. Interestingly, a low velocity MLD is observed below the Moho in the DBIC 1 Sp

and DBIC 2 Ps RF stacks, while no phases between the Moho and 100 km are observed in the DBIC 2 Sp or DBIC 1 Ps RF stacks. This observation makes apparent that upper mantle lateral variation in this region is complex and highlights how MLD occurrence can have significant variation at local scales.

Previous work interprets upper mantle structure from Ps RF stacks (e.g. Bostock, 1998). As we have shown, Ps phases migrated to upper mantle depths can be due to Earth structure that exists at upper mantle depths, but these phases can be dominated by or interfered with by crustal multiples. The Ps stack of JFWS 1 demonstrates the importance of using Sp RF stacks to interpret MLD phases (Fig. SA-4(d)). In the Ps stack, JFWS 1 has a strong negative phase below the Moho, which does not exist out of error in the Sp stack. It is likely this phase has been amplified in the Ps stack by interference with crustal multiples and should not be interpreted as an MLD.

Finally, we find that in some backazimuthal bins, all of these phenomena compound (crustal multiples, higher frequency sampling, lateral variation) to result in a Ps RF stack that is extremely different from the corresponding Sp stack (Fig. SA-4(e)). In this example, we are confident in the results of the BRVK 1 Ps and Sp RF stacks because of the high quantity of data.

S2.2 Global distribution of phases

We analyze phases based on the shear velocity at 150 km depth from the SEMum2 velocity model as well as thermotectonic age at the location of the phase's conversion point. To visualize this, we have plotted all phases out of two standard deviations at their conversion points projected to the surface in map-view in Fig. SA-5.

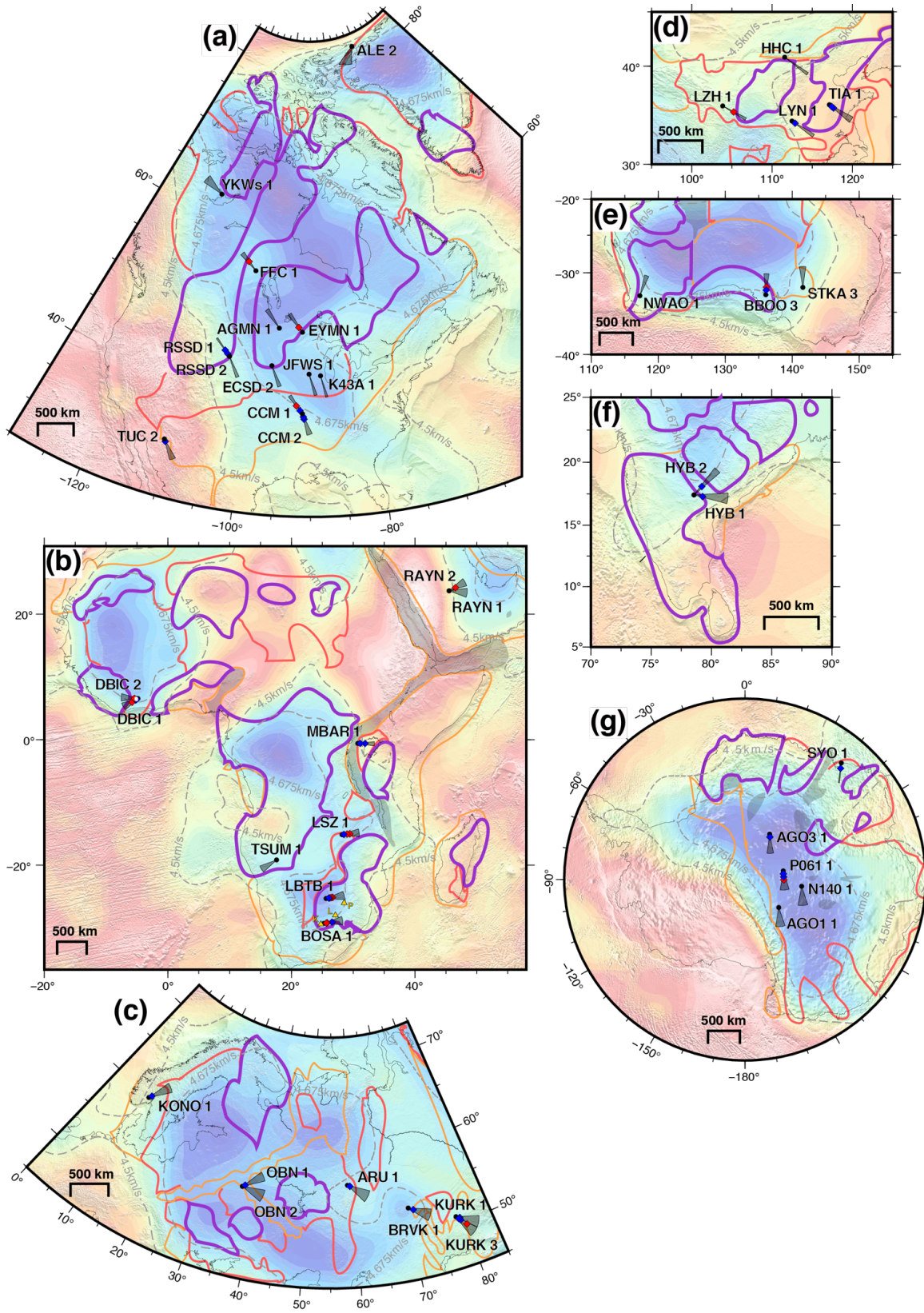


Figure S A-5. Conversion points on SEMum2

Conversion points projected to the surface (diamonds) for PVGs (red) and NVGs (blue) between the Moho and 150 km depth (Fig. 1-6) plotted for the regions of **(a)** North America, **(b)** Africa (white circles are DBIC Ps 75 km conversion points), **(c)** Eurasia, **(d)** China, **(e)** Australia, **(f)** India, and **(g)** Antarctica. Outlines indicate thermotectonic age (Modified from Artemieva (2006)): Archean (purple), Paleoproterozoic (red), Meso- & Neo-Proterozoic (orange). Regions of Phanerozoic age internal to continents are shaded. Yellow triangles are locations for the Kimberley (K), Lace (L), and Premier (P) mines in Africa. Regions are underlain by shear velocity at 150 km depth from SEMum2 (French et al., 2013) and shear velocities of 4.5 and 4.675 km/s are contoured with gray dashed lines. Stations are black circles and backazimuthal bins are represented by gray wedges, which indicate the backazimuthal range that events in that bin sample plotted to the lateral distance of the 300 km conversion point.

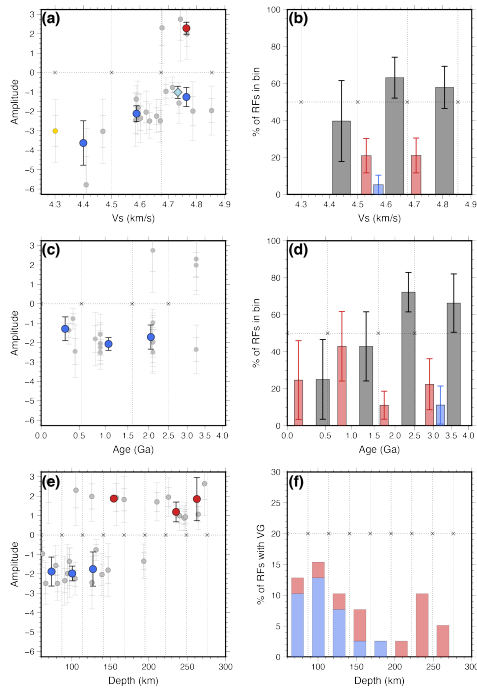


Figure S A-6. Fig. 1-9 with additional error criteria
(a-d) Same as Fig. 1-9 and **(e-f)** Fig. 1-7, but with additional amplitude-to-error criteria.

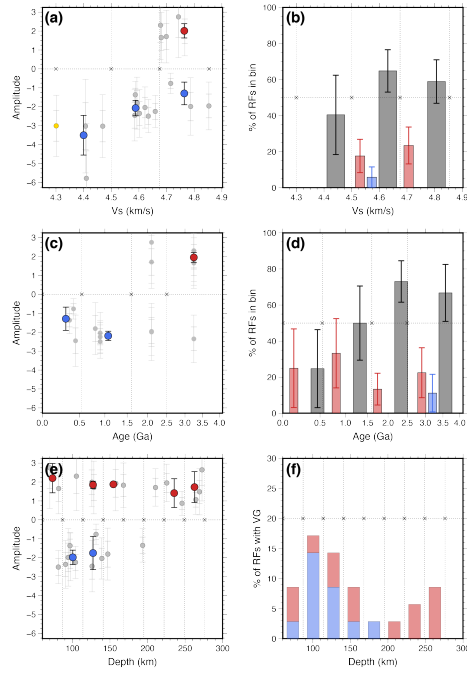


Figure S A-7. Fig. 1-9 without potential side-lobe RFs

(a-d) Same as Fig. 1-9 and (e-f) Fig. 1-7, but without RF stacks ARU 1, OBN 1, DBIC 1, and LSZ 1 which we are skeptical may exhibit side-lobe energy (main text, section 3).

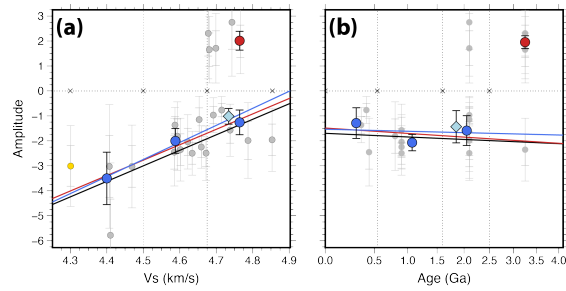


Figure S A-8. Fig. 1-9 with linear fits

Same as (a) Fig. 1-9(a) and (b) Fig. 1-9(c). Linear fits for the individual negative amplitudes (gray circles) shown as a black line; to the binned averages of the negative amplitudes (dark blue circles) shown as a red line; and to the binned averages of the negative amplitudes without Antarctic stations (light blue diamonds replace dark blue circles) shown as a blue line. The best fit linear regressions for the individual amplitudes and binned amplitudes are consistent. (a) R^2 values for the black, red, and blue lines are 0.434, 0.969, and 0.989, respectively. The R^2 value for the individual measurements is much lower, as expected given the scatter in the amplitude measurements as a function of V_s . This scatter illustrates that multiple factors influence velocity gradient amplitude, in addition to the background V_s of the mantle. (b) R^2 values for the black, red, and blue lines are 0.022, 0.107, and 0.014, respectively. Low R^2 values for all linear fits of thermotectonic age with amplitude indicate that a linear trend is not present in the data.

S2.3 Error bars on mantle phases

In order to be cautious in our interpretation due to the skepticism surrounding MLDs, we perform our analysis with a more conservative error criterion as well (Fig. SA-6). We find that this confirms the large-scale trends we see in MLD amplitude and occurrence with shear velocity structure and thermotectonic age. For this additional error criterion, the ratio of the amplitude that is out of 2 standard deviations to one standard deviation, must be at least 50%. We only consider phases, where the following statement is true:

$$\frac{|meanAmp| - 2 * stdDev}{stdDev} > \frac{1}{2}$$

S2.4 Testing the impact of uncertainty in velocity models

Regions with sparse distributions of seismic instrumentation do not produce reliable seismic velocity model results. This argument is supported by the variation in the shape of the velocity gradients in sparsely sampled regions when SEMum2 is compared to other global velocity models such as GLADM15 (Fig. 1-1, S5, SA-9-11; Bozdag et al., 2016). When we perform the analysis from Fig. 1-7 & 1-9 using the GLADM15 model, results are not significantly different, indicating our conclusions are not dependent on the global velocity model (Fig. SA-11). The largest differences in the velocity models are in the Arabian Shield, Antarctica, and southern Australia. We reanalyze the data using the SEMum2 model but omit these regions (Fig. SA-12).

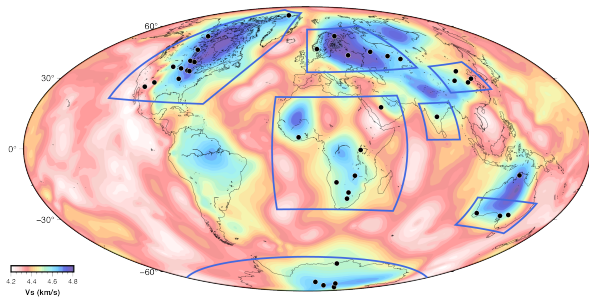


Figure S A-9. Fig. 1-1 with GLADM15

Same as Fig. 1-1, but with the GLADM15 velocity model (Bozdag et al., 2016).

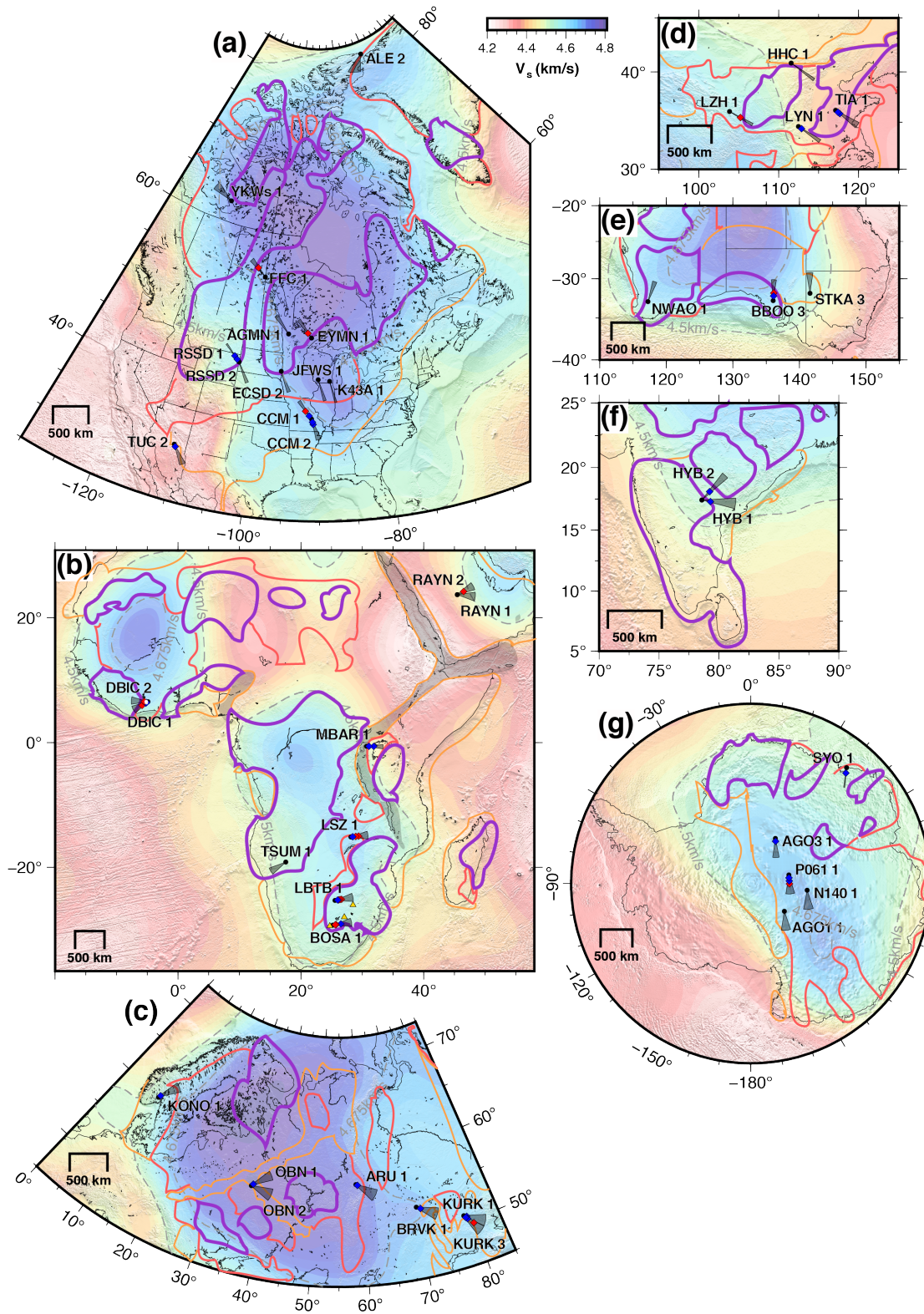


Figure S A-10. Regional maps with GLADM15

Same as Fig. SA-5, but with the GLADM15 velocity model (Bozdag et al., 2016).

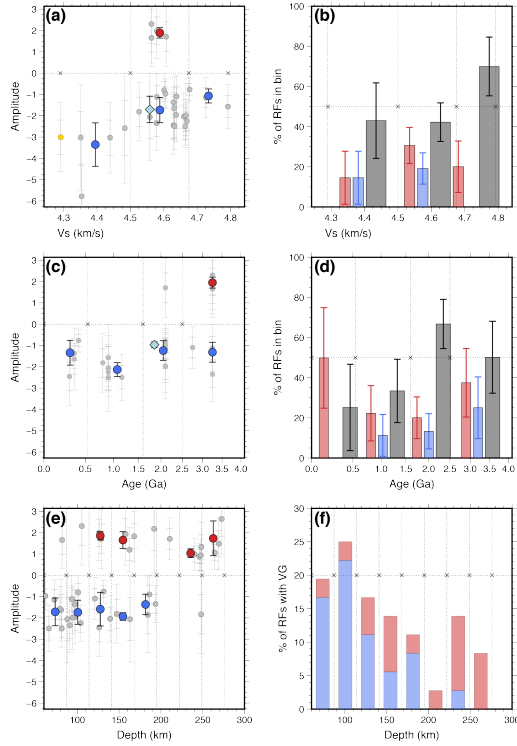


Figure S A-11. Fig. 1-9 with GLADM15

(a-d) Same as Fig. 1-9 and (e-f) Fig. 1-7, but calculated with shear velocities from the GladM15 global model.

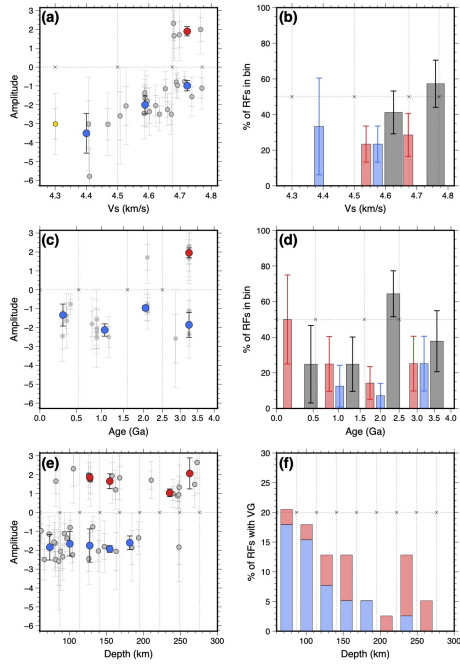


Figure S A-12. Fig. 1-9 without select stacks

Figure S12. (a-d) Same as Fig. 1-9 and (e-f) Fig. 1-7 with RF stacks from the Arabian Shield (RAYN), central Antarctica, and southern Australia omitted.

S2.5 The Kaapvaal craton

The RF stacks we observe in the cratonic interior of the Kaapvaal craton exhibit more significant mid-lithospheric PVGs than other RF stacks in this study. This result is consistent with other geophysical and geochemical observations from the Kaapvaal craton that indicate complexity in the Kaapvaal mantle that is not widely observed in other Archean cratons (e.g. Hoal, 2003; Pedersen et al., 2009). Most craton-interior volcanism globally has occurred in the Phanerozoic, while the Kaapvaal craton experienced significant Paleoproterozoic volcanism during the emplacement of the Bushveld Igneous Complex (Hoal, 2003; Kgaswane et al., 2012). The difference in geochemistry between Kaapvaal craton xenoliths and craton xenoliths from regions of Phanerozoic volcanism indicates that refertilization of the lithospheric mantle involved different processes in the Precambrian than it does today (Hoal, 2003).

One model is that the Bushveld Igneous Complex served as a source for silicate melt that was rich in iron and aluminum, causing alteration to the Kaapvaal lithosphere that differs from more recent potassium-rich alteration of cratonic lithosphere (Hoal, 2003). However, the presence of iron-rich altered lithosphere would predict relatively low seismic velocities in the Kaapvaal craton. This finding contrasts with our observation of predominately PVGs (velocity increases with depth) in the Kaapvaal mantle. In particular, LBTB 1 is the only RF stack that obviously disagrees with the dominant global MLD trends we observe that cannot be explained by the global resolution of the reference velocity model (French et al., 2013). LBTB 1 is located in the Kaapvaal craton and samples near the eastern boundary of the Bushveld Igneous Complex (Kgaswane et al., 2012). This stack has a high amplitude, distributed PVG from 130-250 km depth that requires a ~10% shear velocity increase (Fig. 1-8).

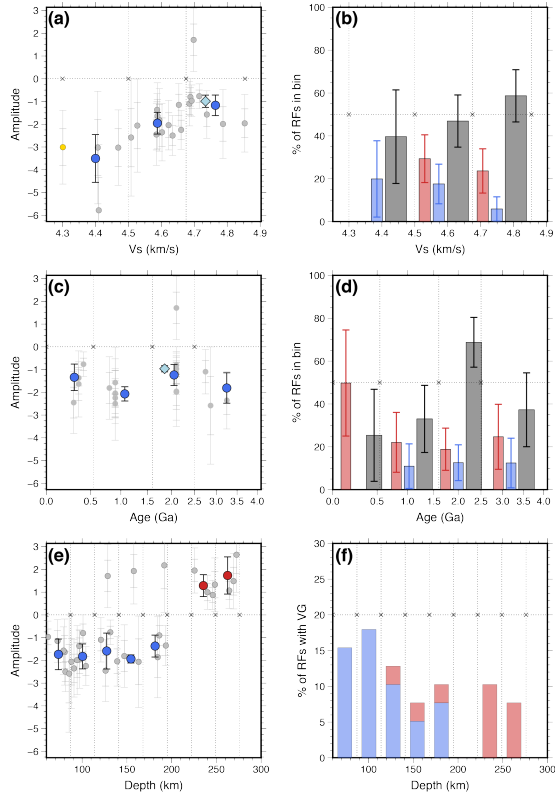


Figure S A-13. Fig. 1-9 without S Africa stacks

Figure S13. (a-d) Same as Fig. 1-9 and **(e-f)** Fig. 1-7 with southern African RF stacks omitted.

In an alternative model, the presence of eclogite and/or diamond could contribute to a PVG in the upper mantle (Garber et al., 2018). It is notable that eclogite sourced from 120-170 km depths from the Lace kimberlite mine in the Kaapvaal craton is found to have exceptionally fast shear velocities (≥ 4.8 km/s; Fig. SA-5(b), yellow triangle - L) (Aulbach et al., 2020), which corresponds clearly to the PVGs from 130-200 km required to model the positive phases observed at station LBTB (Fig. 1-8). Additionally, the shallower mantle PVGs in BOSA could be explained by the transition from spinel to garnet stability, or the Hales discontinuity, expected to be at ~ 100 km depth (Lebedev et al., 2009; Revenaugh & Jordan, 1991).

Due to the exceptional nature of southern African RF results, we reanalyze data omitting these stations (Fig. SA-13), with the most significant change being the presence of positive phases in the upper mantle.

S2.6 Comparison with previous work: North America

Hopper and Fischer (2018) find regions with low velocity MLD phases in the northern Great Plains of the United States (north-central US) using frequency domain deconvolution and common conversion point stacking of Sp phases, while Shearer and Buehler (2019) do not observe upper mantle NVGs in this region using long-period reflection seismology. We compared the results from individual stations in this study in the northern Great Plains (Fig. SA-5(a)) to the map of common conversion point stack amplitudes from *Hopper and Fischer (2018)* (Fig. SA-14). Most stations show good agreement, although in one case we observe a significant phase where the *Hopper and Fischer (2018)* amplitudes are low, and in two cases we do not observe a phase where the *Hopper and Fischer (2018)* amplitudes are relatively high.

Furthermore, Kind et al. (2020) do not observe a ubiquitous low velocity MLD below the Moho in cratonic North America, but do observe scattered low amplitude MLDs throughout the cratonic lithosphere as well as significant PVGs in the upper mantle of the northern Great Plains. Overall, we observe only intermittent NVGs and PVGs at a variety of lithospheric depths in the cratonic US (Fig. 1-6, SA-5(a)).

In order to compare our results with these studies, we chose a large proportion of RF stacks in North America relative to other continental regions. To ensure this over-representation of North America in our global analysis is not skewing our findings, we preform our analysis for all RF stacks that are not in North America and find similar conclusions (Fig. SA-15).

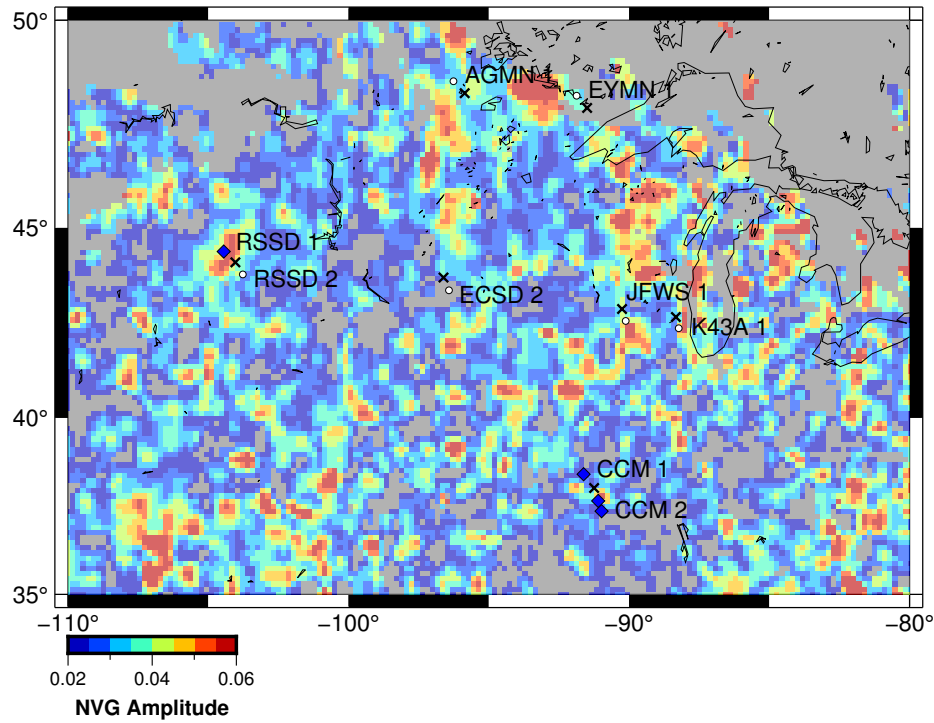


Figure S A-14. Hopper & Fischer (2018) comparison

Our results for the northern Great Plains of the US (blue diamonds = NVGs (Moho to 150 km) at their conversion point locations; white circles represent the MLD depth conversion point from *Hopper and Fischer (2018; Fig. 5(a))* at locations we do not observe an MLD. Background colors represents the negative phase amplitude in the upper mantle found by *Hopper and Fischer (2018; Fig. S1(a))*. We make the assumption that the smallest of the *Hopper and Fischer (2018)* amplitudes (<0.04) to be the least robust, so we consider amplitudes <0.04 to have no MLD and amplitudes >0.04 as MLDs. If our results were in agreement, blue diamonds would correspond to regions of green - red and white circles would correspond to regions of gray - blue. Our results are generally in agreement except for the results at CCM 1 where we observe a significant MLD and the amplitude in the *Hopper and Fischer (2018)* map is relatively low, and at JFWS 1 and K43A 1 where we do not observe MLDs but the *Hopper and Fischer (2018)* amplitude is higher.

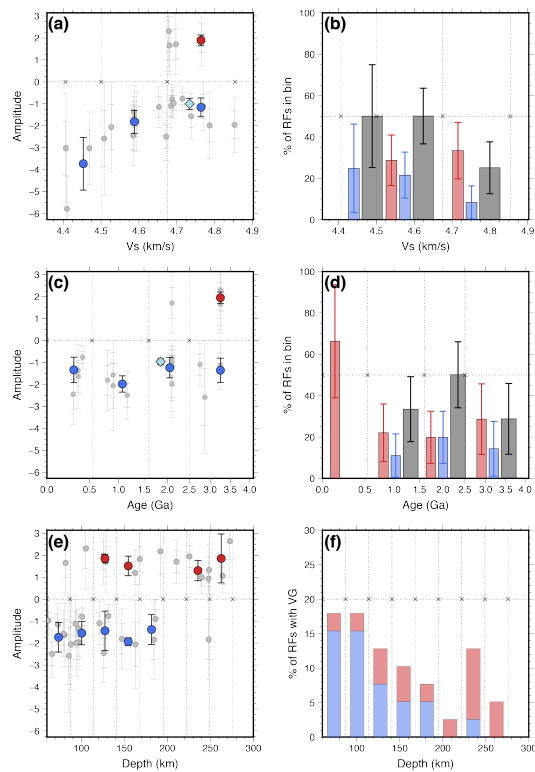


Figure S A-15. Fig. 1-9 without N America

(a-d) Same as Fig. 1-9 and (e-f) Fig. 1-7, with North American RF stacks omitted.

TUC IU 2 Cluster, Tot Ev 87, Rating 143

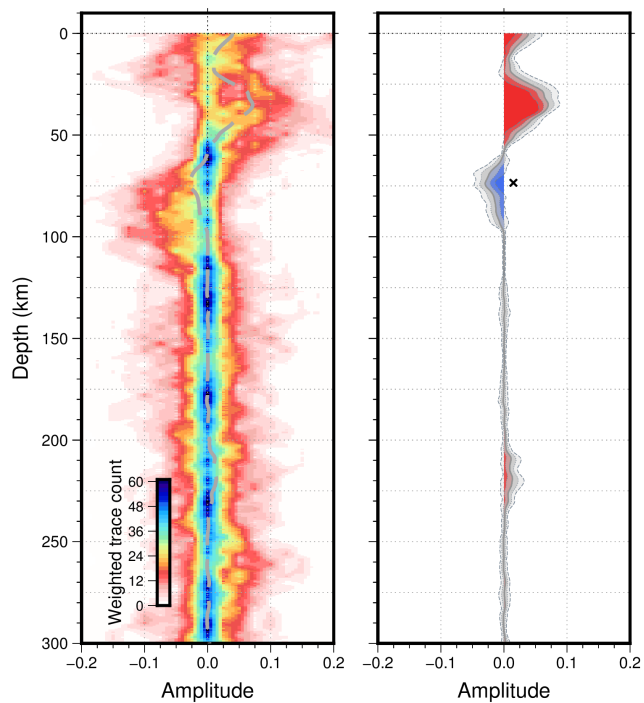


Figure S A-16. Sp TUC 2

Complete Sp RF stack results for TUC 2. Plotted like Figure 1-5.

TIA CB 1 Cluster, Tot Ev 44, Rating 59

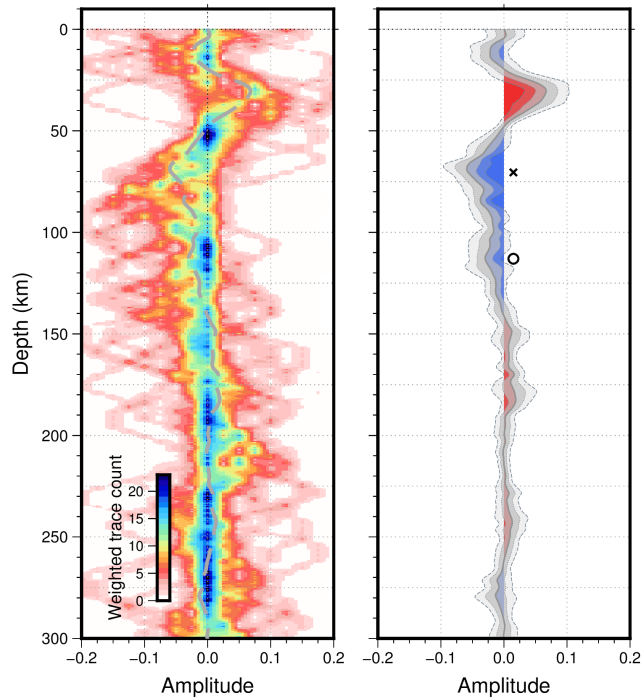


Figure S A-17. Sp RF TIA 1

Complete Sp RF stack results for TIA 1. Plotted like Figure 1-5.

LYN CB 1 Cluster, Tot Ev 67, Rating 95

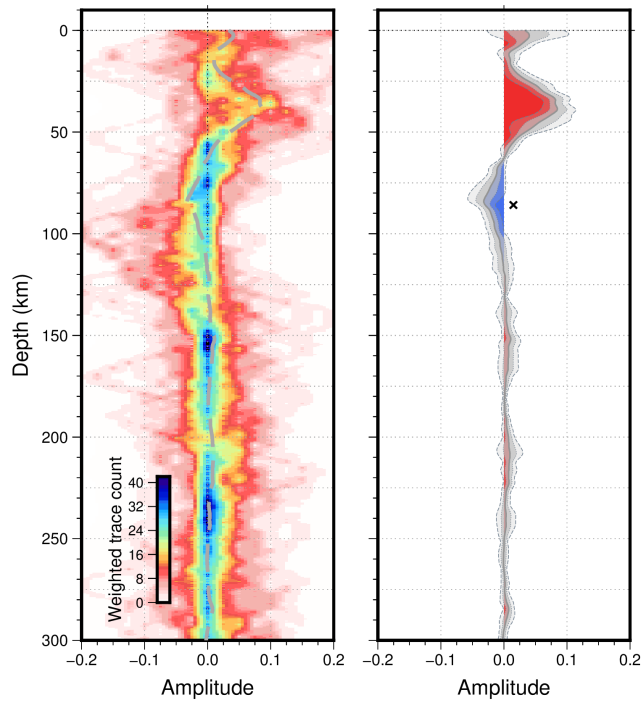


Figure S A-18. Sp RF LYN 1

Complete Sp RF stack results for LYN 1. Plotted like Figure 1-5.

RAYN II 2 Cluster, Tot Ev 46, Rating 88

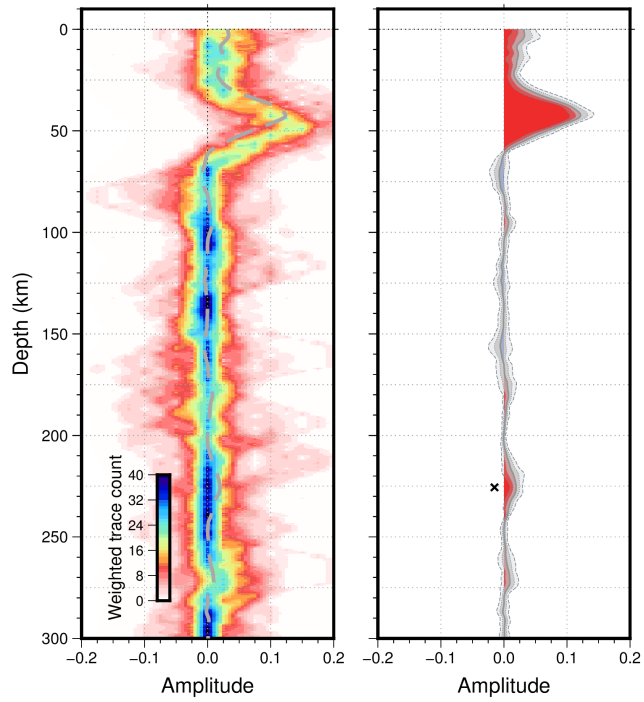


Figure S A-19. Sp RF RAYN 2

Complete Sp RF stack results for RAYN 2. Plotted like Figure 1-5.

RAYN II 1 Cluster, Tot Ev 57, Rating 119

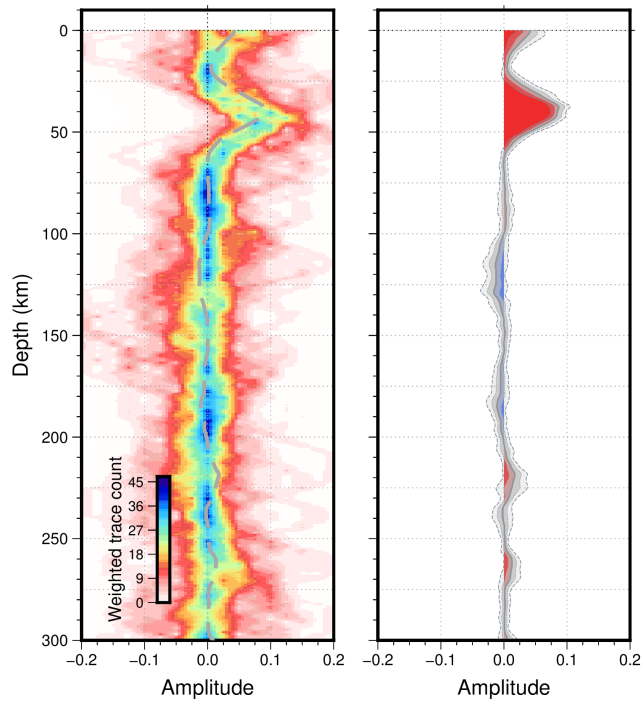


Figure S A-20. Sp RF RAYN1

Complete Sp RF stack results for RAYN 1. Plotted like Figure 1-5.

MBAR II 1 Cluster, Tot Ev 34, Rating 45

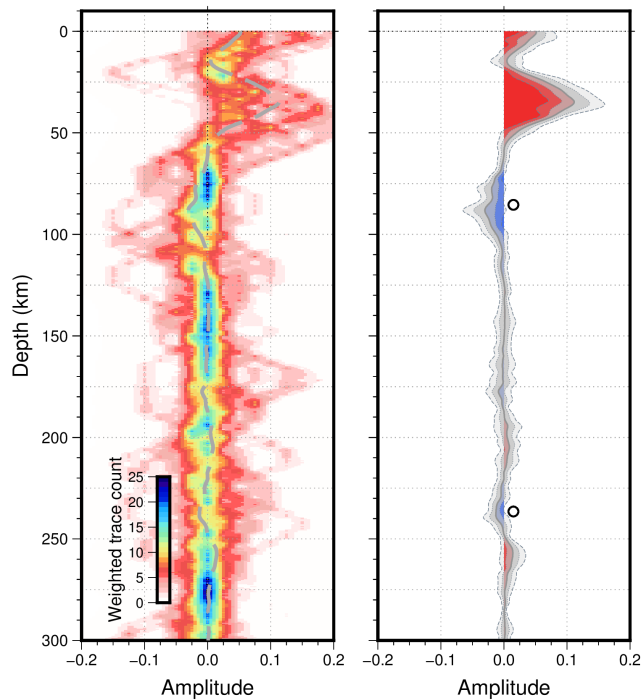


Figure S A-21. Sp RF MBAR 1

Complete Sp RF stack results for MBAR 1. Plotted like Figure 1-5.

KONO IU 1 Cluster, Tot Ev 52, Rating 104

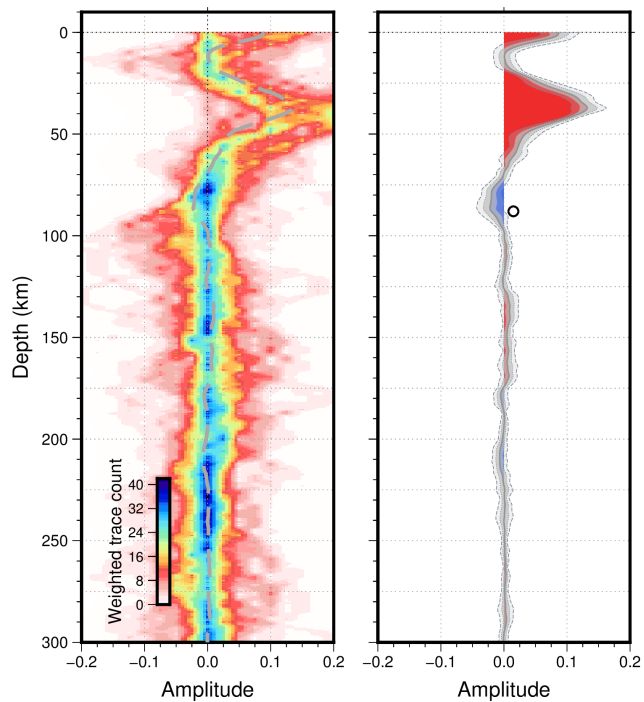


Figure S A-22. Sp RF KONO 1

Complete Sp RF stack results for KONO 1. Plotted like Figure 1-5.

LZH CB 1 Cluster, Tot Ev 52, Rating 84

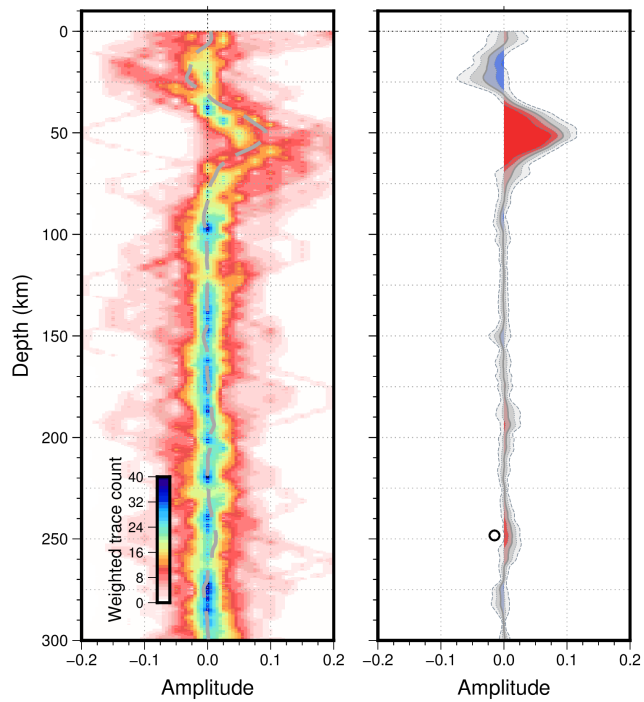


Figure S A-23. Sp RF LZH 1

Complete Sp RF stack results for LZH 1. Plotted like Figure 1-5.

HHC CB 1 Cluster, Tot Ev 40, Rating 83

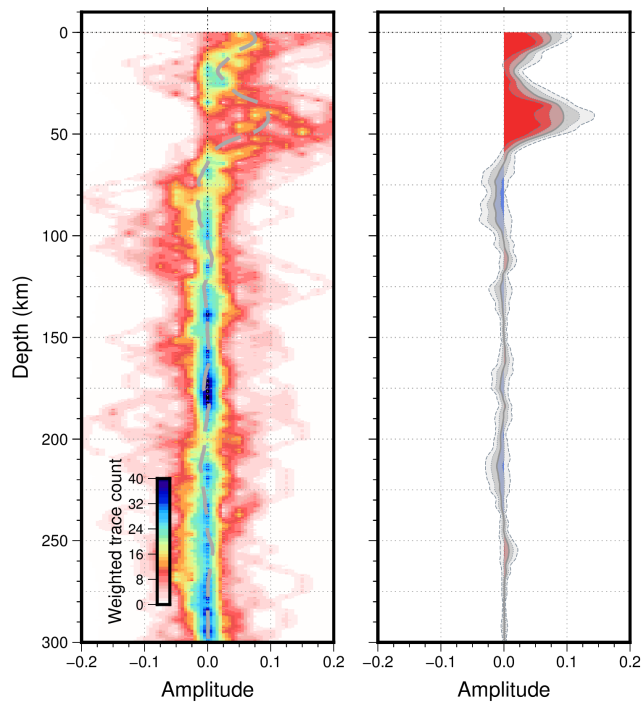


Figure S A-24. Sp RF HHC 1

Complete Sp RF stack results for HHC 1. Plotted like Figure 1-5.

SYO PS 1 Cluster, Tot Ev 37, Rating 75

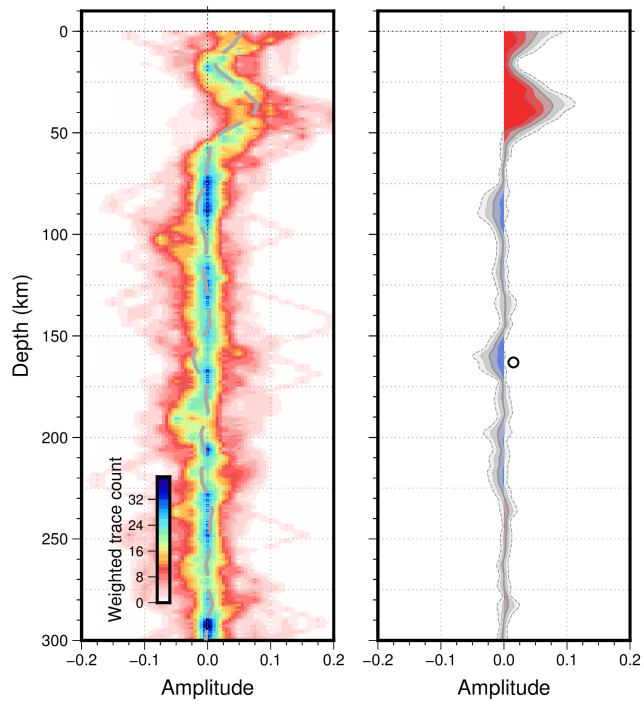


Figure S A-25. Sp RF SYO 1

Complete Sp RF stack results for SYO 1. Plotted like Figure 1-5.

KURK II 3 Cluster, Tot Ev 39, Rating 86

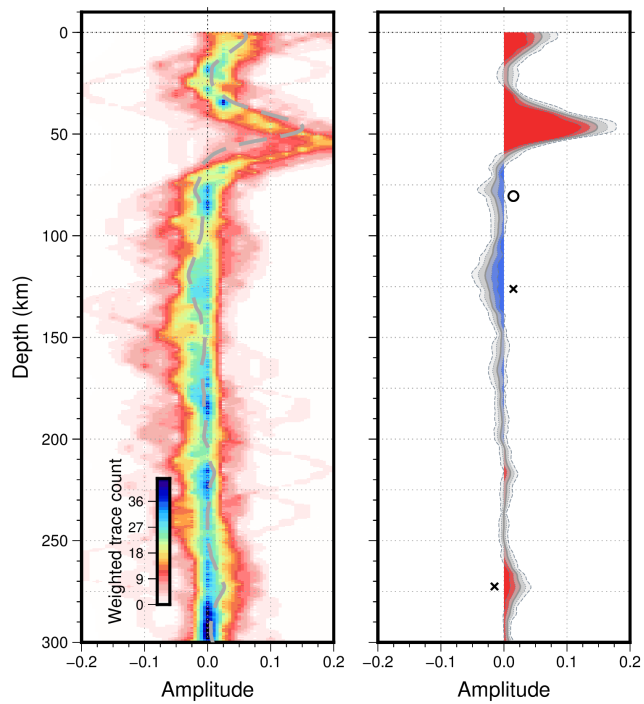


Figure S A-26. Sp RF KURK 3

Complete Sp RF stack results for KURK 3. Plotted like Figure 1-5.

KURK II 1 Cluster, Tot Ev 119, Rating 213

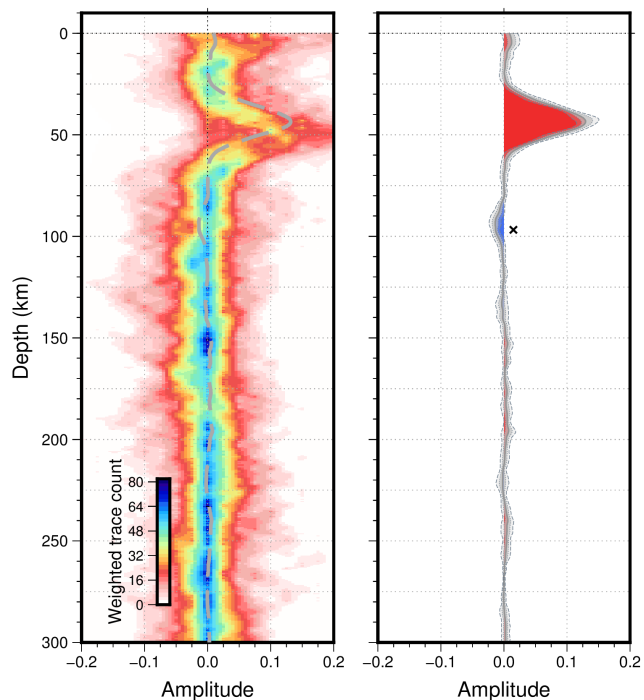


Figure S A-27. KURK 1

Complete Sp RF stack results for KURK 1. Plotted like Figure 1-5.

RSSD IU 1 Cluster, Tot Ev 68, Rating 107

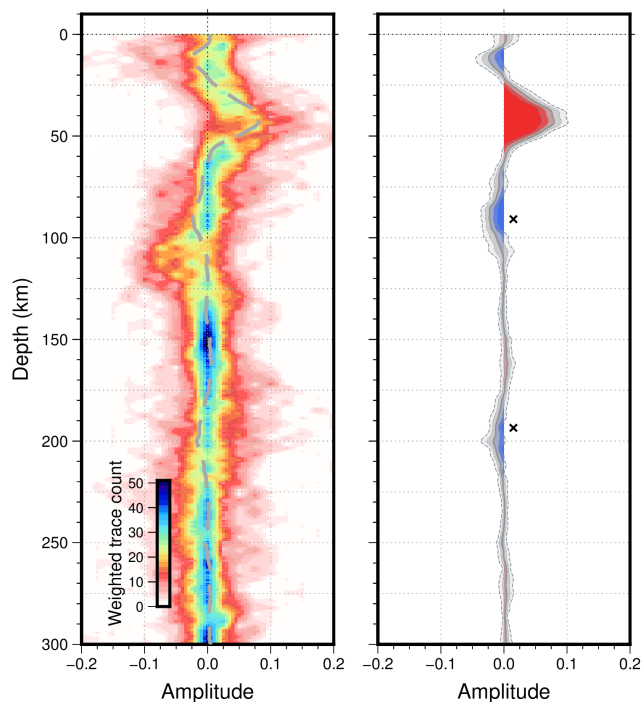


Figure S A-28. Sp RF RSSD 1

Complete Sp RF stack results for RSSD 1. Plotted like Figure 1-5.

RSSD IU 2 Cluster, Tot Ev 35, Rating 53

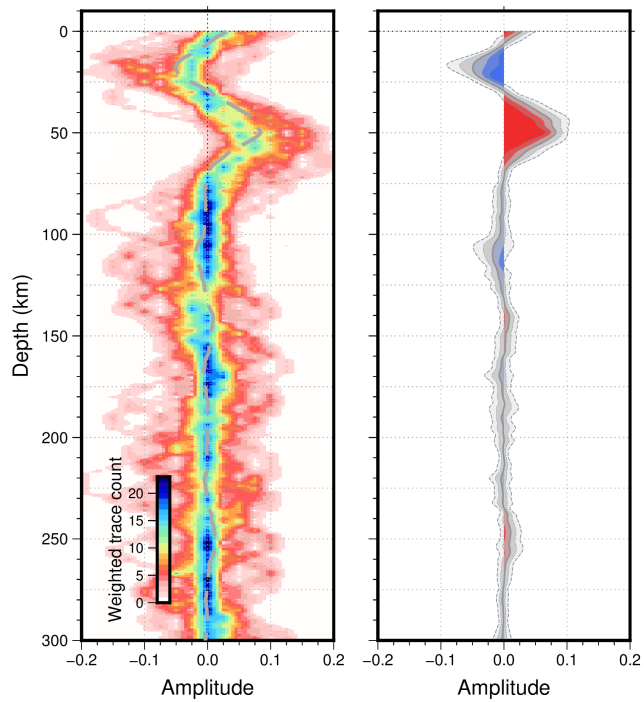


Figure S A-29. Sp RF RSSD 2

Complete Sp RF stack results for RSSD 2. Plotted like Figure 1-5.

ALE II 2 Cluster, Tot Ev 65, Rating 112

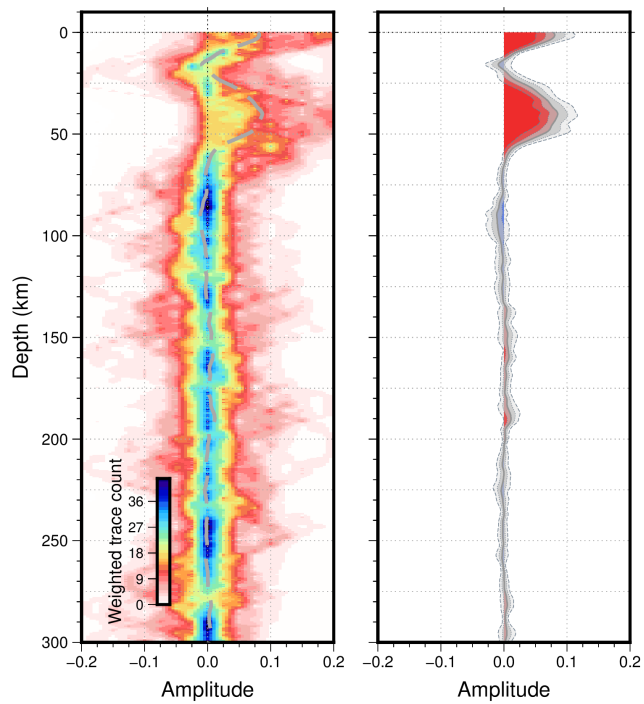


Figure S A-30. Sp RF ALE 2

Complete Sp RF stack results for ALE 2. Plotted like Figure 1-5.

HYB G 1 Cluster, Tot Ev 62, Rating 127

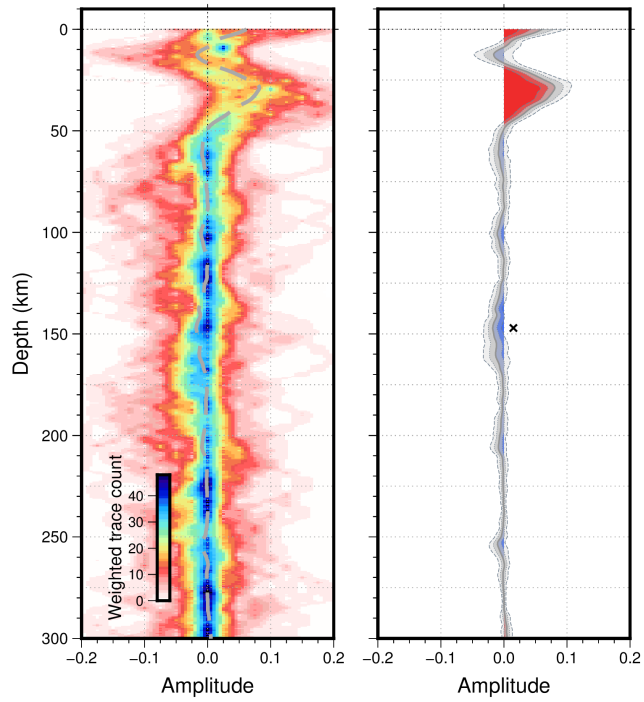


Figure S A-31. Sp RF HYB 1

Complete Sp RF stack results for HYB 1. Plotted like Figure 1-5.

STKA AU 3 Cluster, Tot Ev 50, Rating 84

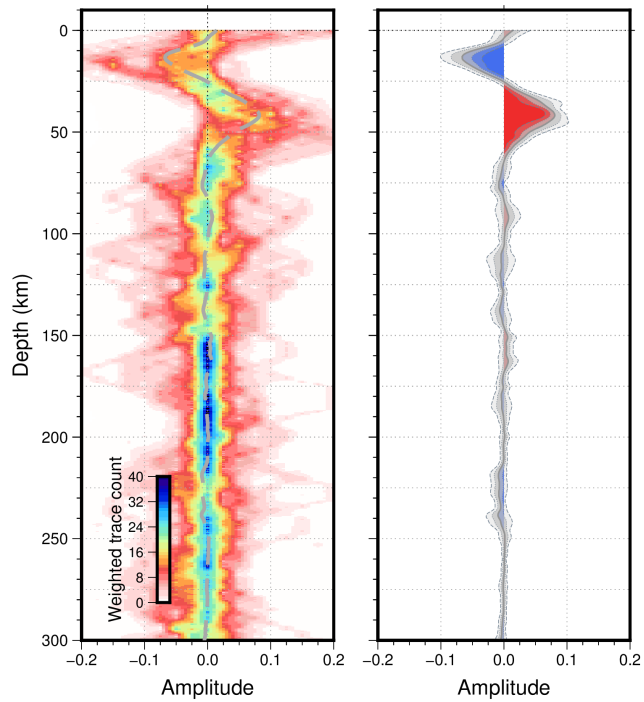


Figure S A-32. Sp RF STKA 3

Complete Sp RF stack results for STKA 3. Plotted like Figure 1-5.

NWAO IU 1 Cluster, Tot Ev 38, Rating 60

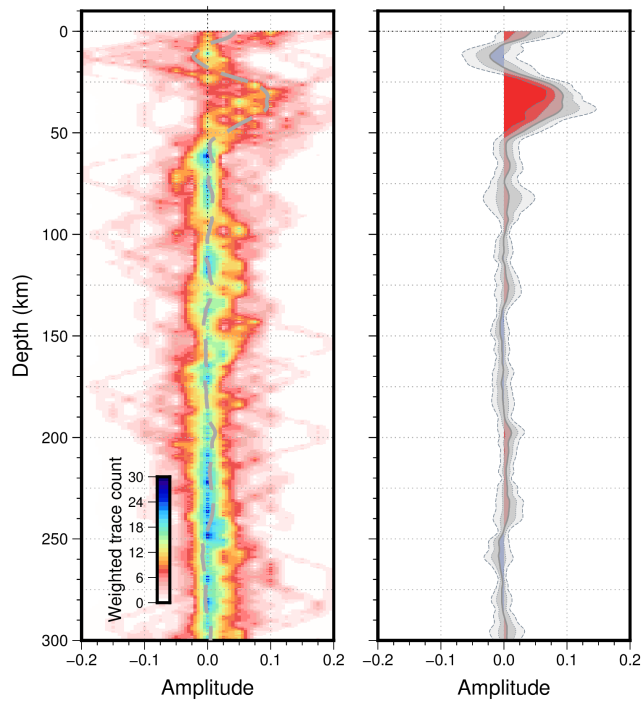


Figure S A-33. NWAO 1

Complete Sp RF stack results for NWAO 1. Plotted like Figure 1-5.

CCM IU 2 Cluster, Tot Ev 56, Rating 113

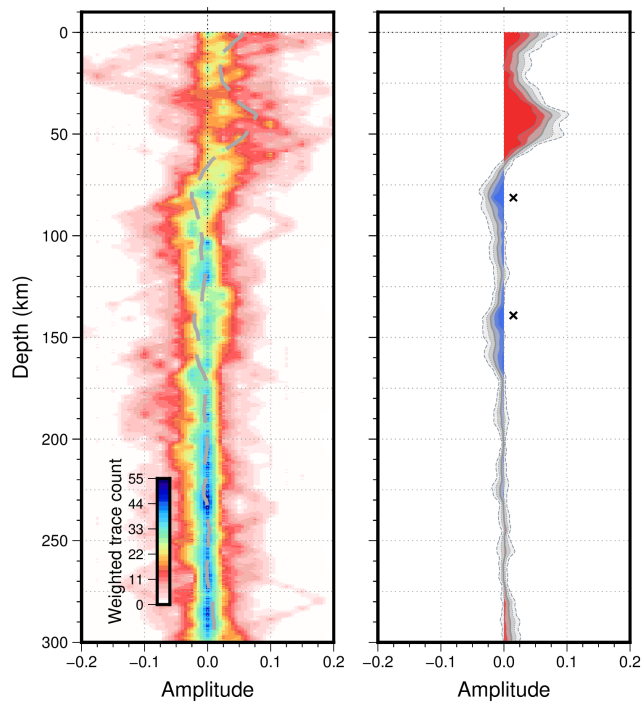


Figure S A-34. Sp RF CCM 2

Complete Sp RF stack results for CCM 2. Plotted like Figure 1-5.

HYB G 2 Cluster, Tot Ev 40, Rating 51

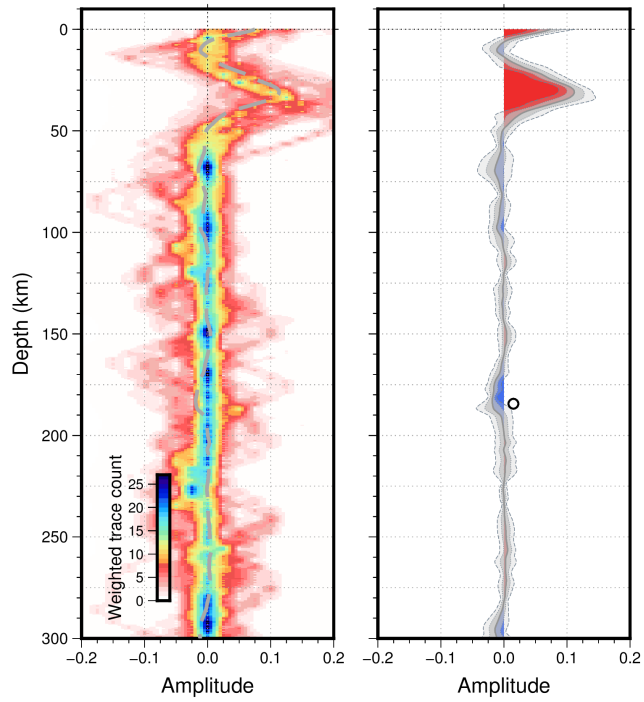


Figure S A-35. Sp RF HYB 2

Complete Sp RF stack results for HYB 2. Plotted like Figure 1-5.

TSUM IU 1 Cluster, Tot Ev 41, Rating 74

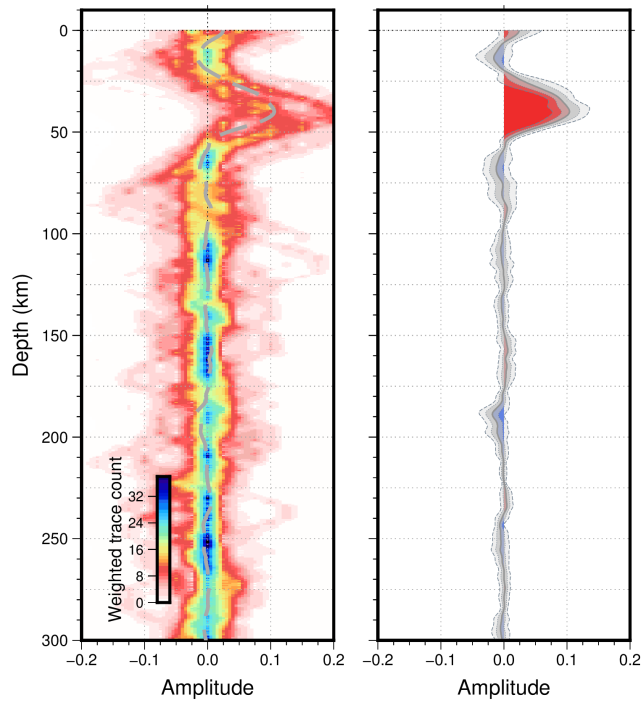


Figure S A-36. Sp RF TSUM 1

Complete Sp RF stack results for TSUM 1. Plotted like Figure 1-5.

ARU II 1 Cluster, Tot Ev 144, Rating 189

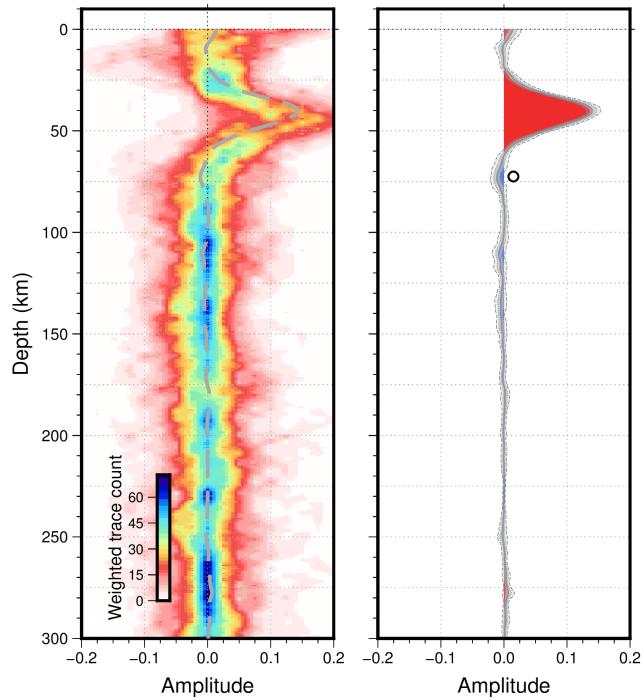


Figure S A-37. Sp RF ARU 1

Complete Sp RF stack results for ARU 1. Plotted like Figure 1-5.

CCM IU 1 Cluster, Tot Ev 90, Rating 130

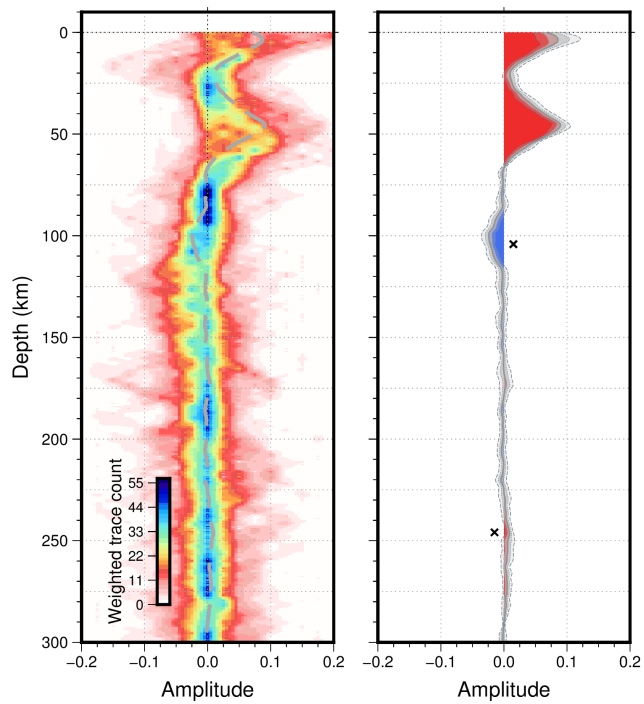


Figure S A-38. Sp RF CCM 1

Complete Sp RF stack results for CCM 1. Plotted like Figure 1-5.

LSZ IU 1 Cluster, Tot Ev 120, Rating 147

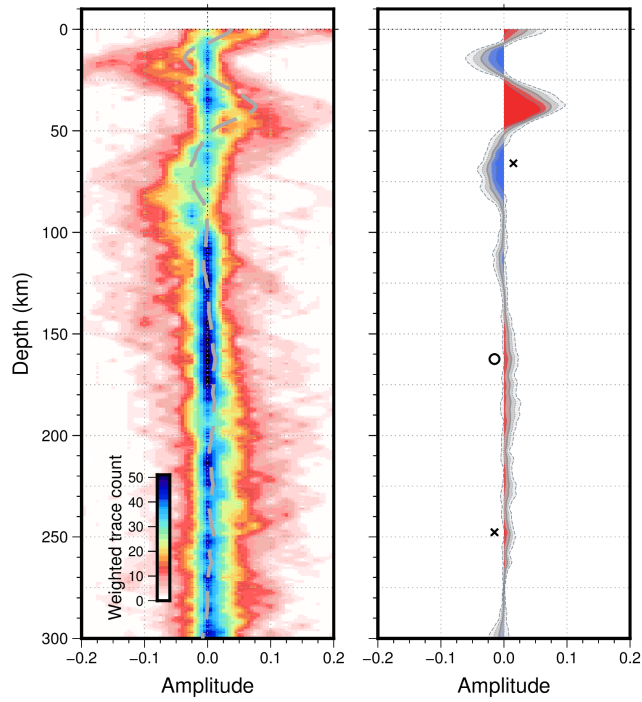


Figure S A-39. Sp RF LSZ 1

Complete Sp RF stack results for LSZ 1. Plotted like Figure 1-5.

BOSA GT 1 Cluster, Tot Ev 41, Rating 84

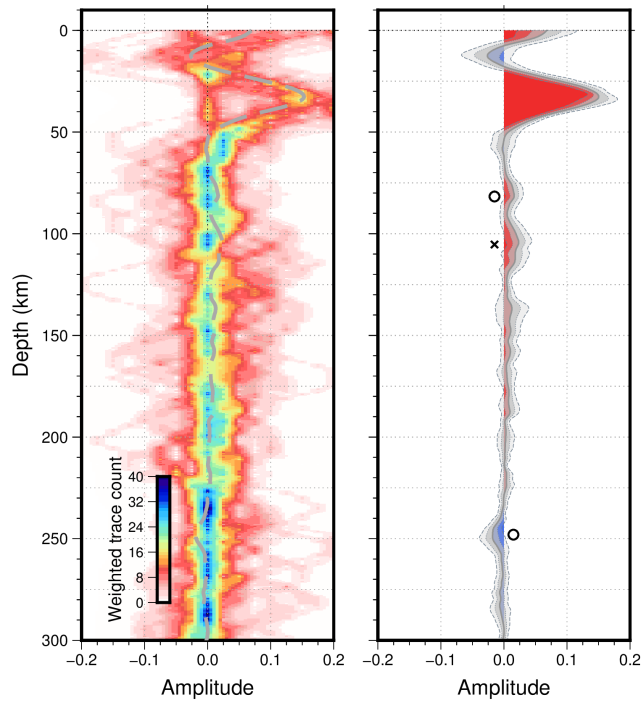


Figure S A-40. Sp RF BOSA 1

Complete Sp RF stack results for BOSA 1. Plotted like Figure 1-5.

BBOO AU 3 Cluster, Tot Ev 40, Rating 91

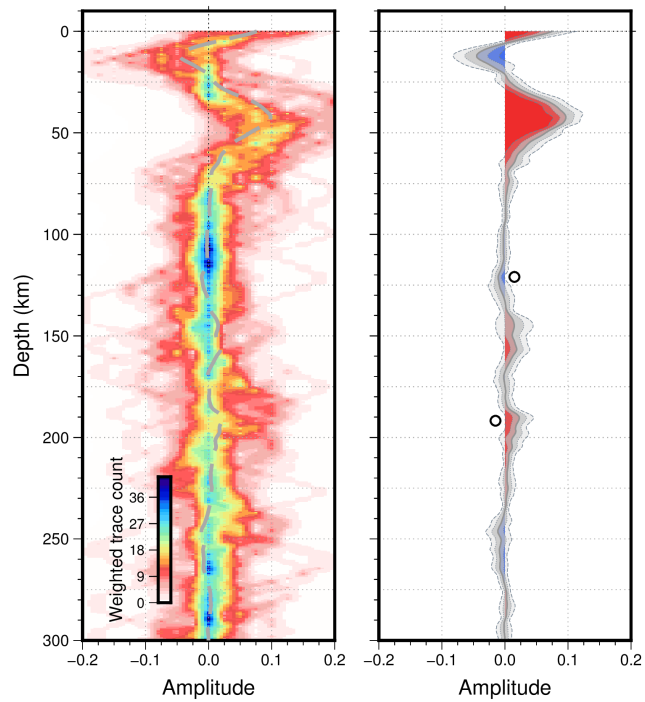


Figure S A-41. Sp RF BBOO 3

Complete Sp RF stack results for BBOO 3. Plotted like Figure 1-5.

FFC II 1 Cluster, Tot Ev 99, Rating 133

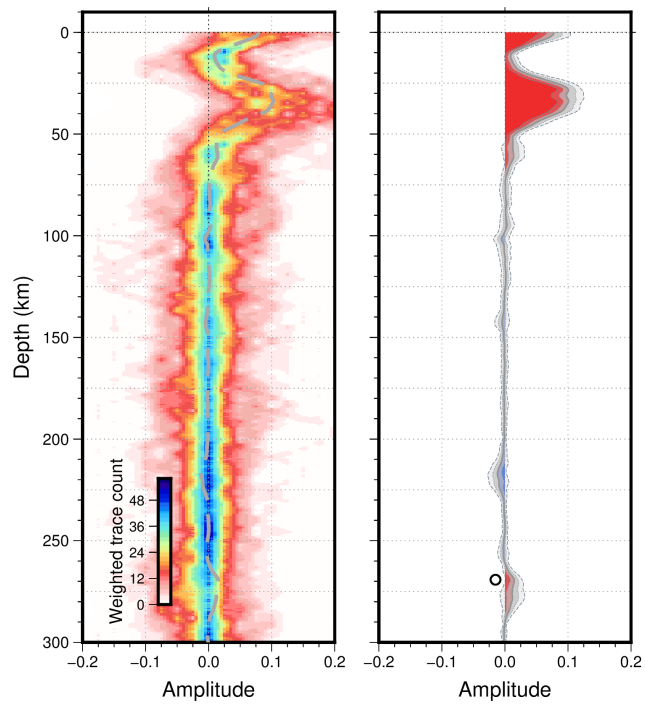


Figure S A-42. Sp RF FFC 1

Complete Sp RF stack results for FFC 1. Plotted like Figure 1-5.

DBIC GT 1 Cluster, Tot Ev 126, Rating 201

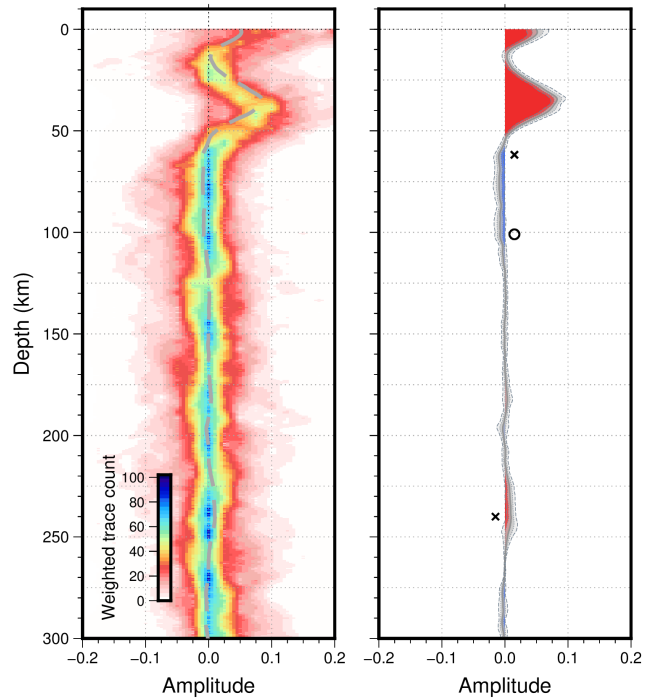


Figure S A-43. Sp RF DBIC 1

Complete Sp RF stack results for DBIC 1. Plotted like Figure 1-5.

DBIC GT 2 Cluster, Tot Ev 51, Rating 101

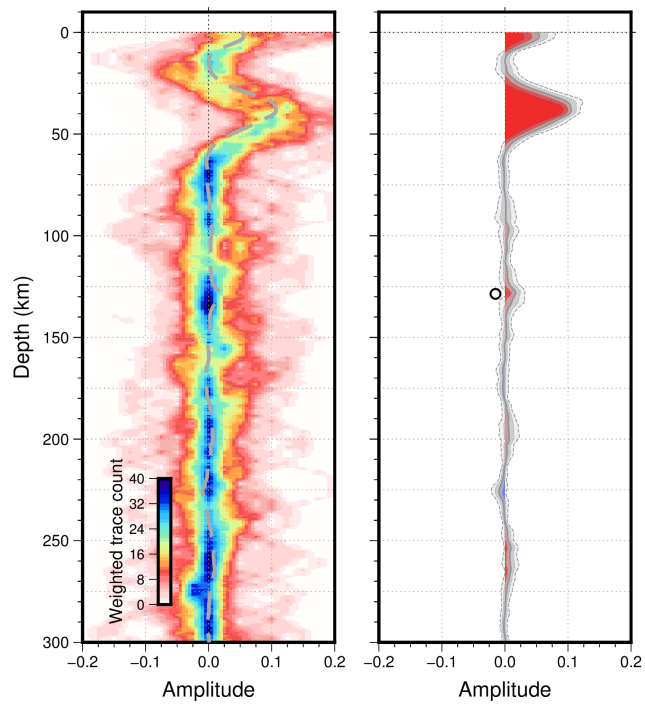


Figure S A-44. Sp RF DBIC 2

Complete Sp RF stack results for DBIC 2. Plotted like Figure 1-5.

BRVK II 1 Cluster, Tot Ev 212, Rating 329

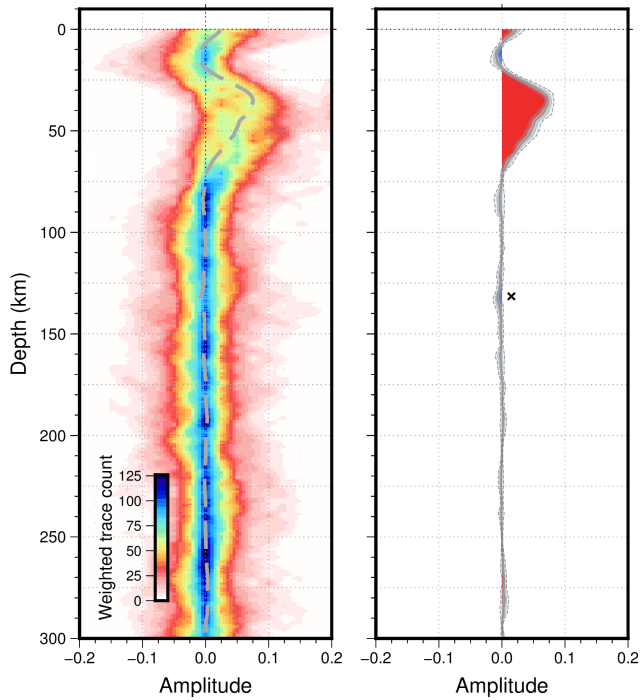


Figure S A-45. Sp RF BRVK 1

Complete Sp RF stack results for BRVK 1. Plotted like Figure 1-5.

YKWs CN 1 Cluster, Tot Ev 47, Rating 68

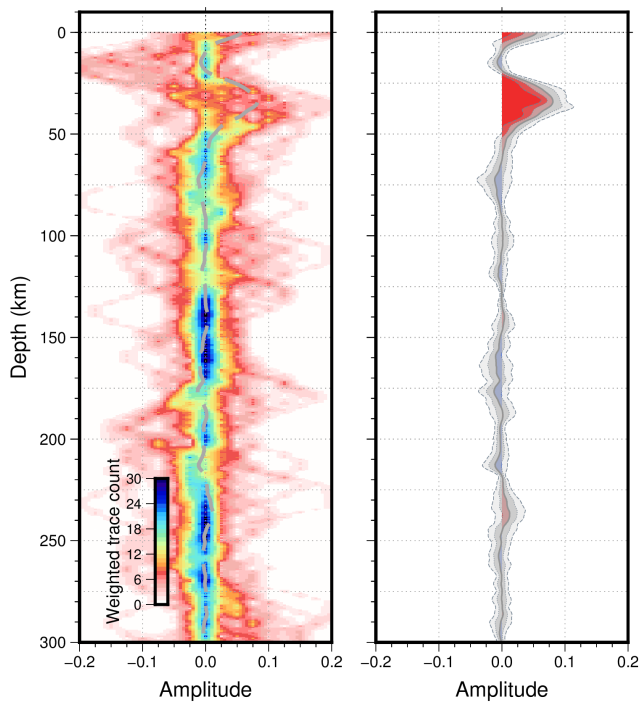


Figure S A-46. Sp RF YKWs 1

Complete Sp RF stack results for YKWs 1. Plotted like Figure 1-5.

ECSD US 2 Cluster, Tot Ev 60, Rating 123

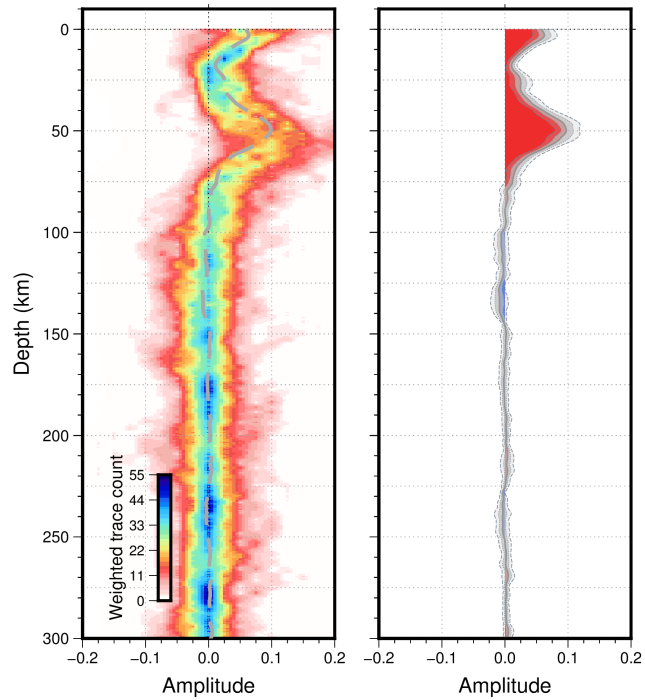


Figure S A-47. Sp RF ECSD 2

Complete Sp RF stack results for ECSD 2. Plotted like Figure 1-5.

EYMN US 1 Cluster, Tot Ev 59, Rating 99

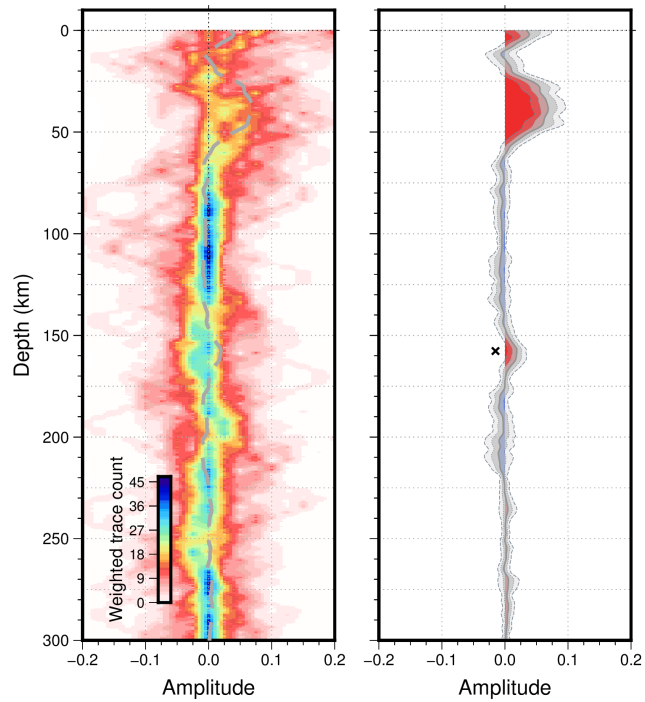


Figure S A-48. Sp RF EYMN 1

Complete Sp RF stack results for EYMN 1. Plotted like Figure 1-5.

OBN II 1 Cluster, Tot Ev 164, Rating 202

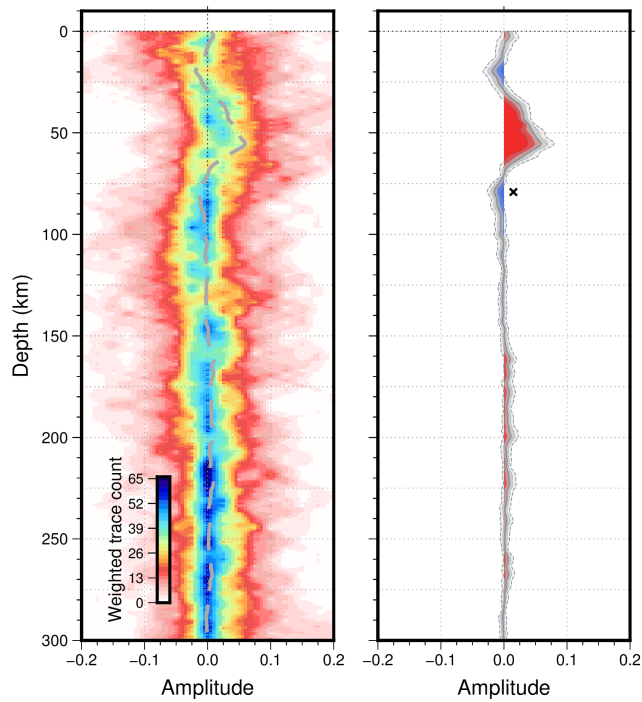


Figure S A-49. Sp RF OBN 1

Complete Sp RF stack results for OBN 1. Plotted like Figure 1-5.

JFWS US 1 Cluster, Tot Ev 63, Rating 128

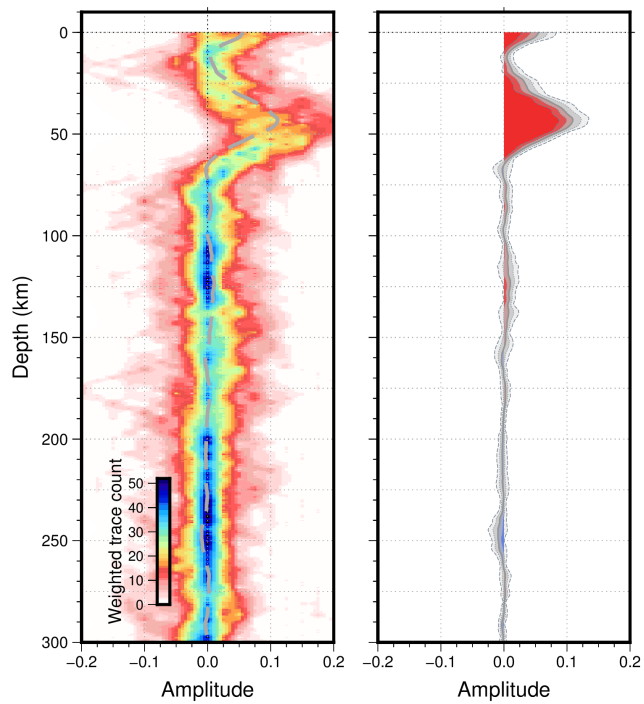


Figure S A-50. Sp RF JFWS 1

Complete Sp RF stack results for JFWS 1. Plotted like Figure 1-5.

OBN II 2 Cluster, Tot Ev 73, Rating 91

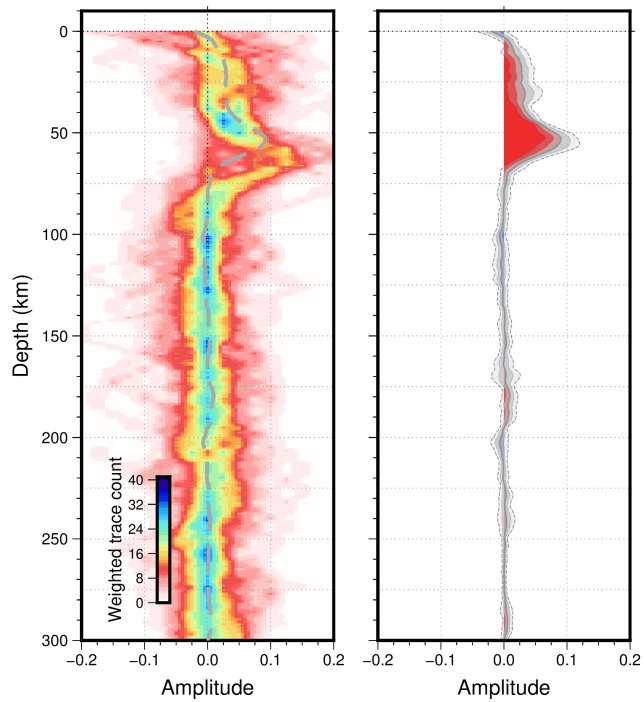


Figure S A-51. Sp RF OBN 2

Complete Sp RF stack results for OBN 2. Plotted like Figure 1-5.

K43A N4-TA 1 Cluster, Tot Ev 37, Rating 82

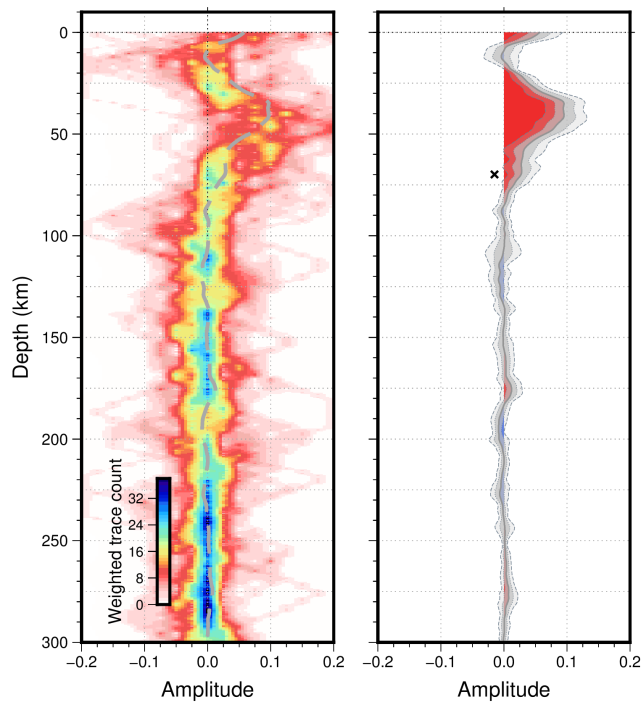


Figure S A-52. Sp RF K43A 1

Complete Sp RF stack results for K43A 1. Plotted like Figure 1-5.

AGMN US 1 Cluster, Tot Ev 58, Rating 96

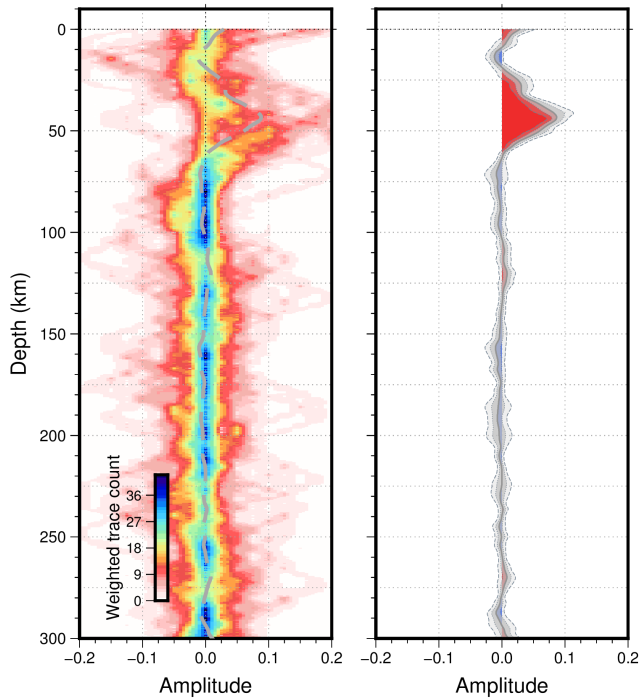


Figure S A-53. Sp RF AGMN 1

Complete Sp RF stack results for AGMN 1. Plotted like Figure 1-5.

LBTB GT 1 Cluster, Tot Ev 71, Rating 137

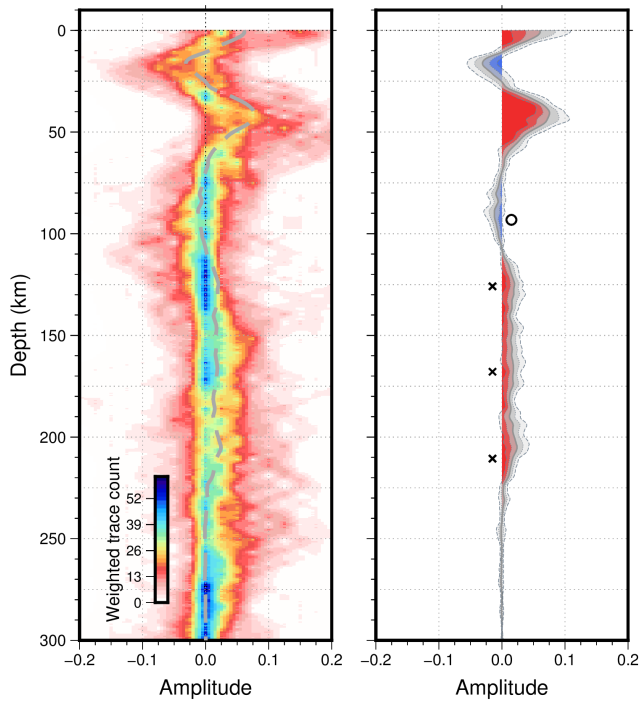


Figure S A-54. Sp RF LBTB 1

Complete Sp RF stack results for LBTB 1. Plotted like Figure 1-5.

AGO1 YT-ZM 1 Cluster, Tot Ev 78, Rating 151

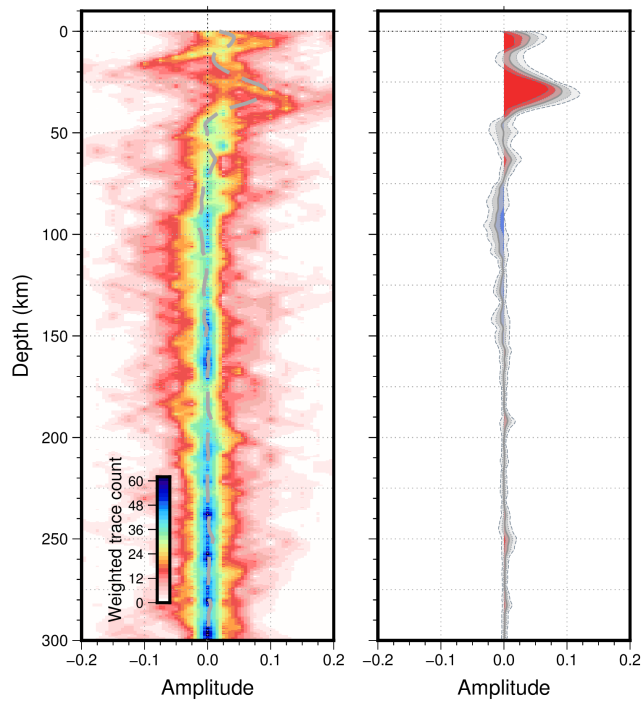


Figure S A-55. Sp RF AGO1 1

Complete Sp RF stack results for AGO1 1. Plotted like Figure 1-5.

N140 ZM-YT 1 Cluster, Tot Ev 165, Rating 272

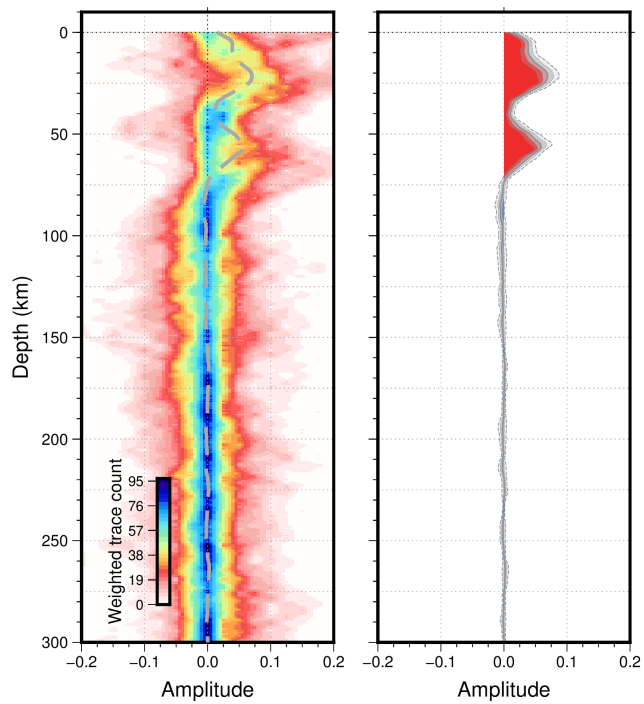


Figure S A-56. Sp RF N140 1

Complete Sp RF stack results for N140 1. Plotted like Figure 1-5.

P061 YT-ZM 1 Cluster, Tot Ev 58, Rating 100

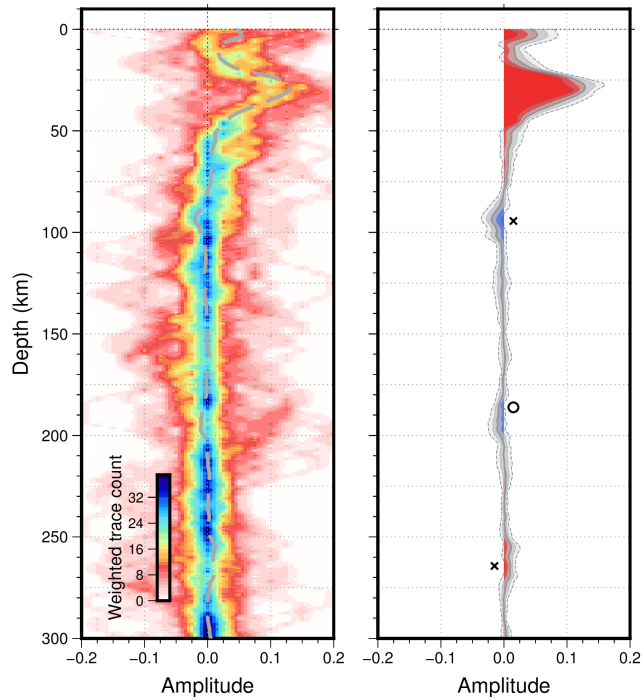


Figure S A-57. Sp RF P061 1

Complete Sp RF stack results for P061 1. Plotted like Figure 1-5.

AGO3 YT-ZM 1 Cluster, Tot Ev 59, Rating 96

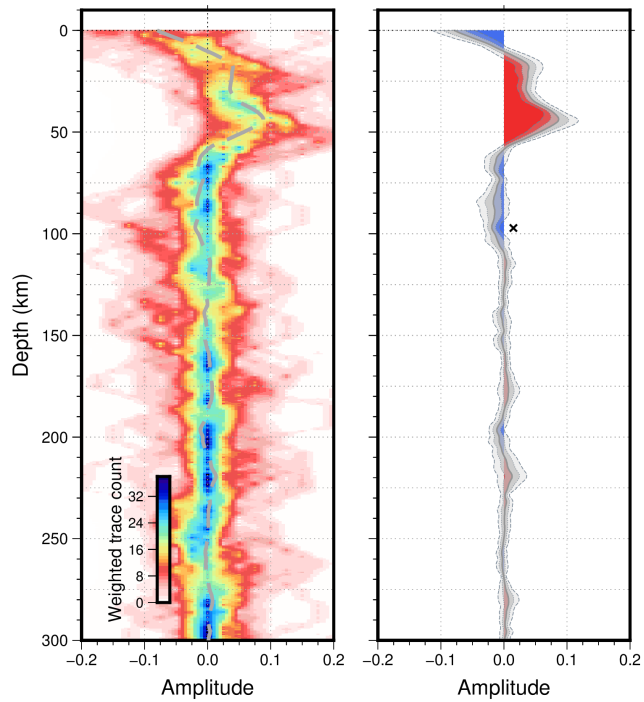


Figure S A-58. Sp RF AGO3 1

Complete Sp RF stack results for AGO3 1. Plotted like Figure 1-5.

TUC IU 2 Cluster, Tot Ev 256, Rating 489

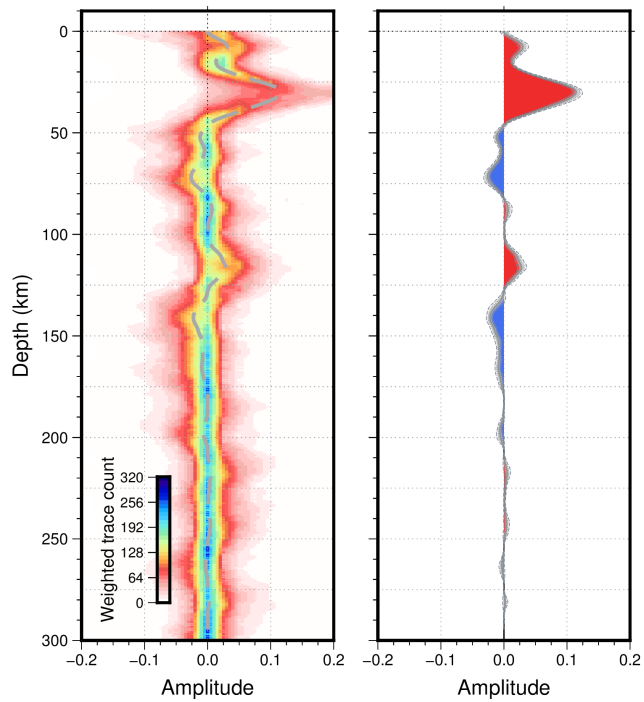


Figure S A-59. Ps RF TUC 2

Complete Ps RF stack results for TUC 2. Plotted like Figure 1-5.

TIA CB 1 Cluster, Tot Ev 282, Rating 630

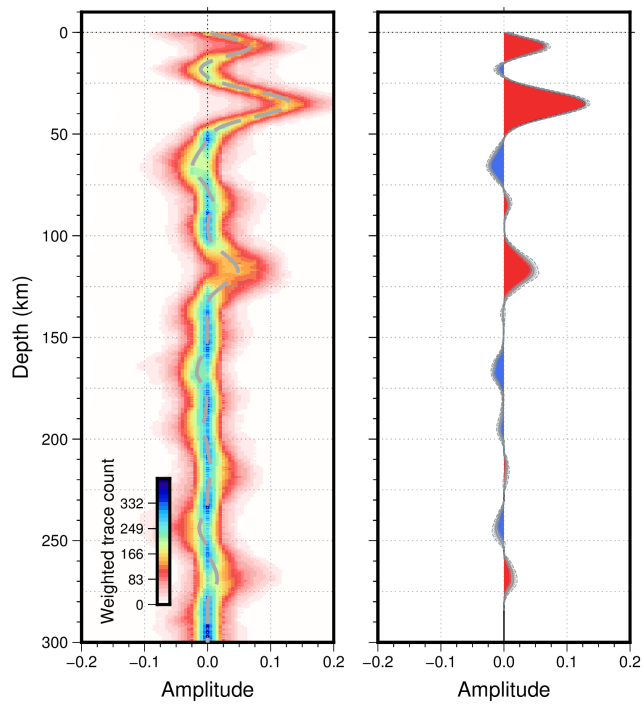


Figure S A-60. Ps RF TIA 1

Complete Ps RF stack results for TIA 1. Plotted like Figure 1-5.

LYN CB 1 Cluster, Tot Ev 303, Rating 758

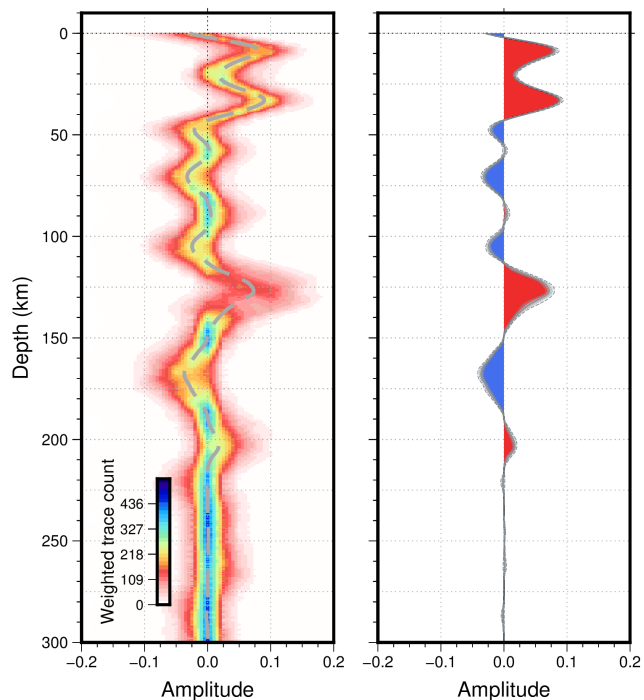


Figure S A-61. Ps RF LYN 1

Complete Ps RF stack results for LYN 1. Plotted like Figure 1-5.

RAYN II 2 Cluster, Tot Ev 210, Rating 465

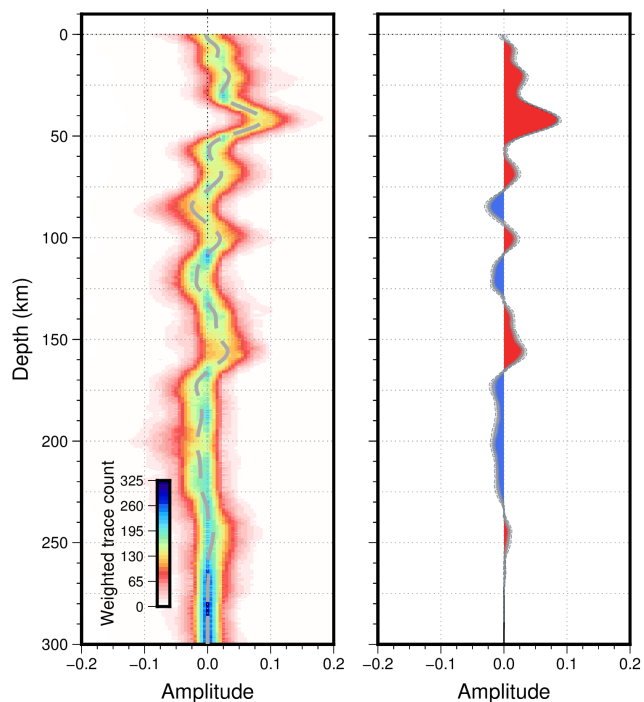


Figure S A-62. Ps RF RAYN 2

Complete Ps RF stack results for RAYN 2. Plotted like Figure 1-5.

RAYN II 1 Cluster, Tot Ev 275, Rating 441

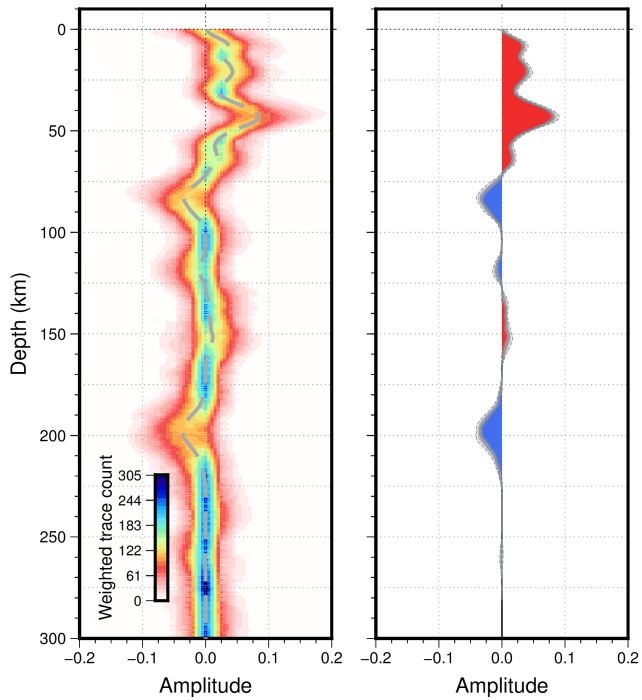


Figure S A-63. Ps RF RAYN 1

Complete Ps RF stack results for RAYN 1. Plotted like Figure 1-5.

MBAR II 1 Cluster, Tot Ev 91, Rating 149

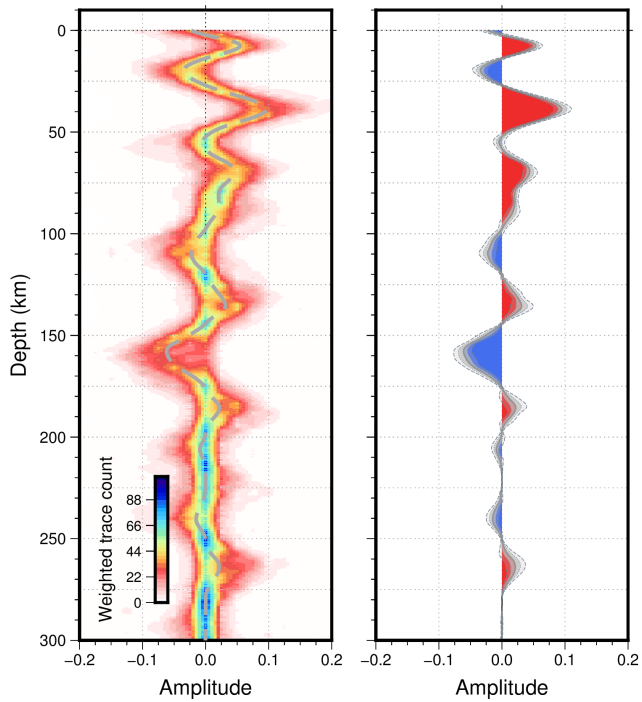


Figure S A-64. Ps RF MBAR 1

Complete Ps RF stack results for MBAR 1. Plotted like Figure 1-5.

KONO IU 1 Cluster, Tot Ev 271, Rating 438

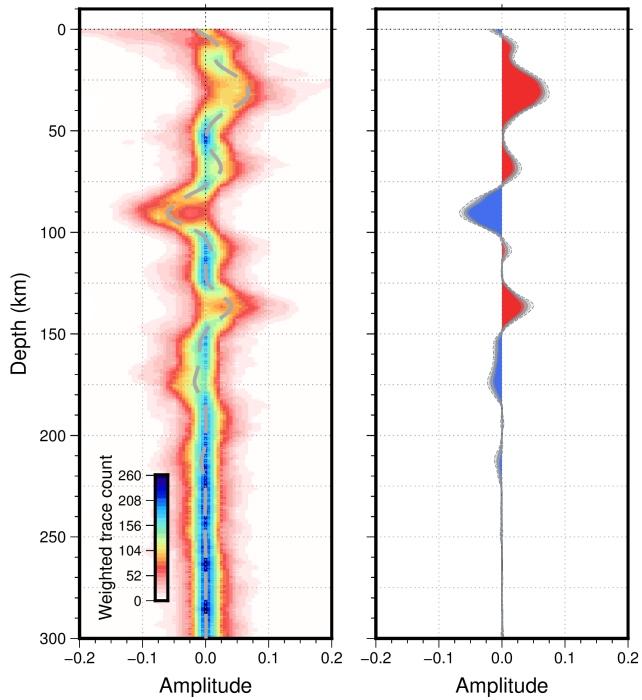


Figure S A-65. Ps RF KONO 1

Complete Ps RF stack results for KONO 1. Plotted like Figure 1-5.

LZH CB 1 Cluster, Tot Ev 288, Rating 520

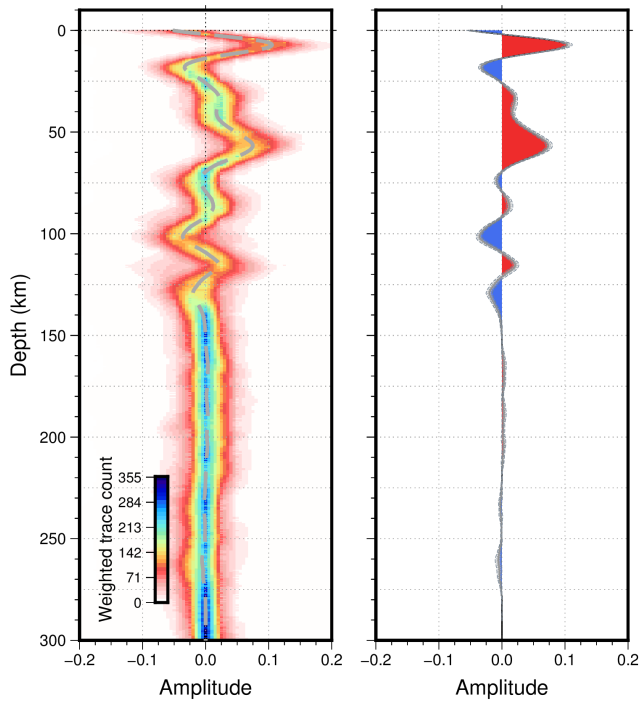


Figure S A-66. Ps RF LZH 1

Complete Ps RF stack results for LZH 1. Plotted like Figure 1-5.

HHC CB 1 Cluster, Tot Ev 304, Rating 596

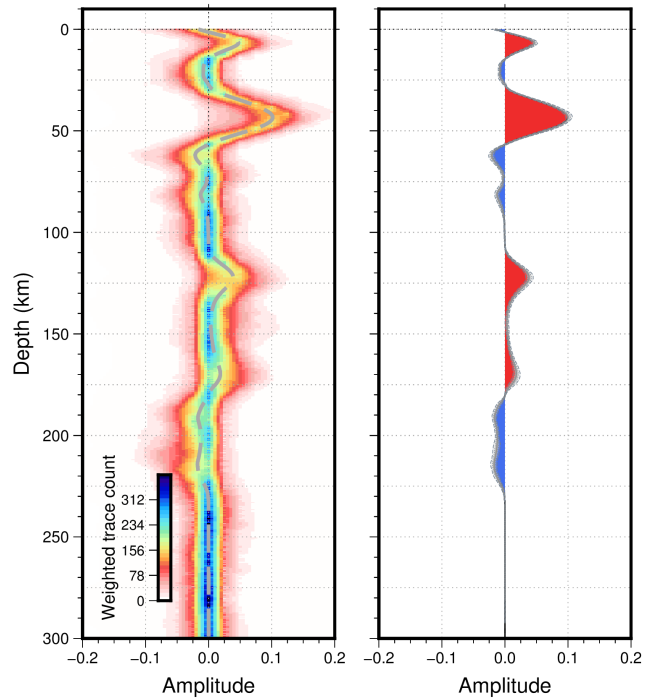


Figure S A-67. Ps RF HHC 1

Complete Ps RF stack results for HHC 1. Plotted like Figure 1-5.

SYO PS 1 Cluster, Tot Ev 107, Rating 259

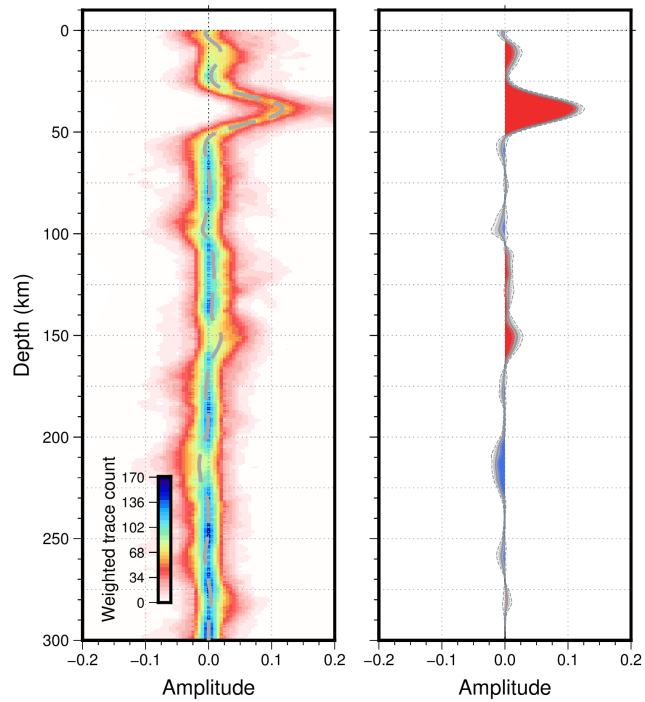


Figure S A-68. Ps RF SYO 1

Complete Ps RF stack results for SYO 1. Plotted like Figure 1-5.

KURK II 3 Cluster, Tot Ev 299, Rating 600

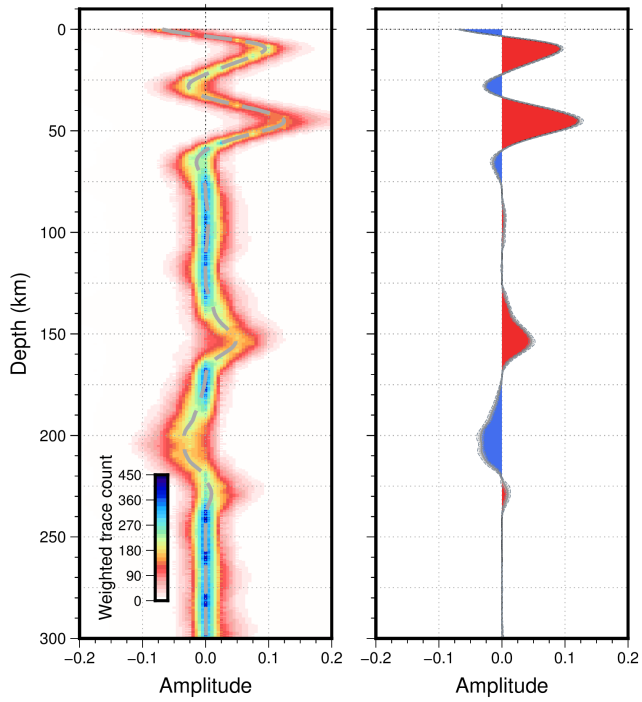


Figure S A-69. Ps RF KURK 3

Complete Ps RF stack results for KURK 3. Plotted like Figure 1-5.

KURK II 1 Cluster, Tot Ev 361, Rating 816

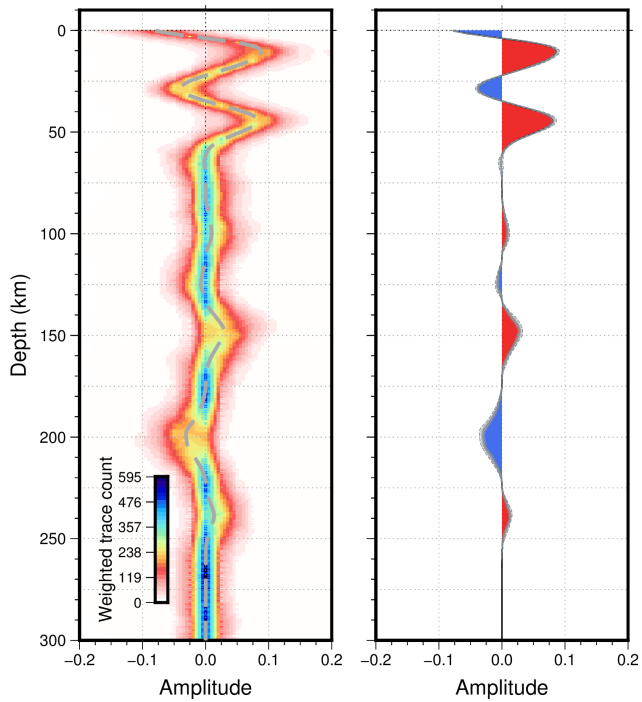


Figure S A-70. Ps RF KURK 1

Complete Ps RF stack results for KURK 1. Plotted like Figure 1-5.

RSSD IU 1 Cluster, Tot Ev 381, Rating 655

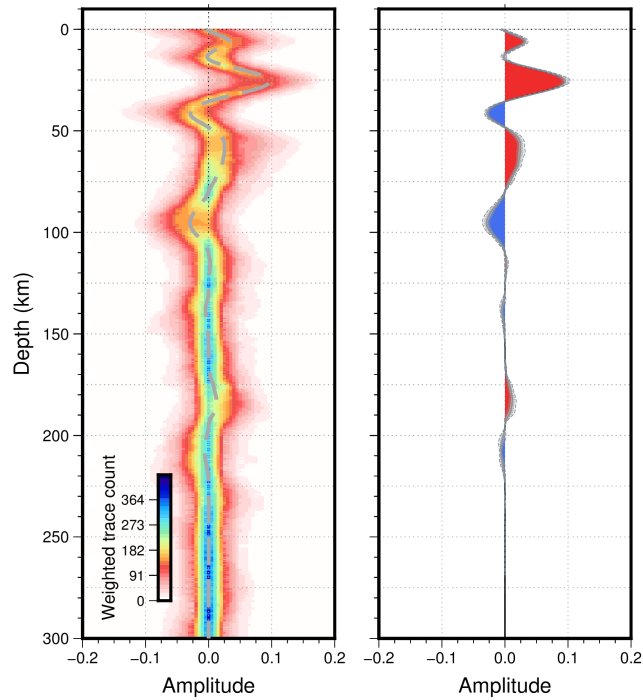


Figure S A-71. Ps RF RSSD 1

Complete Ps RF stack results for RSSD 1. Plotted like Figure 1-5.

RSSD IU 2 Cluster, Tot Ev 209, Rating 418

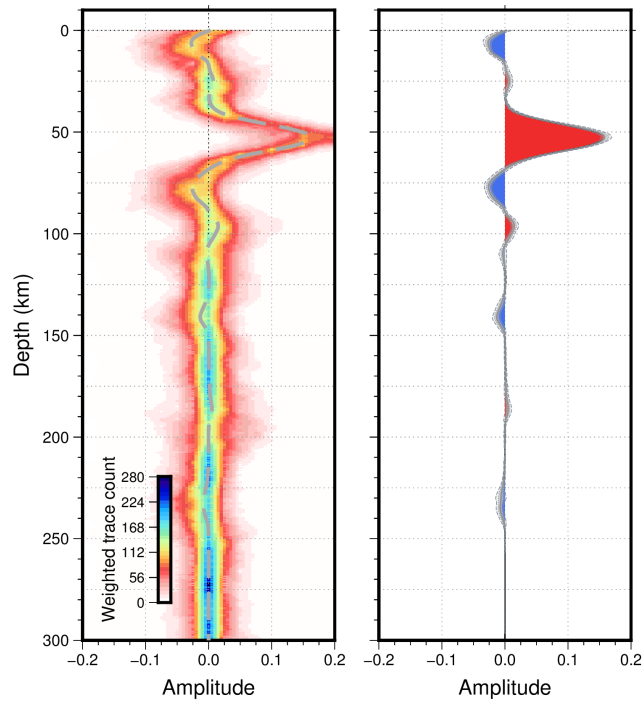


Figure S A-72. Ps RF RSSD 2

Complete Ps RF stack results for RSSD 2. Plotted like Figure 1-5.

ALE II 2 Cluster, Tot Ev 148, Rating 278

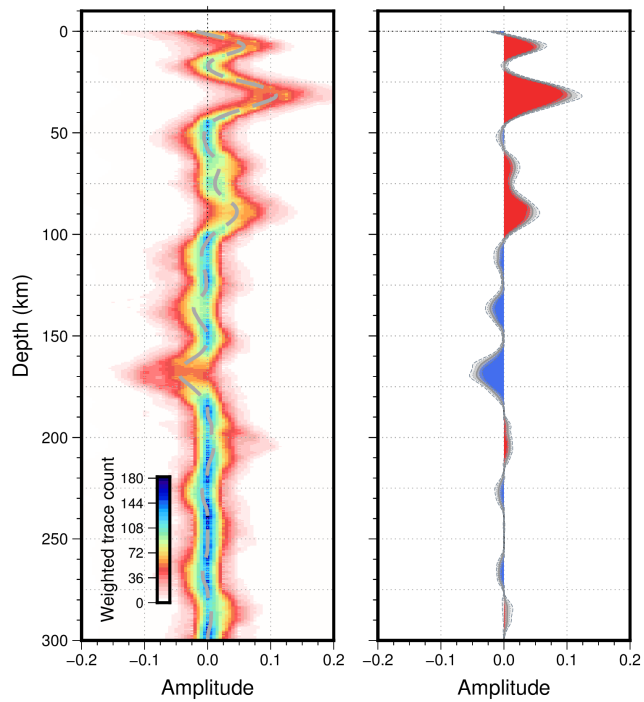


Figure S A-73. Ps RF ALE 2

Complete Ps RF stack results for ALE 2. Plotted like Figure 1-5.

HYB G 1 Cluster, Tot Ev 427, Rating 1022

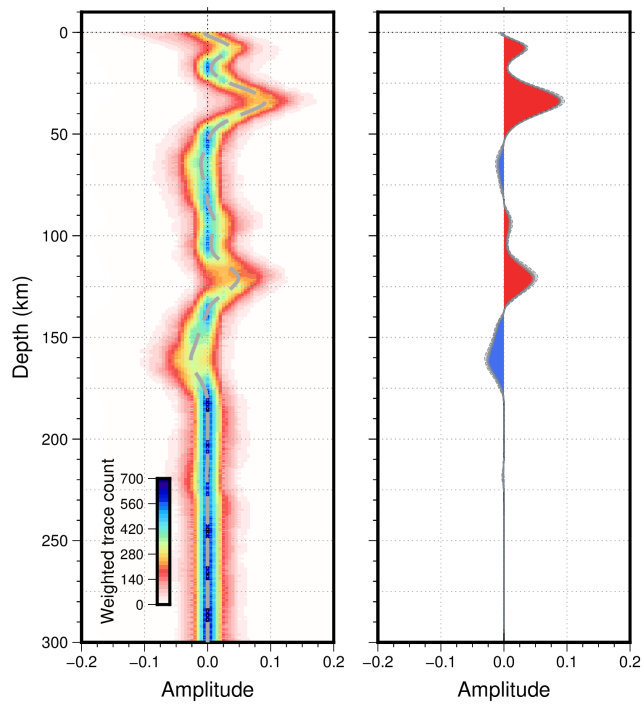


Figure S A-74. Ps RF HYB 1

Complete Ps RF stack results for HYB 1. Plotted like Figure 1-5.

STKA AU 3 Cluster, Tot Ev 80, Rating 134

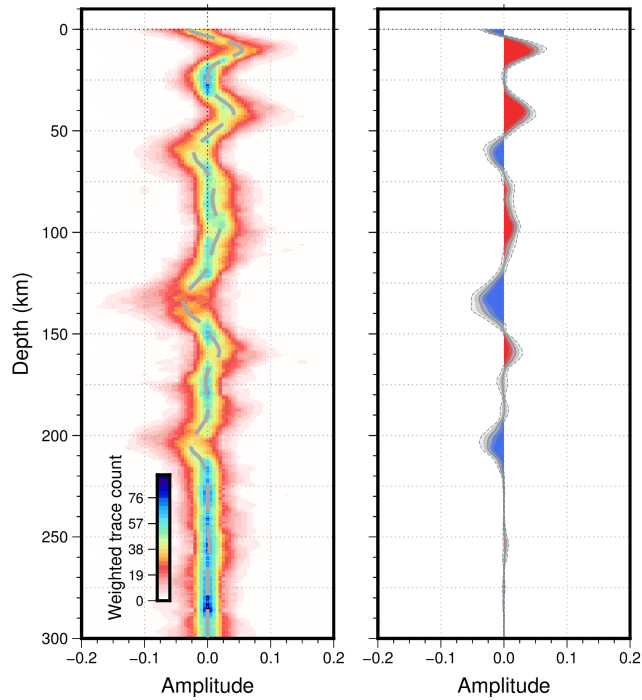


Figure S A-75. Ps RF STKA 3

Complete Ps RF stack results for STKA 3. Plotted like Figure 1-5.

NWAO IU 1 Cluster, Tot Ev 222, Rating 319

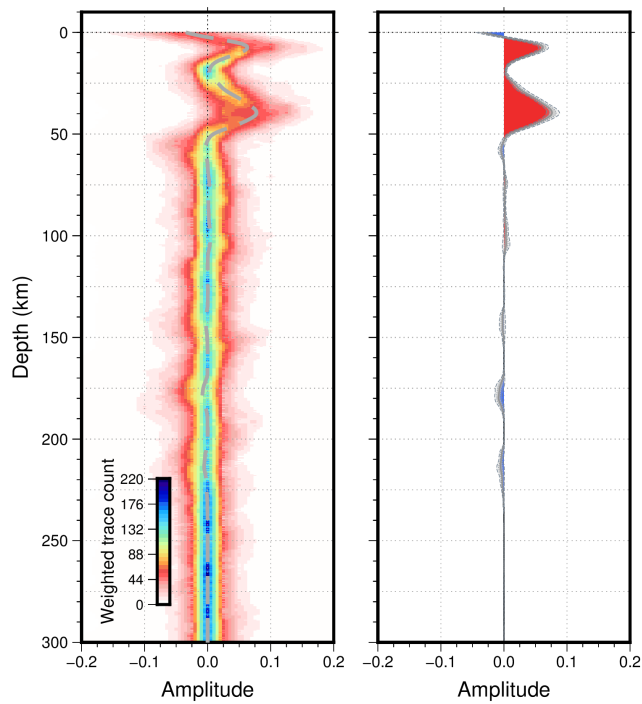


Figure S A-76. Ps RF NWAO 1

Complete Ps RF stack results for NWAO 1. Plotted like Figure 1-5.

CCM IU 2 Cluster, Tot Ev 195, Rating 439

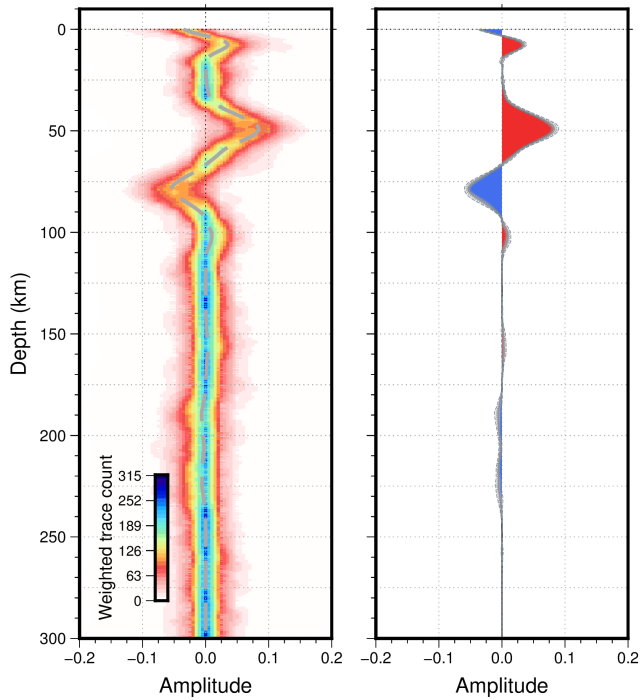


Figure S A-77. Ps RF CCM 2

Complete Ps RF stack results for CCM 2. Plotted like Figure 1-5.

HYB G 2 Cluster, Tot Ev 238, Rating 541

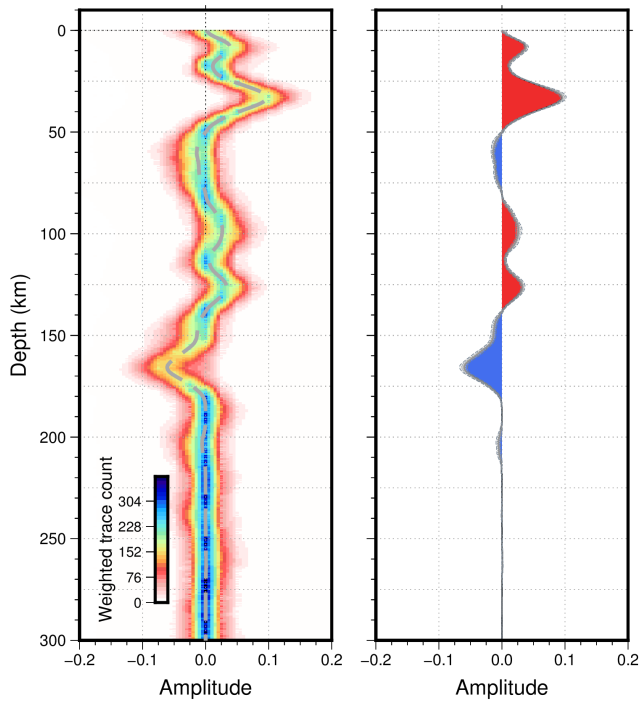


Figure S A-78. Ps RF HYB 2

Complete Ps RF stack results for HYB 2. Plotted like Figure 1-5.

TSUM IU 1 Cluster, Tot Ev 122, Rating 201

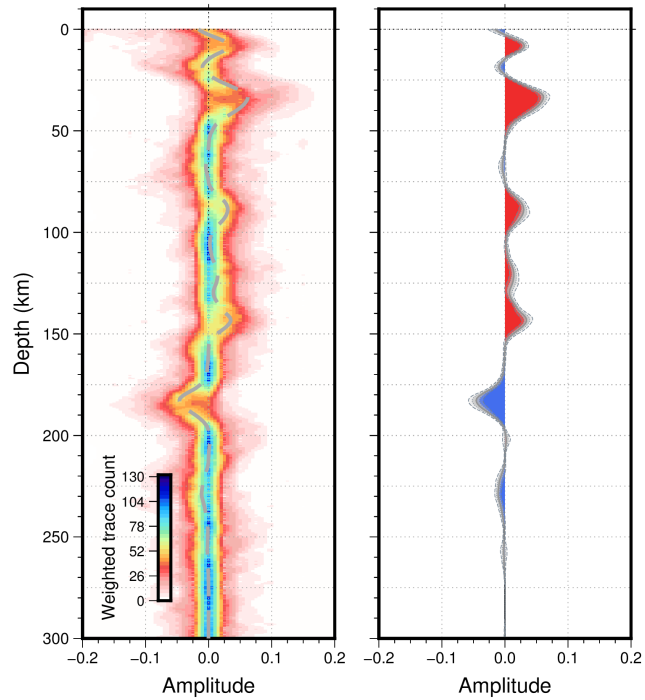


Figure S A-79. Ps RF TSUM 1

Complete Ps RF stack results for TSUM 1. Plotted like Figure 1-5.

ARU II 1 Cluster, Tot Ev 357, Rating 694

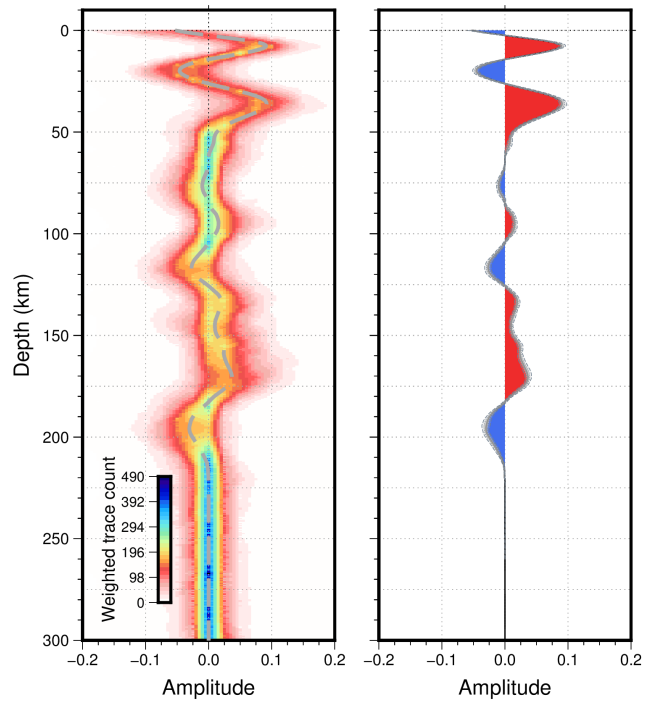


Figure S A-80. Ps RF ARU 1

Complete Ps RF stack results for ARU 1. Plotted like Figure 1-5.

CCM IU 1 Cluster, Tot Ev 229, Rating 307

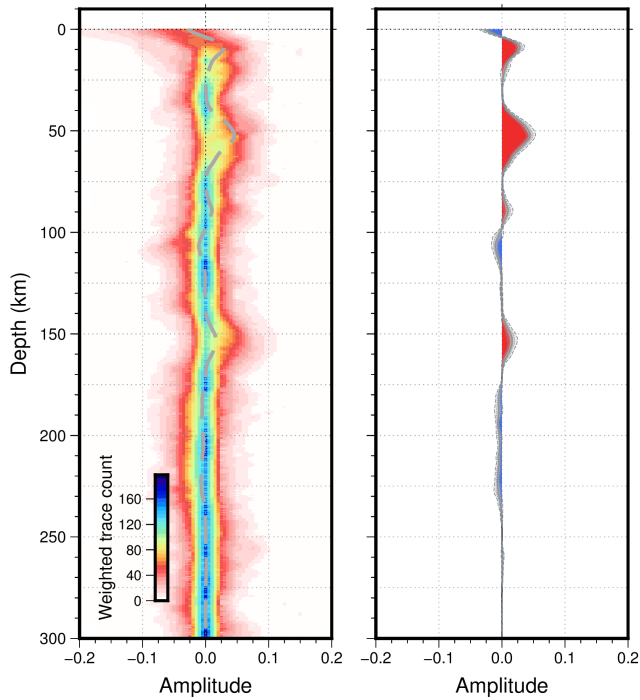


Figure S A-81. Ps RF CCM 1

Complete Ps RF stack results for CCM 1. Plotted like Figure 1-5.

LSZ IU 1 Cluster, Tot Ev 115, Rating 226

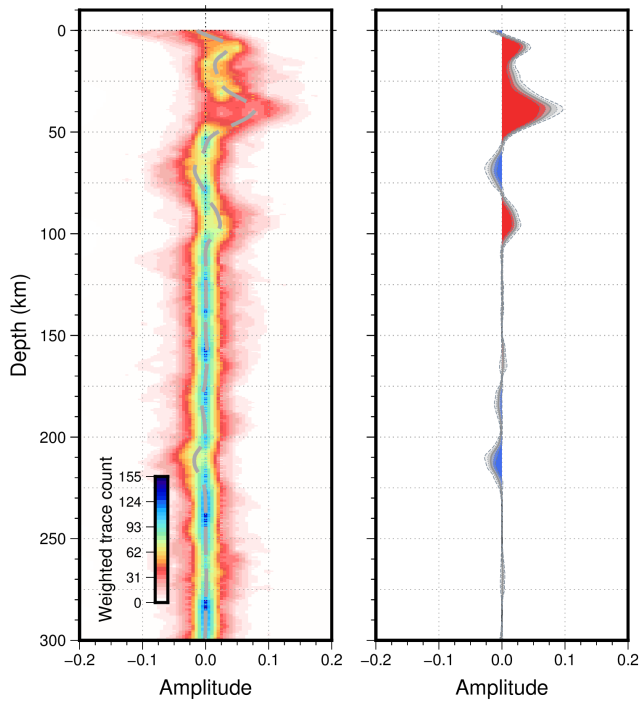


Figure S A-82. Ps RF LSZ 1

Complete Ps RF stack results for LSZ 1. Plotted like Figure 1-5.

BOSA GT 1 Cluster, Tot Ev 39, Rating 78

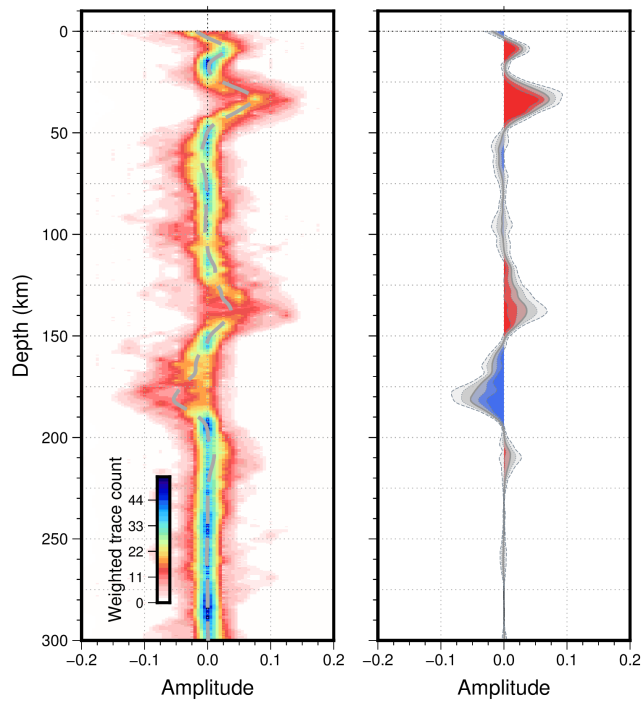


Figure S A-83. Ps RF BOSA 1

Complete Ps RF stack results for BOSA 1. Plotted like Figure 1-5.

BBOO AU 3 Cluster, Tot Ev 60, Rating 109

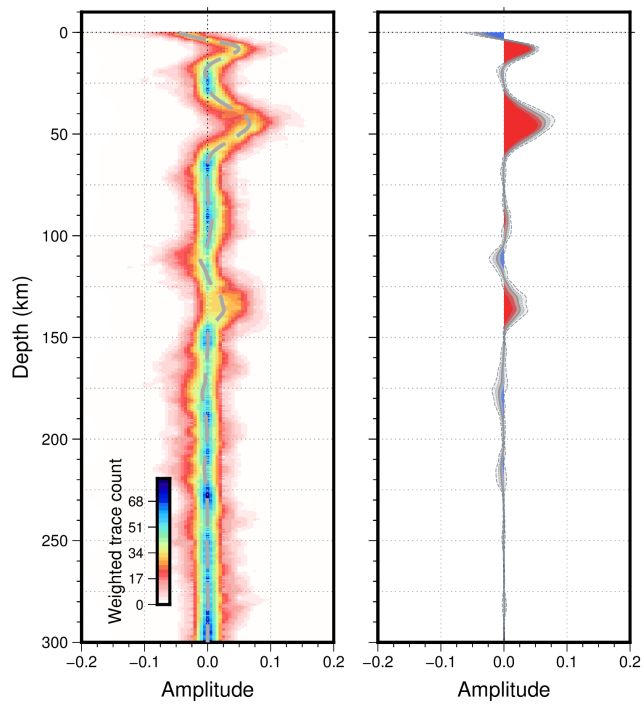


Figure S A-84. Ps RF BBOO 3

Complete Ps RF stack results for BBOO 3. Plotted like Figure 1-5.

FFC II 1 Cluster, Tot Ev 315, Rating 546

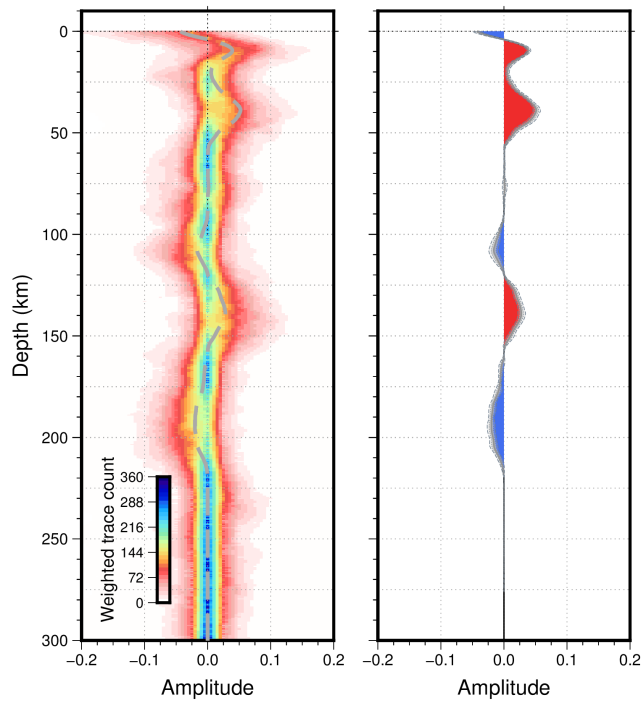


Figure S A-85. Ps RF FFC 1

Complete Ps RF stack results for FFC 1. Plotted like Figure 1-5.

DBIC GT 1 Cluster, Tot Ev 170, Rating 312

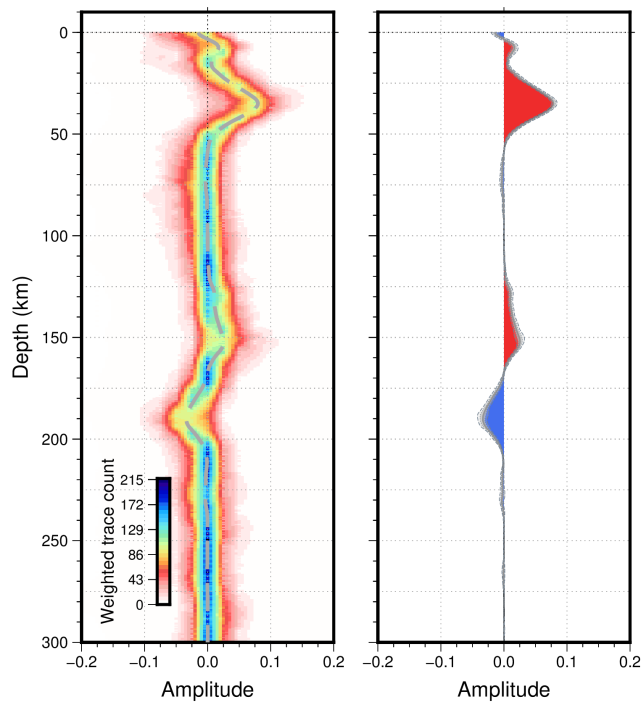


Figure S A-86. Ps RF DBIC 1

Complete Ps RF stack results for DBIC 1. Plotted like Figure 1-5.

DBIC GT 2 Cluster, Tot Ev 126, Rating 254

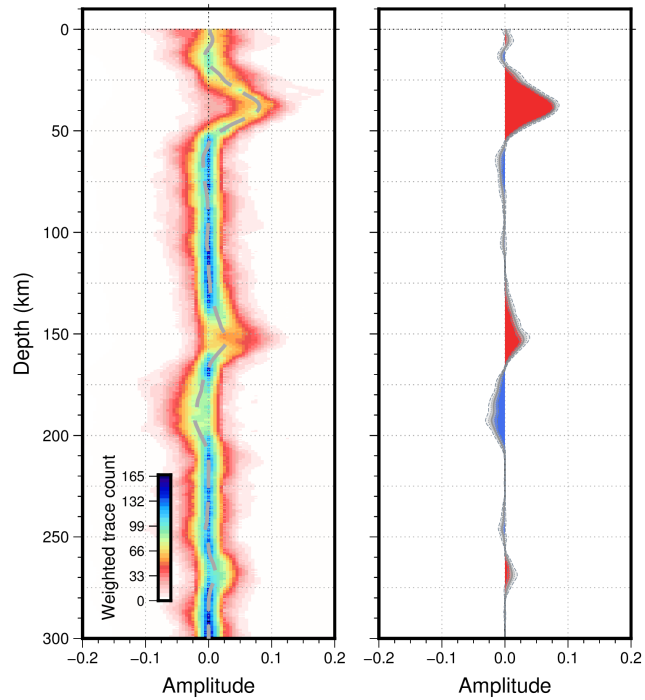


Figure S A-87. Ps RF DBIC 2

Complete Ps RF stack results for DBIC 2. Plotted like Figure 1-5.

BRVK II 1 Cluster, Tot Ev 547, Rating 641

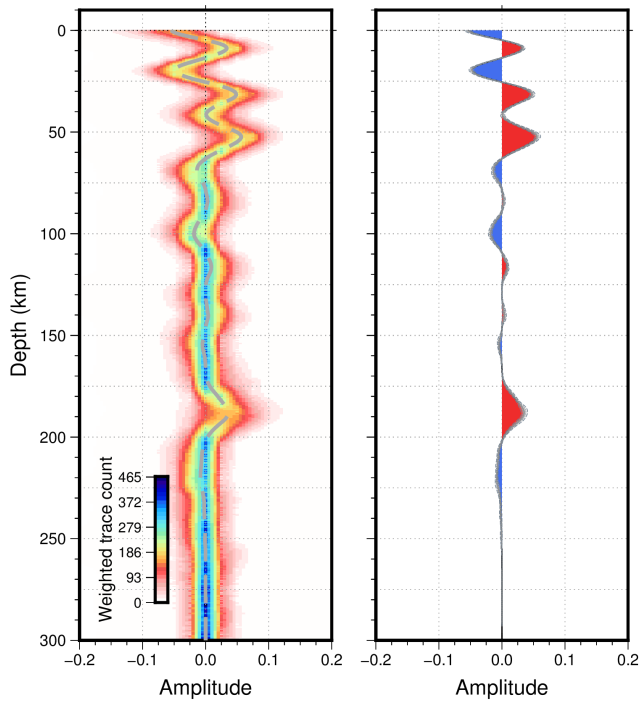


Figure S A-88. Ps RF BRVK 1

Complete Ps RF stack results for BRVK 1. Plotted like Figure 1-5.

YKWs CN 1 Cluster, Tot Ev 265, Rating 398

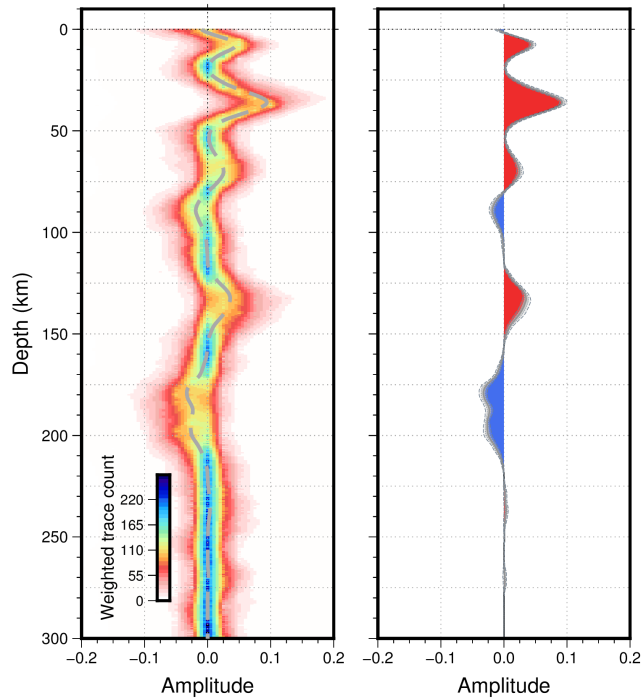


Figure S A-89. Ps RF YKWS 1

Complete Ps RF stack results for YKWs 1. Plotted like Figure 1-5.

ECSD US 2 Cluster, Tot Ev 140, Rating 292

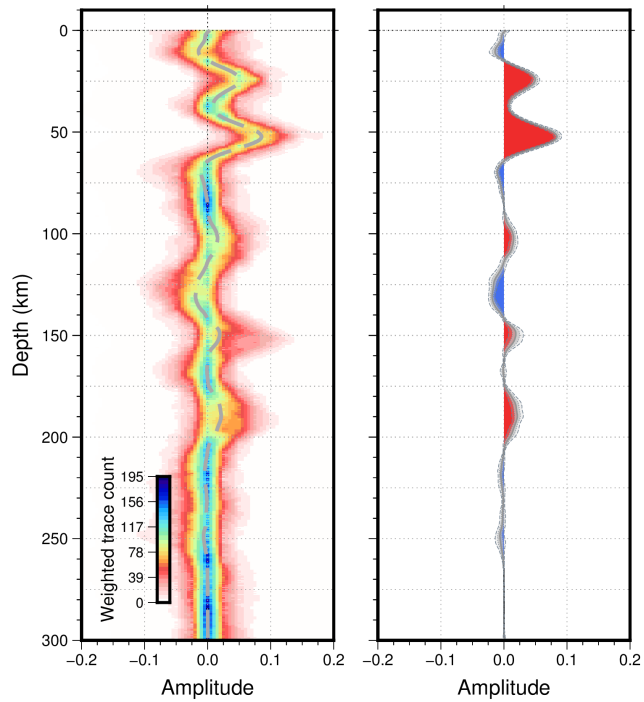


Figure S A-90. Ps RF ECSD 2

Complete Ps RF stack results for ECSD 2. Plotted like Figure 1-5.

EYMN US 1 Cluster, Tot Ev 249, Rating 275

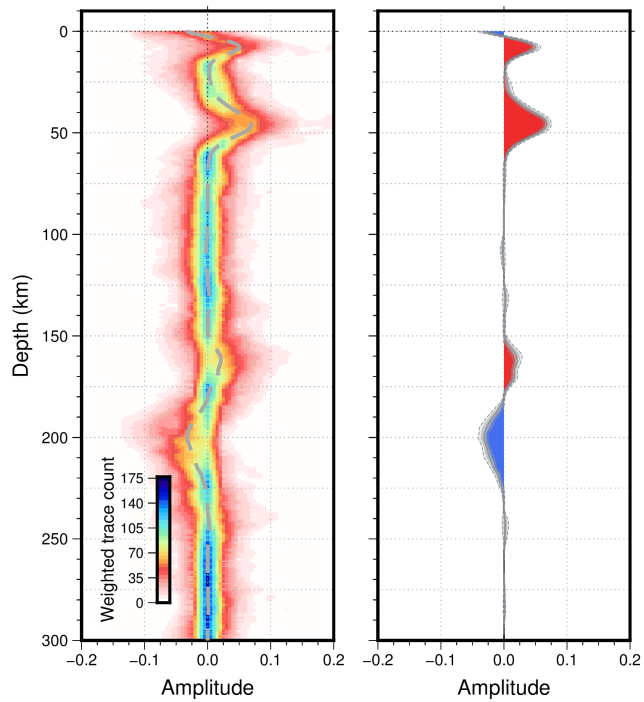


Figure S A-91. Ps RF EYMN 1

Complete Ps RF stack results for EYMN 1. Plotted like Figure 1-5.

OBN II 1 Cluster, Tot Ev 263, Rating 309

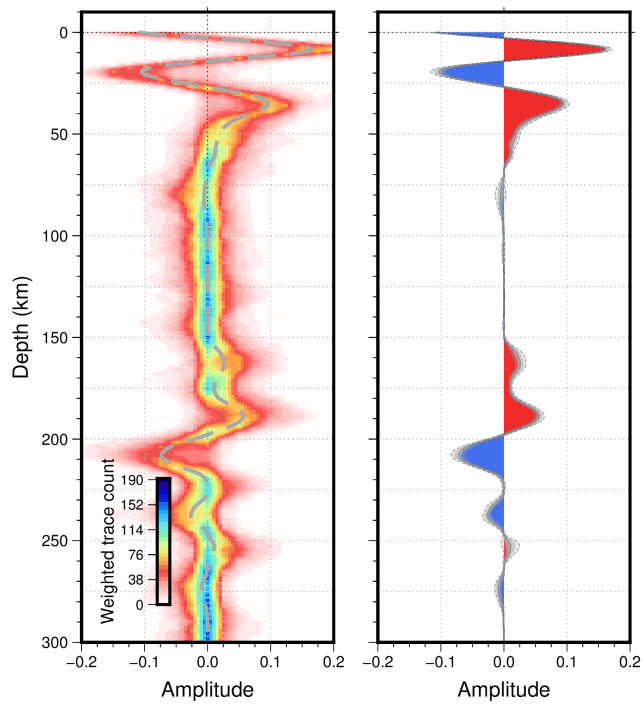


Figure S A-92. Ps RF OBN 1

Complete Ps RF stack results for OBN 1. Plotted like Figure 1-5.

JFWS US 1 Cluster, Tot Ev 180, Rating 335

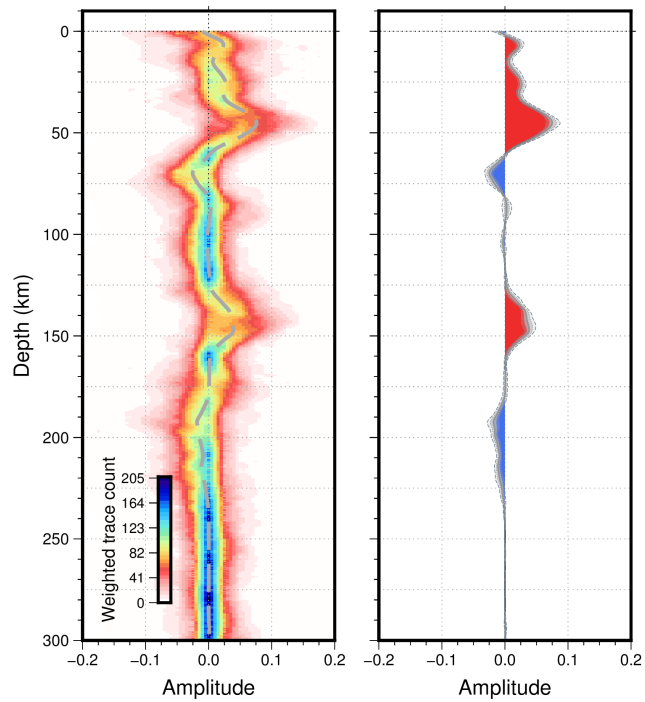


Figure S A-93. Ps RF JFWS 1

Complete Ps RF stack results for JFWS 1. Plotted like Figure 1-5.

OBN II 2 Cluster, Tot Ev 189, Rating 313

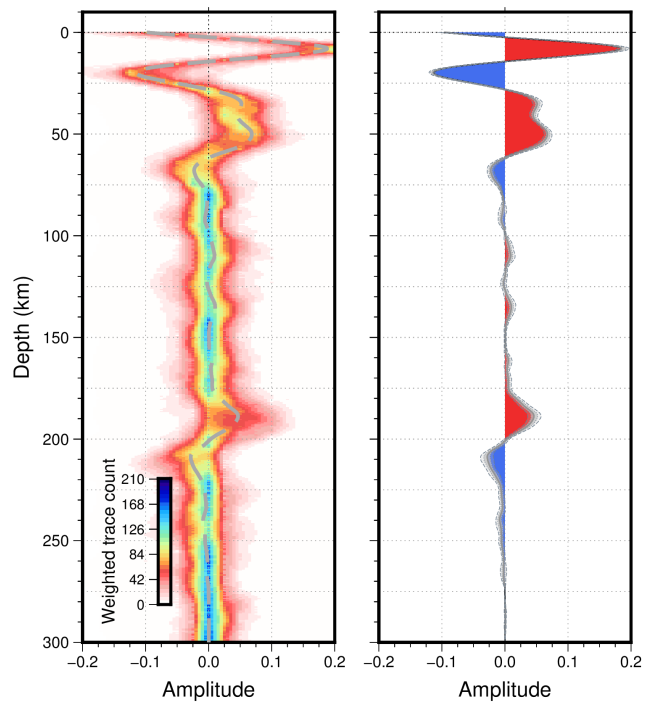


Figure S A-94. Ps RF OBN 2

Complete Ps RF stack results for OBN 2. Plotted like Figure 1-5.

K43A N4-TA 1 Cluster, Tot Ev 85, Rating 140

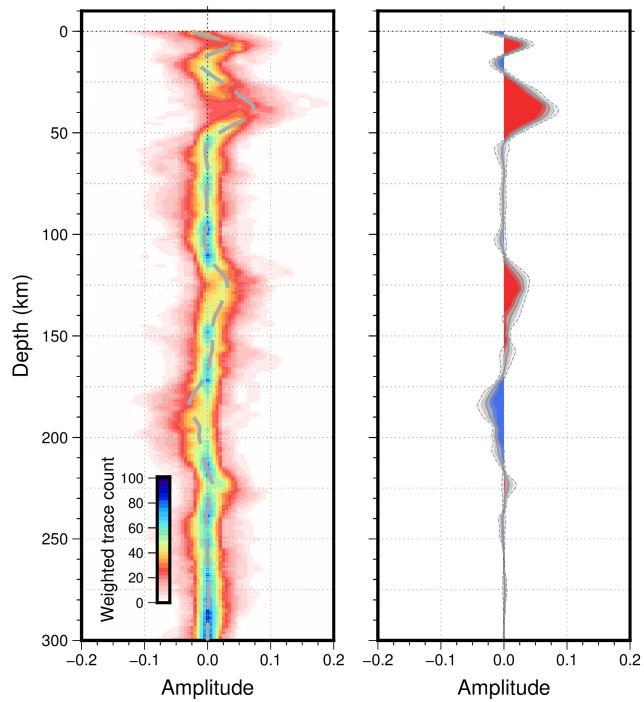


Figure S A-95. Ps RF K43A 1

Complete Ps RF stack results for K43A 1. Plotted like Figure 1-5.

AGMN US 1 Cluster, Tot Ev 147, Rating 229

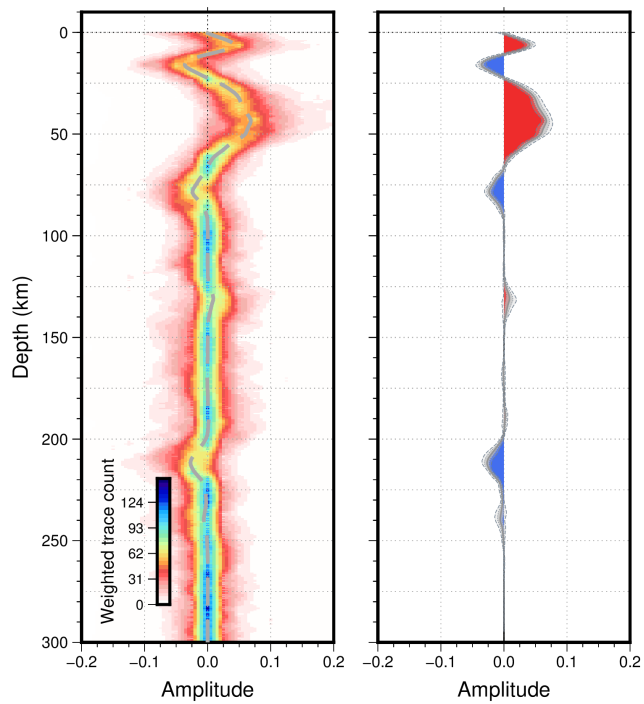


Figure S A-96. Ps RF AGMN 1

Complete Ps RF stack results for AGMN 1. Plotted like Figure 1-5.

LBTB GT 1 Cluster, Tot Ev 92, Rating 180

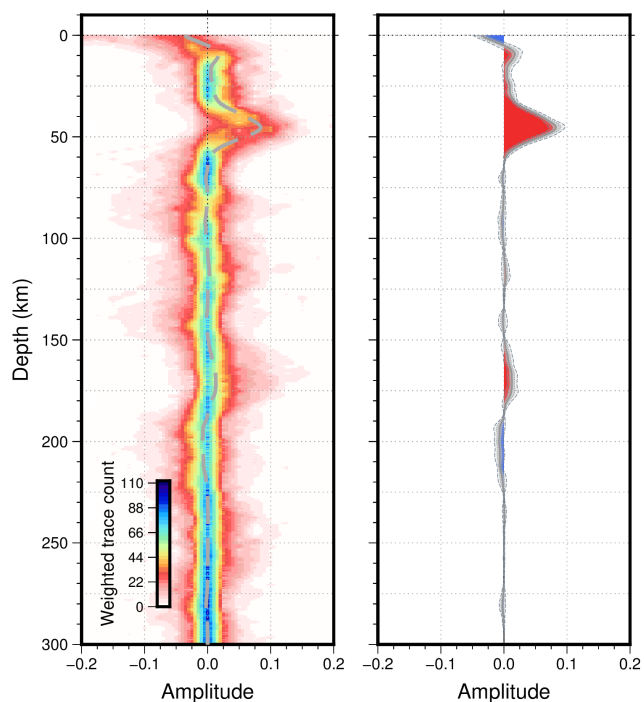


Figure S A-97. Ps RF LBTB 1

Complete Ps RF stack results for LBTB 1. Plotted like Figure 1-5.

AGO1 YT-ZM 1 Cluster, Tot Ev 77, Rating 143

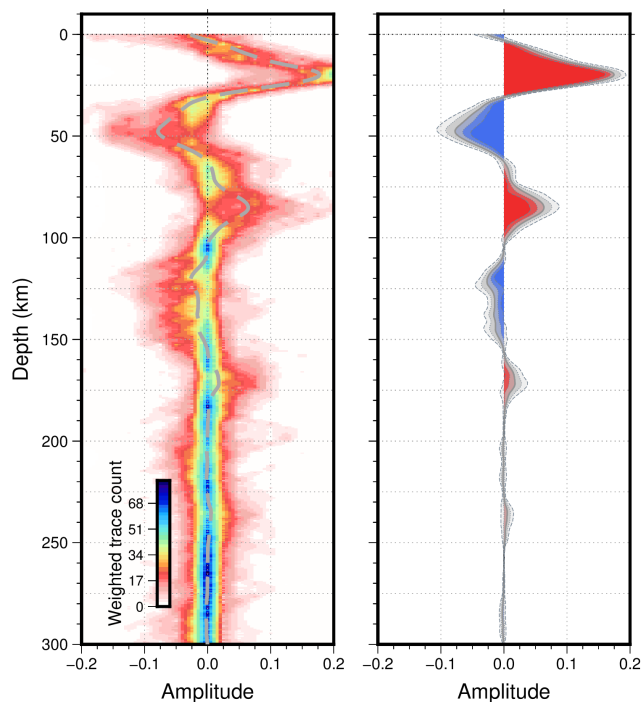


Figure S A-98. Ps RF AGO1 1

Complete Ps RF stack results for AGO1 1. Plotted like Figure 1-5.

N140 ZM-YT 1 Cluster, Tot Ev 35, Rating 61

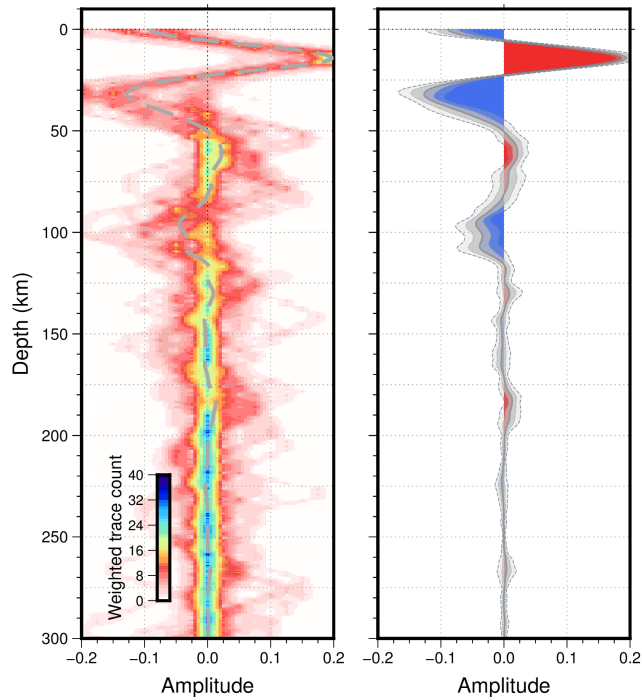


Figure S A-99. Ps RF N140 1

Complete Ps RF stack results for N140 1. Plotted like Figure 1-5.

P061 YT-ZM 1 Cluster, Tot Ev 56, Rating 120

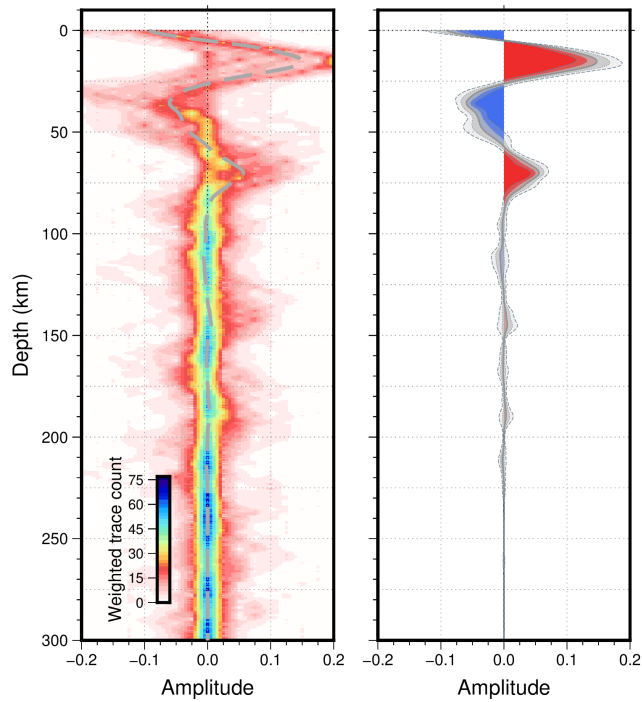


Figure S A-100. Ps RF P061 1

Complete Ps RF stack results for P061 1. Plotted like Figure 1-5.

Appendix B Supplementary Materials: A Common Origin for Layering in the Cratonic Lithosphere

S1 Supplementary Methodology

S1.1 Bayesian Joint Inversion limitations

A variety of limitations are placed on both the inversion parameters and resulting acceptable velocity profiles. (1) A monotonic increase of seismic velocity of the crust is required, while negative velocity gradients are allowed in the mantle. (2) The crust is confined to one V_p/V_s value which is solved for as a model parameter. (3) Velocity gradients at the Moho are required to be positive and cannot exceed a 30% change in velocity. (4) Strong positive velocity gradients are not allowed on either side of the Moho to isolate the Moho velocity gradient to the location of the Moho. (5) Spline knots are allowed to exist for the full depth of the model (300 km) to resolve sharp velocity gradients throughout the upper mantle, but must be spaced at least 2 km in depth from one another. (6) The V_p/V_s ratio within the crust is a model parameter, but the V_p/V_s ratio within the mantle as well as mantle and crustal density are determined using the AK135 reference model (Kennett et al., 1995), the scaling of Brocher (2005), and Abers and Hacker (2016), respectively.

S1.2 LAB & VG identification

We use Matlab's `ischange` function with a threshold of 1.4 to identify abrupt linear changes in the velocity profiles up to 300 km depth. These changes are indicated by white circles in Figure 2-4. Velocity gradients (VGs) are calculated as the percent velocity change between two inflection points. We do not consider VGs that exist completely below 250 km.

In order to discuss lithosphere structure, it is crucial we identify the bottom of the cratonic lithosphere, or the cratonic lithosphere-asthenosphere boundary (LAB), in each V_s profile. For cratonic locations, we identify LABs as the center of the largest magnitude negative velocity gradient (NVG) below 125 km depth that occurs over less than 80 km, with the additional requirement that the lower 2-sigma bound of the V_s profiles lies within the 1-MAD shear velocity range of Phanerozoic regions (Artemieva, 2006) in the SEMum2 global model (French et al., 2013) below the LAB. If no LAB is identified with this approach, we consider the LAB depth to be where the lower 1-sigma bound of the V_s profile overlaps with the upper MAD bound of the Phanerozoic V_s median. In both cases, we also required that the 1-sigma V_s values of the profile are below the upper 1-MAD upper bound for Phanerozoic mantle for 75% or more of depths from the identified LAB to 250 km. This latter condition avoids placing the LABs above significant PVGs and/or high shear velocities that are unlikely to represent the asthenosphere.

S2 Supplementary Results

S2.1 Shear Velocity Profiles with Regional Surface Wave Data

S2.1.1 United States' cratonic lithosphere

We invert for V_s us Sp RFs and regional Rayleigh wave phase velocities in the United States (Babikoff & Dalton, 2019; Herrmann et al., 2021) for stations RSSD, ECSD, CCM, JFWS, and K43A. In the United States, the use of regional Rayleigh wave phase velocities do not seem to systematically improve V_s profile results. This is likely because global models of shear velocity tend to be better resolved for well-instrumented continents such as North America.

Exceptionally, at CCM1 and RSSD1 the median absolute velocity in the lithosphere is shifted significantly, even changing the depth identified as the LAB by ~ 100 km in both cases. These

two profiles exist at the edge of the Wyoming craton which is suggested to be delaminating (RSSD) and on the boundary of the Mazatzal and Granite-Rhyolite provinces (Wirth & Long, 2014). The significant change in absolute Vs observed when comparing inversions using global and regional Rayleigh phase velocities at the station location indicates the importance of regional Rayleigh wave phase velocity models to capture lateral variability in mantle structure.

S2.1.2 Antarctic cratonic lithosphere

We jointly invert for Vs using Sp RFs and regional Rayleigh phase velocities from (Shen, Wiens, Stern, et al., 2018) at stations P061, AGO1, AGO3, and N140. In general, the regional surface wave data inversion in Antarctica appear to be similar to the inversion performed with GDM52, but with narrower PDFs. These narrower PDFs indicate a higher confidence in the median Vs profile. In some Antarctic receiver functions, there are PVGs above 75 km that the inversion methodology had difficulty reconciling. This is the most obvious when comparing the global and regional Vs profiles for N1401 (Fig. 2-4d). The two separate inversion results identify the two distinct PVGs in the Sp RF separately as either a shallow Moho followed by a strong PVG or deep Moho followed by only a minor PVG. In the Sp RF for AGO31, there are also two shallow large PVGs, but the inversion results are more consistent for this location.

S2.1.3 South African cratonic lithosphere

Vs profiles calculated with global and regional surface wave data exhibit similar shapes and absolute velocities (Malory et al., 2022; White-Gaynor et al., 2021). The regional surface wave data for LBTB1 allows for the improved fit of the PVGs indicated by the Sp RF at LBTB1 (Krueger et al., 2021) and, subsequently, significantly increases the absolute velocities and the VG in the depth range 100-250 km. The combined PVG increases from 3.5% to 8.9% and the

maximum velocity in the deep lithosphere increases ~ 0.25 km/s. The regional inversion of BOSA1 also exhibits higher mid-lithosphere velocities than the inversion that uses GDM52.

S3 Supplementary Discussion

S3.1 Regional variations in the cratonic lithosphere

We find lithospheric seismic structure to be extremely heterogeneous. This heterogeneity is obvious at the global scale but is also prevalent at the regional scale, and even a scale of ~ 100 km or less. Here we discuss our results by region.

3.1.1 North America

In North America, we are unable to model the compositions of 5 of the 12 inversion profiles located in this region due to high velocities or complicated layering structure not represented in the range of models we test. Of the seven profiles we do model, four contain velocity increases that can be modeled with diamond presence, while three (RSSD2, ECSD2, CCM2) can be modeled by peridotite only compositions, although we believe that the structure of ECSD2 may be underfit. This variation in mantle structure throughout the continent is consistent with previous work (Altoe et al., 2020; Liu & Shearer, 2021).

Inversion profiles RSSD2 and RSSD1 differ significantly. RSSD1 displays significant positive and negative velocity gradients within the upper mantle, while the lithosphere of RSSD2 has no significant velocity gradients. This result is consistent with the NVGs observed by Hopper 2018 (Krueger et al. (2021) Figure SB-14). Notably, station station RSSD is located on the easterd edge of the Wyoming craton which has experienced significant alteration since the Precambrian. Episodes of metasomatism and delamination of the mantle lithosphere at the western edge of the craton have likely resulted in alteration of lithosphere velocity structure with significant lateral variation (Carlson et al., 2004; Eilon et al., 2018). This is evidenced in the

change in mid-lithosphere absolute velocities below 100 km between inversions of RSSD1 using GDM52 and regional Rayleigh phase velocities.

Another set of profiles that show significant local variability are CCM1 and CCM2. CCM1 is at the boundary of the Mazatzal and Granite-Rhyolite provinces (Wirth & Long, 2014) which is north of the Reelfoot Rift with CCM2 sampling southward towards the Reelfoot Rift, and CCM1 sampling northward, away from the rift into the Mazatzal. The most obvious variation between these two profiles is a 30km difference in LAB depth, with the shallower LAB in the profile sampling closer to the Reelfoot rift (Liu & Shearer, 2021). Liu and Shearer (2021) observe significant negative velocity gradients in the upper mantle near the Reelfoot Rift and Wirth and Long (2014) observe multiple anisotropic NVGs at CCM. While the Sp RFs we use for the inversion do exhibit substantial upper mantle NVGs, we do not fit any low velocity layers in our compositional models at CCM. It is possible CCM2 is being underfit due to its relatively broad 1-sigma bounds and CCM1 is not modeled because it's high uppermost mantle velocities cannot be represented by the range of compositions we consider. It is also possible that the profile of CCM1 has an LAB identified too shallowly due to absolute velocities from GDM52. The regional profile for CCM1 has higher velocities below 100km than the global profile.

K43A, JFWS1, and ECSD2 sample the lithosphere near the mid-continent rift. These profiles exhibit relatively high upper mantle velocities with no significant negative velocity gradients. Interestingly, this is contrary to the results of *Liu and Shearer* [2021], who observe negative velocity gradients at the mid-continent rift. Their general observation of two common negative polarity reflections in the lithosphere do correspond with velocity gradients in inversion profiles AGMN1, EYMN1, RSSD2, and CCM1 (Liu & Shearer, 2021).

Chen et al. (2021) observes seismic anisotropy in the Superior craton, although their observations are northwest of the locations we observe. They see positive and negative impedance contrasts within the lithosphere of the Superior craton which corresponds with our results at EYMN1 and AGMN1. Specifically, the observation of negative impedance contrasts at ~100 km coincides with our results.

We sample the edge slave craton with at YKWs1 where we fit compositions of a low velocity layer above 80km and a velocity increase at 110km. The depths of the layering we observe is inconsistent with the MT anomaly from 80-120km observed by Chen et al. (2009). Although, Bostock (1998) does observe complicated layering structures in this region.

3.1.2 East Europe

In eastern Europe (and easternmost central Eurasia) we model four profiles that require velocity increases that can be modeled with diamond (KONO1, OBN1, OBN2 KURK1), although the structures of KONO1 and KURK1 could also be represented by a low velocity layer in mantle with a lower geotherm. KONO1 can be fit by either a velocity increase or a wide low velocity layer. This profile samples the Oslo rift, which may indicate that a wide low velocity layer due to metasomatism from fluid infiltration is more likely (Bruneton et al., 2004; Neumann et al., 1992).

One profile is explained by peridotite only but has clearly had significantly thinned lithosphere (KURK3) and indicate steep LAB topography in this region as we identify the LAB at KURK1 to be >250km. Profiles ARU1 and BRVK1 are not explained by the limited scope of our compositional models.

3.1.3 Africa

Within the African continent we consider the lithospheric structure in four distinct cratons, the West African craton (DBIC), Arabian Shield (RAYN), Tanzanian craton (MBAR1), and

Kalahari craton (southern Africa). In Archean cores, DBIC and BOSA1 lithosphere structure can be explained by the presence of diamond. LBTB1 also exists in an Archean core, but is not modeled with diamond, although it may be underfit (see section 4.4). TSUM1 and MBAR1 are modeled with peridotite only and exhibit thinned lithosphere, although they may be better fit with a low velocity layer in the uppermost mantle. Throughout the continent, the results of Lui & Klemperer are remarkably consistent with the inversion profiles we calculate: ~100km NVG near LBTB; ~80km NVG near TSUM; ~70km NVG at DBIC; and no shallow NVGs at RAYN.

In the Kaapvaal craton, we observe high velocities in the lower lithosphere in LBTB1 and BOSA1, which are pronounced in inversion profiles from regional Rayleigh phase velocities. The Kaapvaal craton is considered to be geochemically complex (Hoal, 2003; Stachel & Harris, 2008). Significant Paleoproterozoic volcanism, which emplaced the Bushveld Igneous Complex, likely modified the cratonic lithosphere significantly with melt rich in silica (Hoal, 2003; Kgaswane et al., 2012). Additionally, these profiles are modeled to have low velocity layers in the upper lithosphere, although the low velocity layer at LBTB1 is more distinct in appears, with a 4% velocity gradient, and requires a higher velocity drop to model. Paulssen et al. (2022) propose that rifting related to the western arm of the East African Rift system continues through Botswana where upward migrating fluids are indicated based on earthquake occurrences. This rifting could explain the distinct low velocity layer observed at LBTB1.

We also sample the mantle structure at the eastern edge of the Tanzanian craton with MBAR1. This location is surrounded by the East Africa Rift. Celli et al. (2020) observe a cratonic keel below MBAR that extends to approximately 150km, which corresponds with the LAB we identify at 170km. Although the mid-lithosphere of MBAR1 exhibits a NVG of 3.8%,

this location is modeled with peridotite only composition due to its relatively broad 1-sigma bounds.

DBIC1 and DBIC2 have inversion profiles that are significantly different in appearance (Fig. 2-3) but have surprisingly consistent compositional fits with 1-2% velocity reductions in the upper 150 km and velocity increases from diamond at approximately 200km depth.

RAYN samples the edge of the Arabian shield at the location of a sharp lateral velocity gradient (Celli et al., 2020). While in our inversions it appears we have resolved a lateral discontinuity in the LAB, RAYN2 does not exhibit the large velocity drop that is expected at in the non-cratonic continental lithosphere. It is possible that the absolute velocity structure we model using the global GDM52 model simply does not adequately resolve velocities at the location of these stacks.

3.1.4 Asia

Inversion profiles in Asia consist of the Dharwar craton in India and the North China craton, both of which have evidence of lithospheric thinning or delamination (Shi et al., 2021). At all profiles in these regions we estimate LAB depths less than 200km.

Thinning of the Dharwar craton has been proposed to be related to plume activity that emplaced the Deccan volcanic province (Shi et al., 2021). Our results are consistent with (Mitra et al., 2006) that interprets LAB depth to be 155 km at HYB.

In the eastern block of the North China craton delamination of the lithosphere likely occurred in the early cretaceous (Shi et al., 2021). Our profiles sample the eastern block (TIA1), Trans-North CHian Orogen (LYN1) and the edges of the western block of the North China craton (HHC1 and LZH1). TIA1 and LYN1 exhibit low velocities and a sharp NVG in the

uppermost mantle indicating a lack of cratonic lithosphere (Fig. 2-4b). The LABs we estimate at LZH1 and HHC1 agree with those estimated by Chen et al. (2014). HHC1 exhibits a distinct low velocity layer modeled with a 3% velocity drop that agrees with the NVG observed by Chen et al. (2014). This low velocity layer is likely related to alteration from nearby historic subduction and subsequent delamination of adjacent cratonic lithosphere (Jiang et al., 2013; Shi et al., 2021). LZH1 exhibits a strange structure. It is successfully modeled with 4-6% velocity reductions in the upper 80km of modeled depths underlain by 3% graphite that does not extend below the graphite-diamond transition with a 45 mW/m² geotherm despite having no significant velocity gradients within the lithosphere. It is likely that the model for LZH1 is artificially modeling a complicated compositional structure because the true geotherm at this location is higher than those we explore. LZH1 does not have any significant VGs and doesn't appear to reflect mid-lithospheric discontinuities observed in the Ordos block by Jiang et al. (2013).

3.1.5 Australia

In the inversion profiles that sample the Australian lithosphere, we observe remarkably similar structure. BBO03, NWA01, and STKA3 all can be modeled with the presence of diamond beginning at 170-200km, although STKA3 could also be modeled by the presence of a broad low velocity layers throughout the upper lithosphere. All NVGs we observe in these profiles are extended in depth and do not correspond with negative phases observed by Ford et al. (2010). Although, it does appear they observe a PVG at ~150km depth at STKA. Furthermore, Chen et al. (2021) does not observe positive impedance contrasts in the Australian cratonic lithosphere.

3.1.6 Antarctica

We observe a variety of structures in Antarctica. AGO11 is best fit by a with 3% reduced velocities above 90km and had a LAB at 172km indicating a thinned lithosphere. It also appears

there may be higher velocities in the lower lithosphere at AGO11 that may be underfit by our compositional model. AGO11 is located near the southern Transantarctic Mountains. In this region, Shen, Wiens, Anandakrishnan, et al. (2018) presents clear evidence of lithospheric foundering. Our interpretations of a thinned lithosphere as well as a low velocity zone followed by a positive velocity gradient are all consistent with lithospheric foundering in this region.

The P0611 profile exhibits high velocities in the uppermost lithosphere and a velocity increase that can be modeled with 2% diamond at 220km, although results using regional Rayleigh phase velocities do not resolve either of these features (Fig. 2-4c). N1401 displays an odd Moho signal that could be interpreted either as a shallow Moho followed by a significant PVG or an exceptionally deep (>50km) Moho. To fit this odd Moho structure, the compositional fit of the global inversion models a 3-4% velocity drop from above 60km. The regional profile for this location indicated a velocity increase at 200km depth that is not reflected in the profile using GDM52. This strange structure is consistent with previous work that finds that the in the Gamburtsev Subglacial Mountains (GSM), where N1401 is located, Moho estimations calculated with different geophysical techniques are not consistent. For example, Moho depth estimates calculated with joint inversion of Rayleigh wave phase velocities and Ps RFs find a deep Moho (~55 km) and studies of gravity find a shallow Moho (Block et al., 2009; Shen, Wiens, Anandakrishnan, et al., 2018 & references therein) Shen 2018 proposes this could be due to eclogized lower crust that is high density, but lower seismic velocity than the underlying cratonic mantle. Our shallow Moho result at N1401 is consistent with gravity data, but inconsistent with other seismic studies that interpret deep crust-mantle boundaries (Ramirez et al., 2016; Shen, Wiens, Anandakrishnan, et al., 2018). P0611 is also adjacent to the GSM and does exhibit a

smaller PVG (4.2%) in the uppermost mantle, but its Moho velocity gradient very clearly at shallow depths, supporting our identification of a shallow Moho at N1401.

Additionally, AGO31 is not modeled due to complex structure. AGO31 is located in the Recovery Subglacial Highlands west of the Gamburtsev Subglacial Mountains in East Antarctica. SYO1 is located on the northern boundary of the East Antarctic Orogen, adjacent to the East Antarctic coast.

S4 Supplementary Tables

Table S B-1. Ranges of the distributions of inversion parameters. Standard deviation describes the standard deviation used in the perturbation to describe how a model can vary between iterations.

Parameter	Minimum	Maximum	Standard deviation
Crustal Structure			
Layer thickness	20	60	3.3
Shear Velocity	2.5	4.2	0.08
Vp/Vs	1.62	1.9	0.01
Splines	2	5	2
Mantle Structure			
Vs	3.5	5.2	0.08
Splines	7	22	4
Data			
Sp RF sigma	-inf	0.5	
Rayleigh phV sigma	1e-2	0.25	

Table S B-2. Stacks organized by their velocity gradient (VG) structure. Positive velocity gradients (PVGs), negative velocity gradients (NVGs). Gray italics indicate the appearance of structure without VGs that do not meet the 2% VG criteria.

PVG Only	NVG + PVG	NVG only	No sharp VGs
OB2	P0611	CCM2	RSSD2
KON01	BBOO3	HYB1	KURK3
AGO11	NWAO1	CCM1	ECSD2
N1401	OB1	BRVK1	HYB2
STKA3	HHC1		LZH1
BOSA1	LBTB1		RAYN1
DBIC2	RSSD1		FFC1
ALE2	DBIC1		
K43A1	LSZ1		
JFWS1	AGMN1		
SYO1	EYMN1		
KURK1	ARU1		
YKW1	AGO31		
	MBAR1		
	TSUM1		

Table S B-3. Detailed results of the compositional models fit to inversion models.

Station	Age	Geotherm	Vel Increase Composition	Vel Incr. Depth	Velocity decrease	LVL Depth
RSSD2	pProt.	40				
KURK3	Phan.	35				
MBAR1	pProt.	40				
ECSD2	pProt.	35				
CCM2	m/nProt.	35				
TSUM1	Arch.	35				

HYB2	m/nProt.	35				
OBN2	m/nProt.	35	20% Ecl, 1% Diamond	169		
P0611	pProt.	35	1% Diamond	220		
JFWS1	pProt.	35	1% Diamond	58		
BBOO3	Arch.	40	20% Ecl, 3% Diamond	172		
NWAO1	Arch.	40	3% Diamond	200		
OBN1	m/nProt.	35	20% Ecl, 2% Diamond	168		
HHC1	pProt.	45			3%	70-110
		45			3%	70-100
LBTB1	Arch.	35			3%	90-140
AGO11	Phan.	35			3%	50-90
		35			3%	50-100
N1401	pProt.	35			4%	50-60
		35			3%	50-60
SYO1	pProt.	40	20% Ecl, 1% Diamond	50		
		40	20% Ecl, 1% Diamond	60		
		40	20% Ecl, 1% Diamond	70		
		35			1%	60-120
		35			1%	60-130
		35			1%	70-120
		35			1%	70-130
KURK1	Phan.	40	20% Ecl, 1% Diamond	160		
		40	20% Ecl, 1% Diamond	170		
		35			2%	50-150
STKA3	m/nProt.	40	20% Ecl, 1% Diamond	170		

		35			2%	50-150
KONO1	m/nProt.	40	2% Diamond	50		
		40	2% Diamond	140		
		35			2%	50-120
		35			2%	50-110
RSSD1	Arch.	35	2% Diamond	160	3%	80-140
DBIC1	pProt.	35	3% Diamond	220	1%	90-150
		35	3% Diamond	220	2%	100-140
		35	3% Diamond	220	2%	100-150
		35	3% Diamond	220	1%	80-150
		35	20% Ecl, 2% Diamond	210	1%	90-150
		35	3% Diamond	220	2%	90-140
		35	3% Diamond	220	2%	90-150
		35	20% Ecl, 2% Diamond	210	2%	100-140
		35	20% Ecl, 2% Diamond	210	2%	100-150
		35	20% Ecl, 2% Diamond	210	1%	80-150
		35	3% Diamond	220	1%	100-150
BOSA1	Arch.	35	20% Ecl, 1% Diamond	180	2%	50-100
YKWsl	pProt.	35	2%	110	2%	50-70
		35	2%	110	2%	50-80
		35	2%	110	1%	50-80
		35	2%	110	1%	50-90
LZH1	pProt.	45	4%	66	6%	56-66
		45	3%	66	6%	56-66
		45	3%	76	4%	56-76
		45	3%	66	5%	56-66
DBIC2	pProt.	35	1%	190	2%	50-130
		35	1%	190	2%	50-120

		35	1%	200	2%	50-130
		35	1%	200	2%	50-120
ALE2	Phan.	35	20% Ecl, 1% Diamond	160	2%	50-130
		35	20% Ecl, 1% Diamond	160	2%	50-140
		35	2% Diamond	170	2%	50-130
		35	2% Diamond	170	2%	50-140

S5 Supplementary Figures

S5.1 Methods

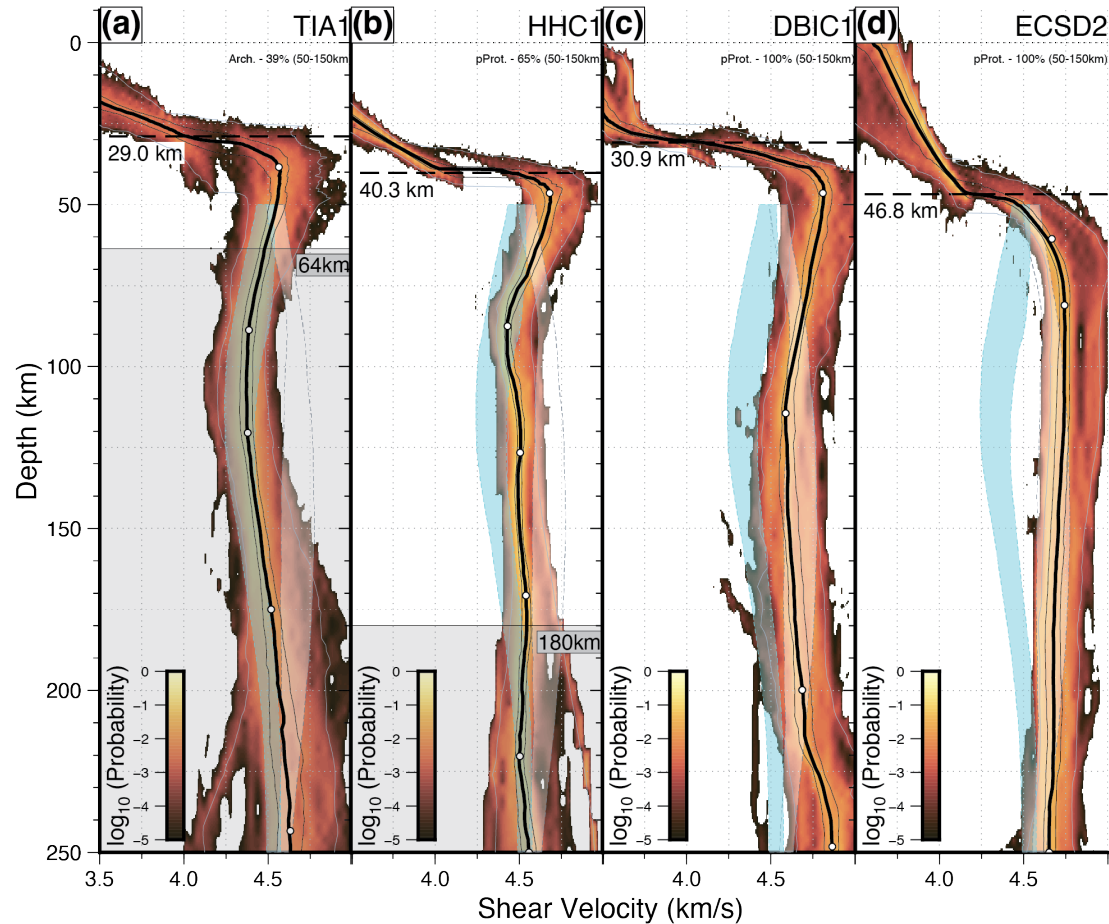


Figure S B-1. Explanation of cratonic lithosphere characterization

Explanation of cratonic lithosphere characterization. Inversion results plotted as in Figure 2-2(a) overlain by the median average deviation bounds for noncraton (cyan) and cratonic (white-shaded) classification based on shear velocities from SEMum2 global model (REF) for locations Artemieva (YEAR) identifies as Phanerozoic (cyan-shaded) and Archean-Paleoproterozoic (white-shaded). We compare the lower Vs deviation bound for Archean and Paleoproterozoic thermotectonic age regions (white-shaded) to the 2-sigma upper bound (gray outline) of the velocity profile inversion result. If the 2-sigma upper bound of the velocity profile exceeds the lower bound of the Archean and Paleoproterozoic profile in the depth range 50-150km for at least 50% of this range, we consider the profile to be cratonic. (a) TIA1 is considered noncratonic. (b) HHC1 is considered cratonic but has a low velocity layer that intersects with Phanerozoic velocities at depths < 100 km. The overlap of the lower bound of the HHC1 profile with Phanerozoic velocities below the LAB depth of 180 km is also shown. (c) DBIC1 is a 100% cratonic profile with a low velocity zone that stays within the Archean-Paleoproterozoic Vs

range. **(d)** ECSD2 exhibits high velocities that are within or above the Archean-Paleoproterozoic Vs range.

S5.2 Compositional Modelling

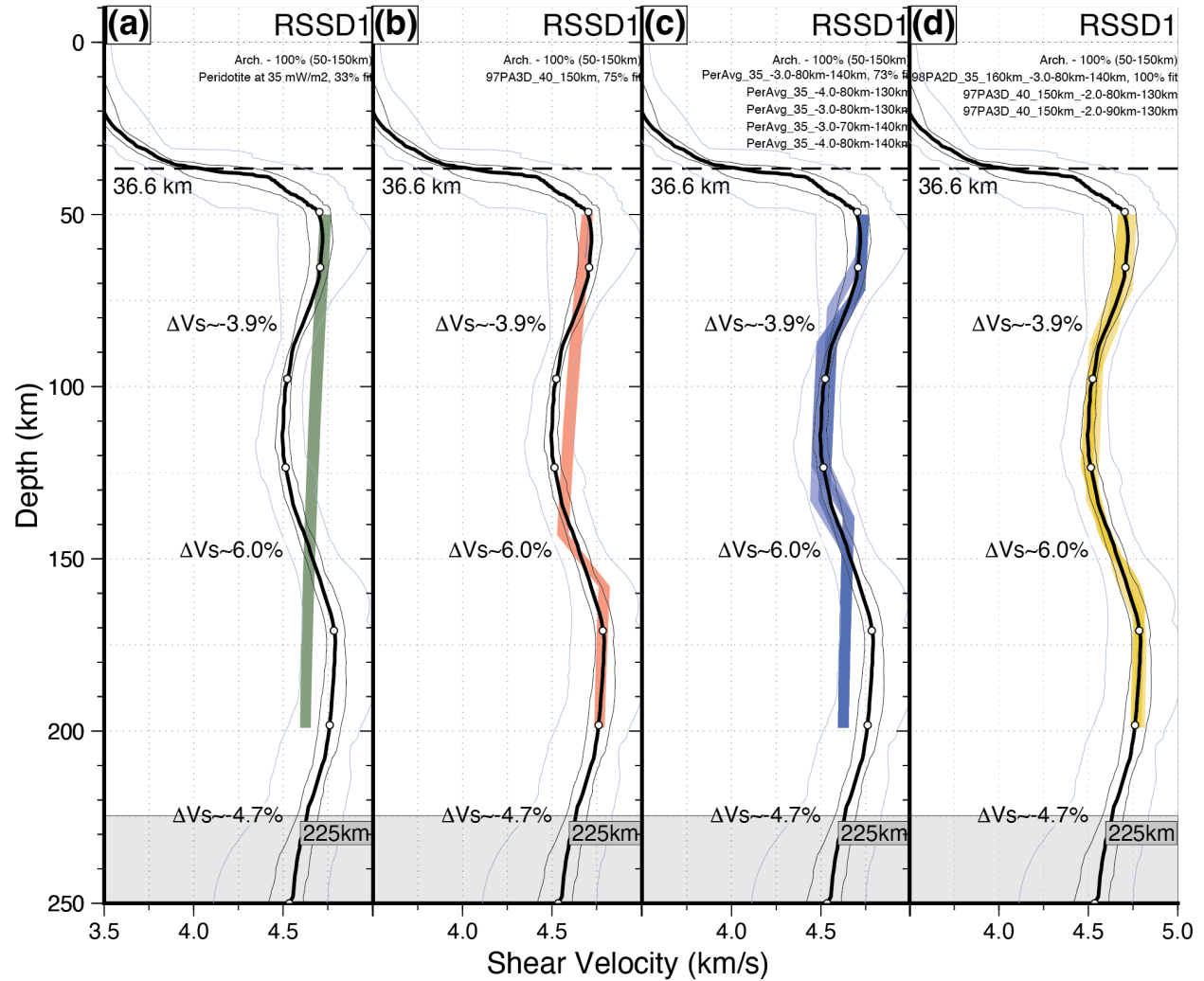


Figure S B-2. Progression of compositional model fits for RSSD 1

Progression of compositional model fits for RSSD1 **(a)** peridotite only, 33% fit; **(b)** velocity increase only 75% fit; **(c)** low velocity layer only, 73% fit; **(d)** low velocity layer and velocity increase 100% fit.

S5.3 Inversion Results

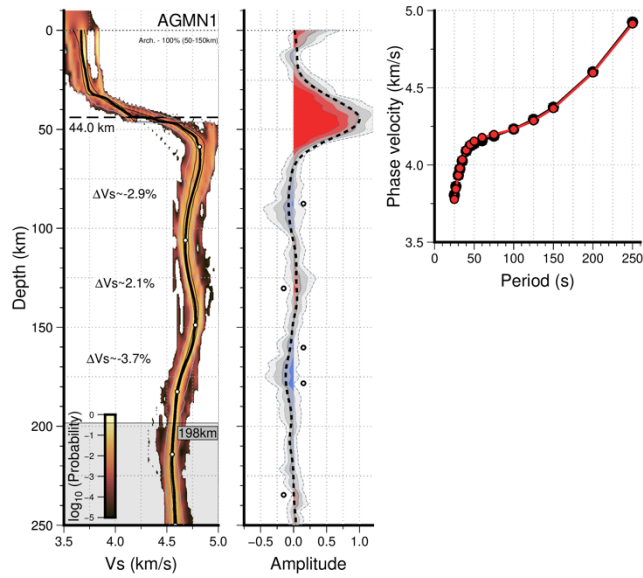


Figure S B-3. Inversion results AGMN 1 (global)

Inversion results for AGMN1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

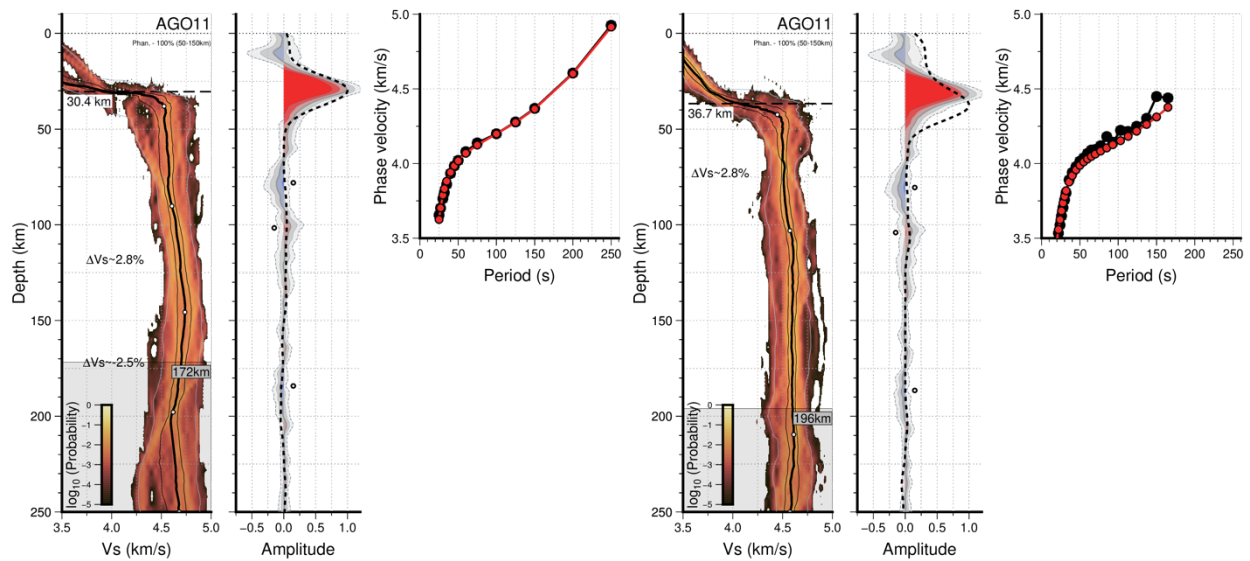


Figure S B-4. Inversion results AGO11 (global and regional)

Inversion results for AGO11 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

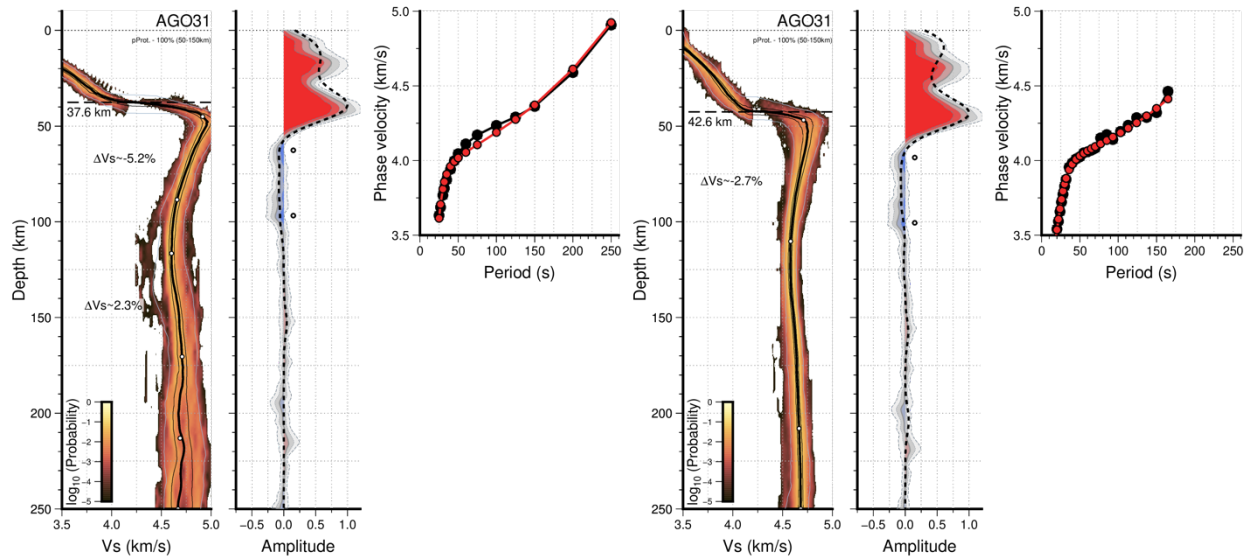


Figure S B-5. Inversion results AGO31 (global and regional)

Inversion results for AGO31 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

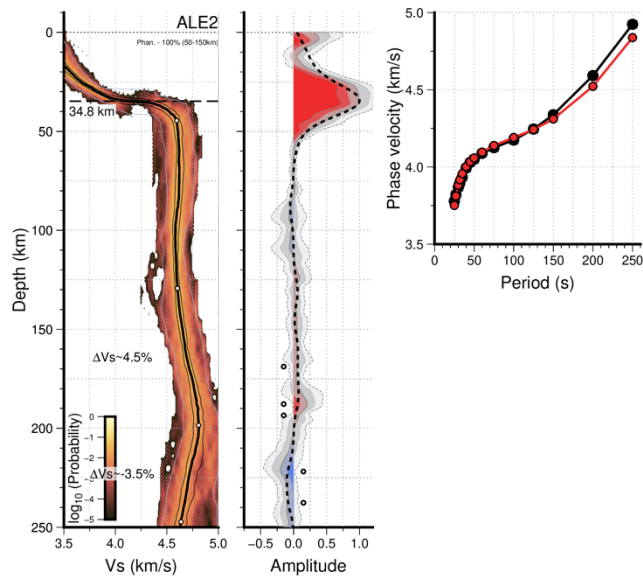


Figure S B-6. Inversion results ALE 2 (global)

Inversion results for ALE2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

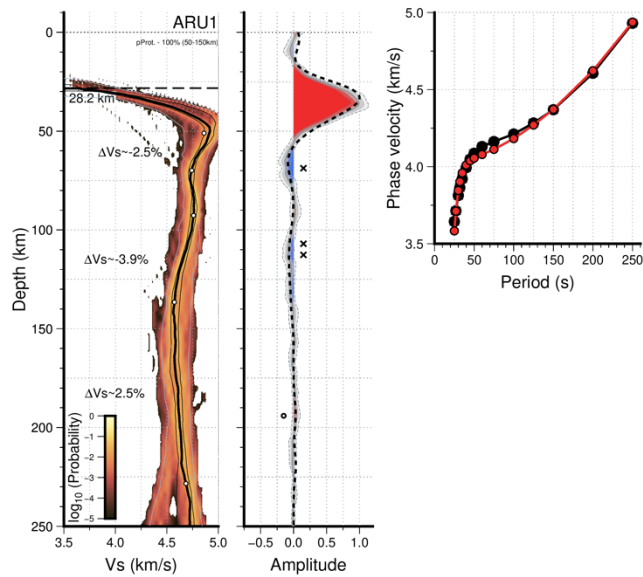


Figure S B-7. Inversion results ARU1 (global)

Inversion results for ARU1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

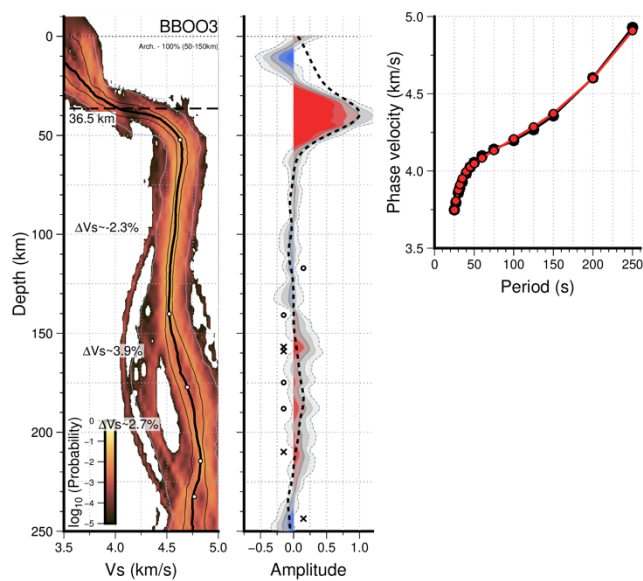


Figure S B-8. Inversion results BBOO3 (global)

Inversion results for BBOO3 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

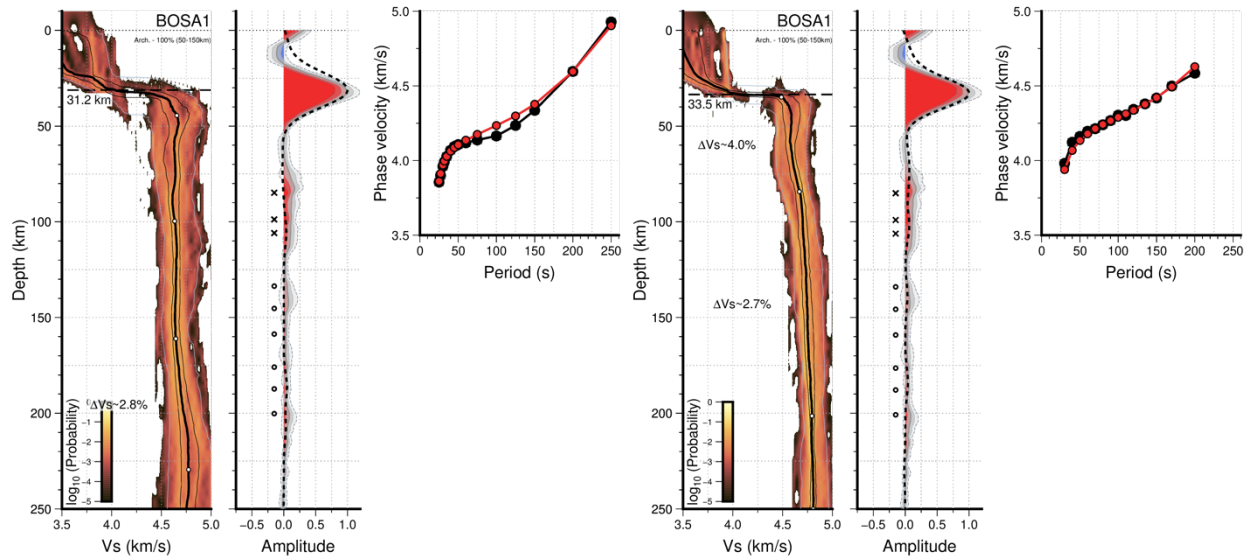


Figure S B-9. Inversion results BOSA1 (global and regional)

Inversion results for BOSA1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

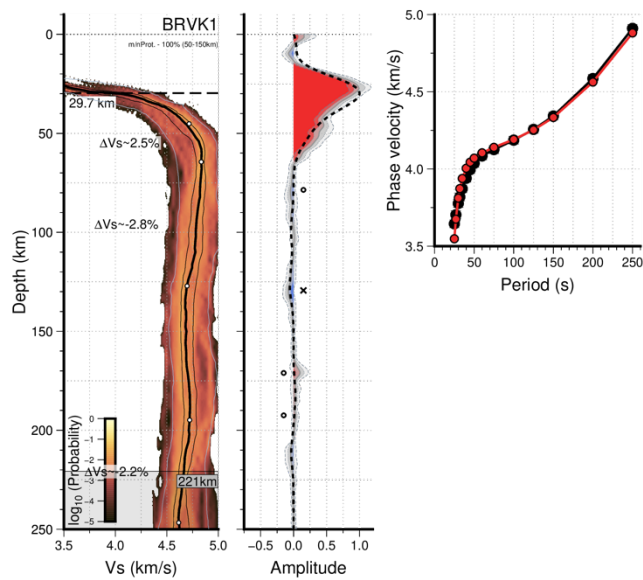


Figure S B-10. Inversion results BRVK1 (global)

Inversion results for BRVK1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

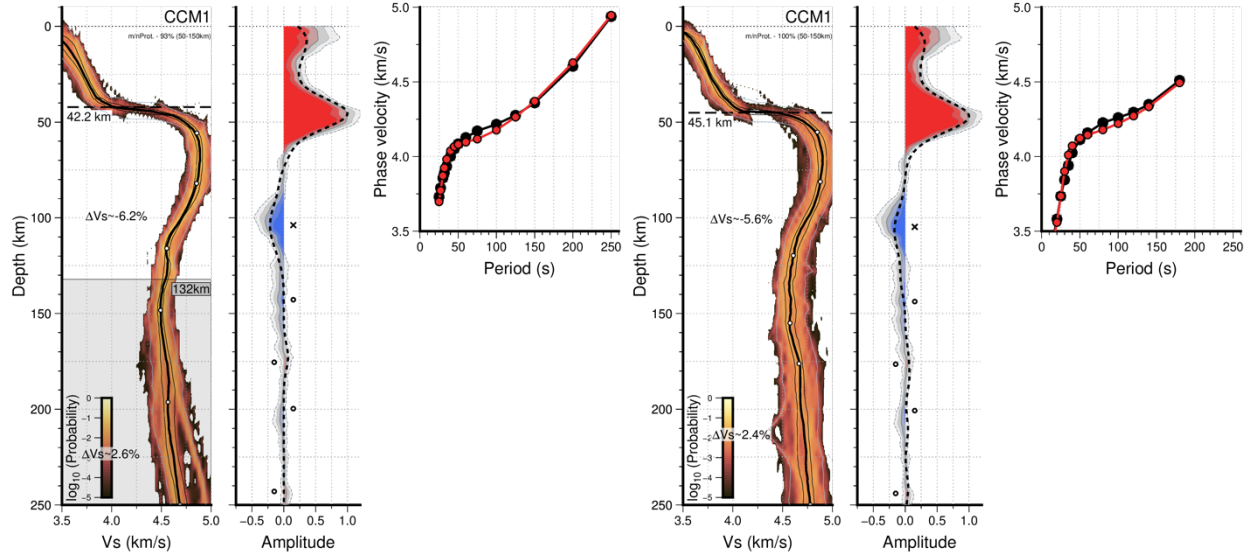


Figure S B-11. Inversion results CCM1 (global and regional)

Inversion results for CCM1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

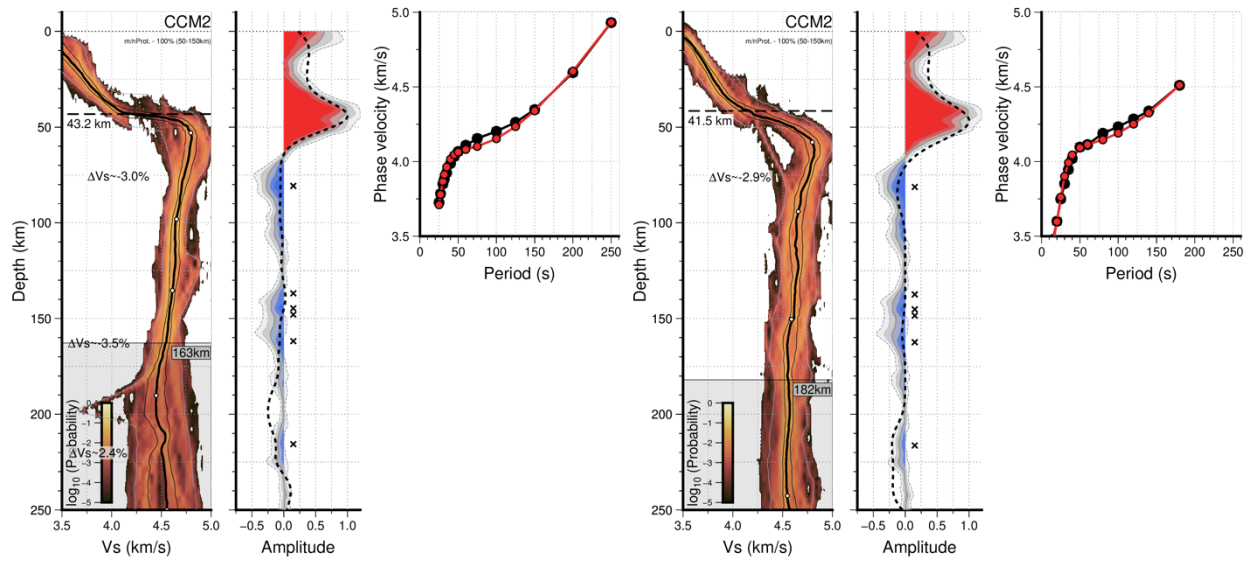


Figure S B-12. Inversion results CCM2 (global and regional)

Inversion results for CCM2 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

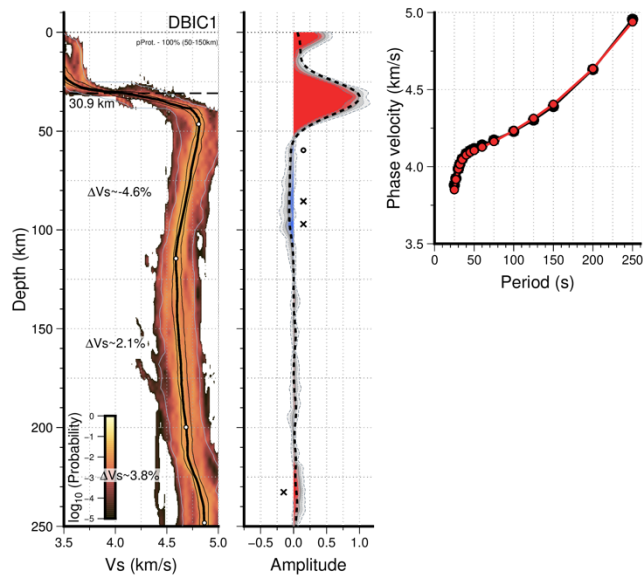


Figure S B-13. Inversion results DBIC1 (global)

Inversion results for DBIC1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

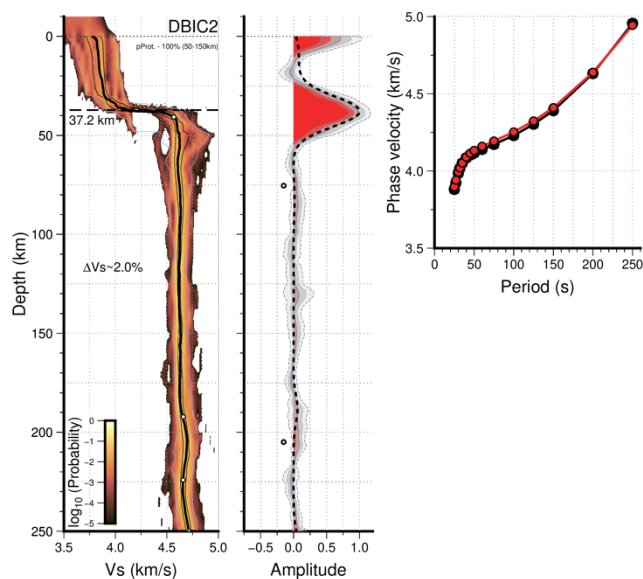


Figure S B-14. Inversion results DBIC2 (global)

Inversion results for DBIC2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

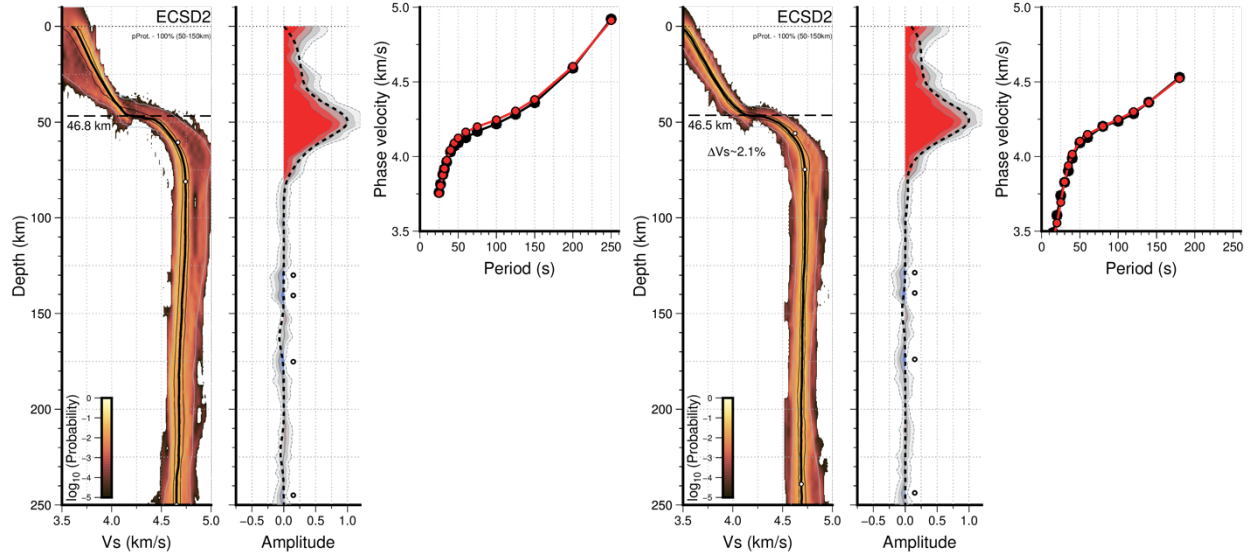


Figure S B-15. Inversion results ECSD2 (global and regional)

Inversion results for ECSD2 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

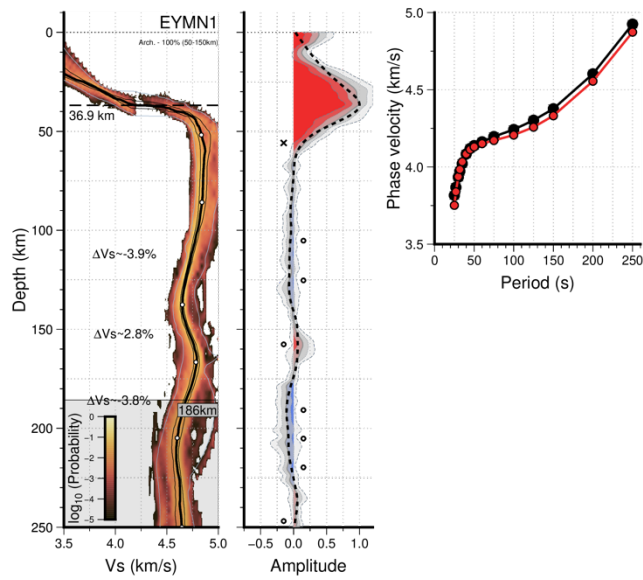


Figure S B-16. Inversion results EYMN1 (global)

Inversion results for EYMN1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

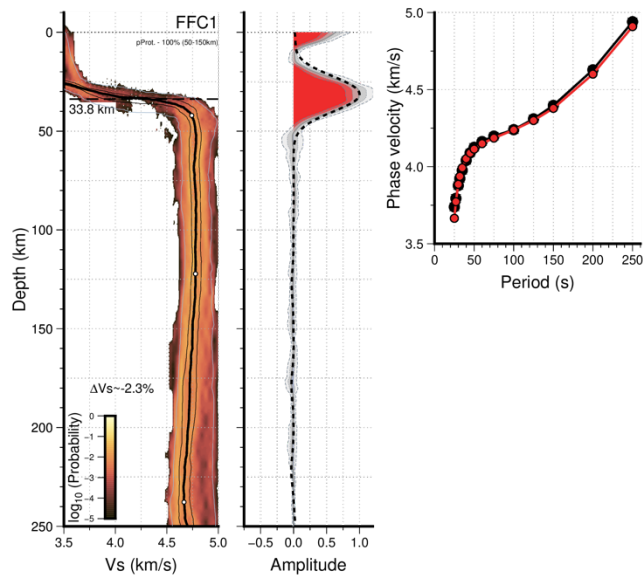


Figure S B-17. Inversion results FFC1 (global)

Inversion results for FFC1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

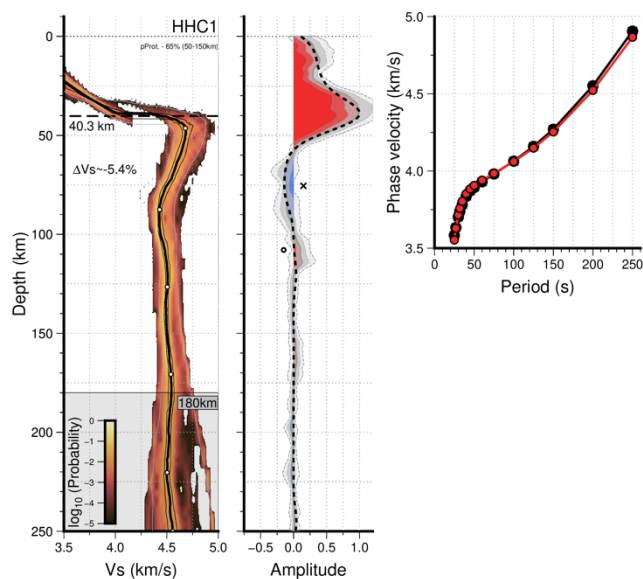


Figure S B-18. Inversion results HHC1 (global)

Inversion results for HHC1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

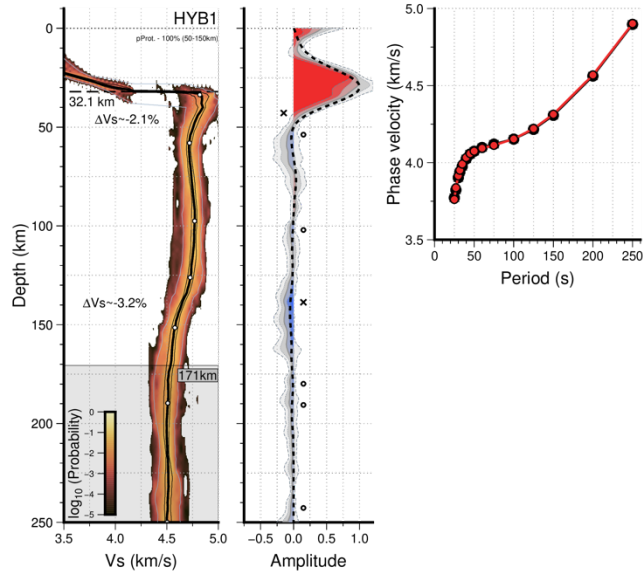


Figure S B-19. Inversion results HYB1 (global)

Inversion results for HYB1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

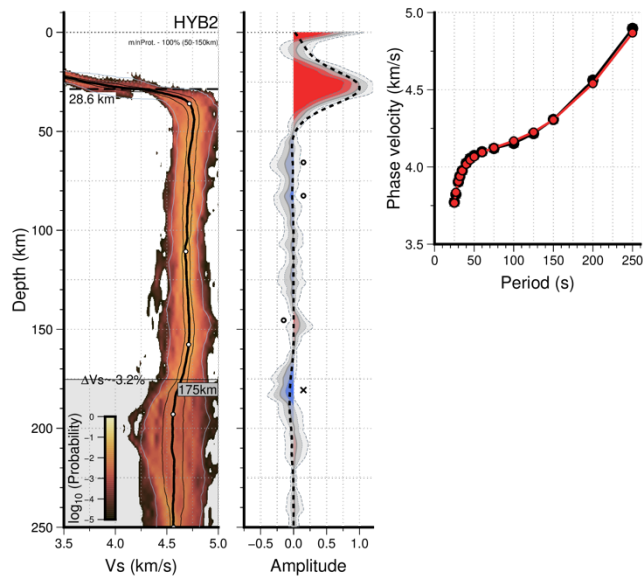


Figure S B-20. Inversion results HYB2 (global)

Inversion results for HYB2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

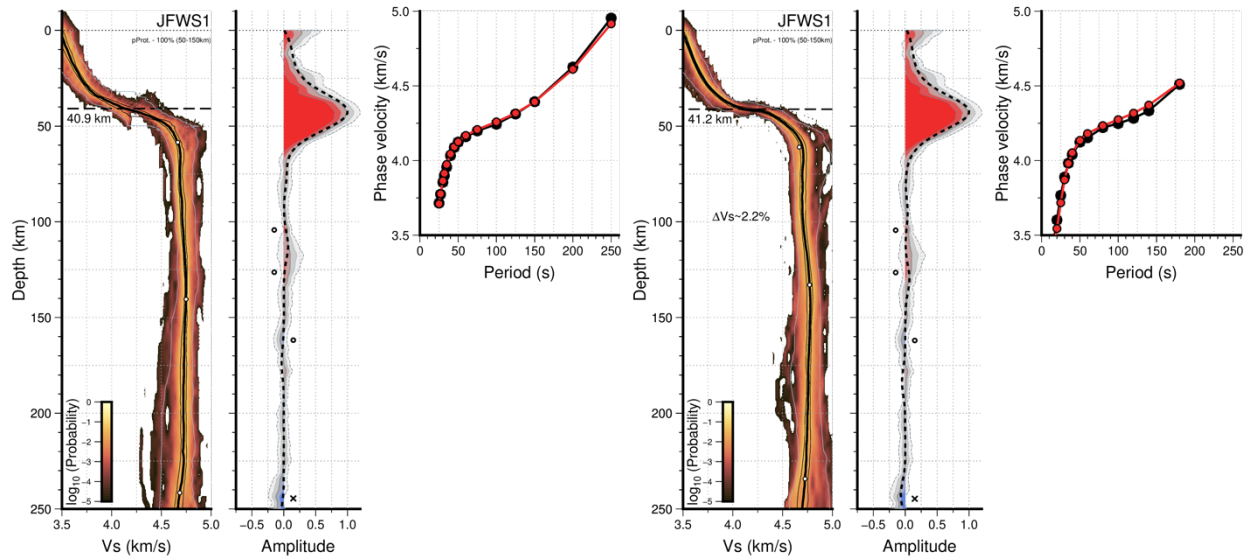


Figure S B-21. Inversion results JFWS1 (global and regional)

Inversion results for JFWS1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

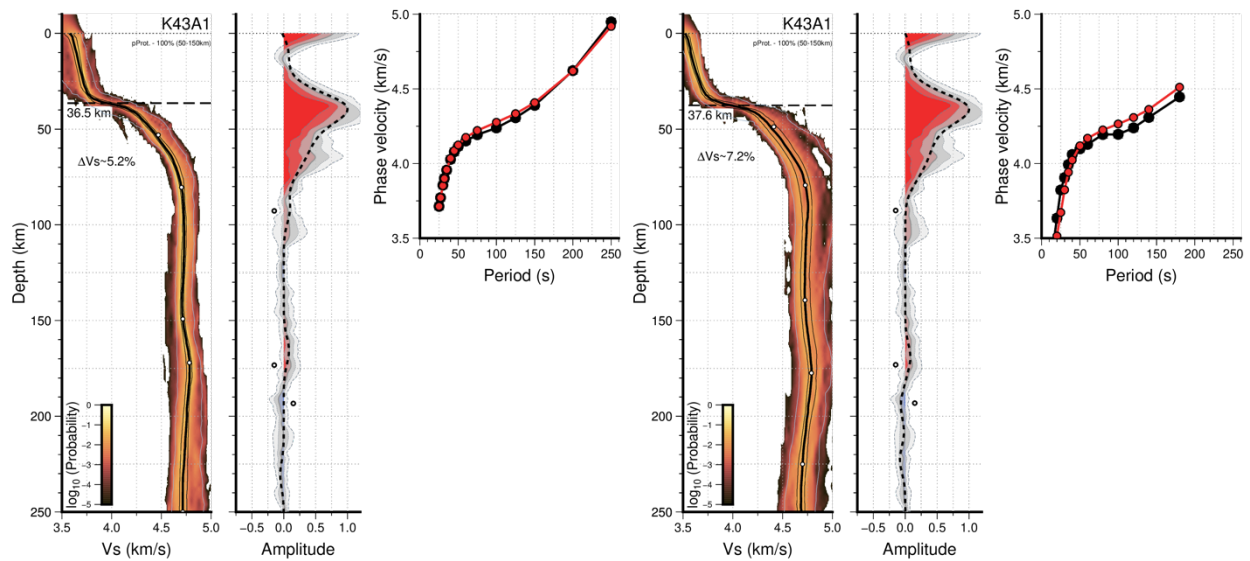


Figure S B-22. Inversion results K43A1 (global and regional)

Inversion results for K43A1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

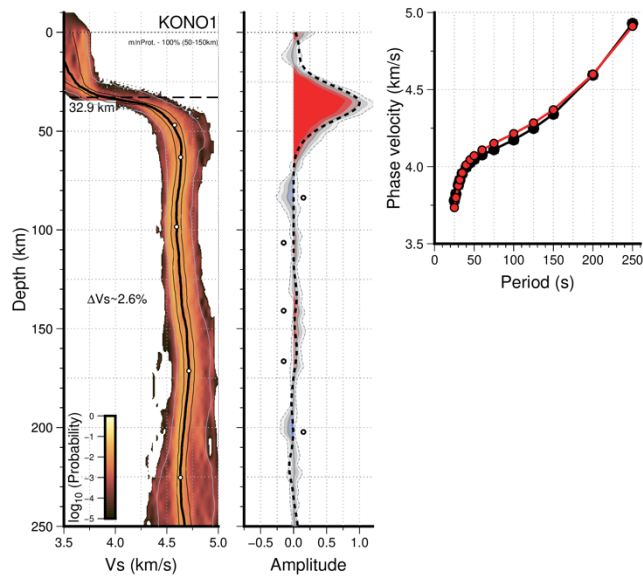


Figure S B-23. Inversion results KONO1 (global)

Inversion results for KONO1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

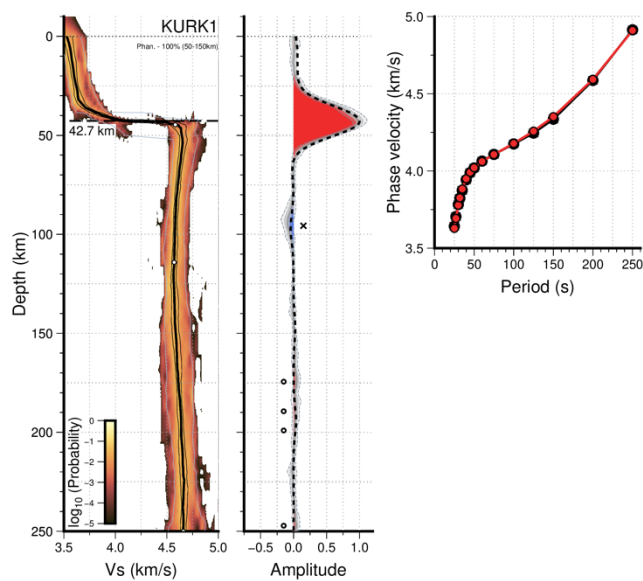


Figure S B-24. Inversion results KURK1 (global)

Inversion results for KURK1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

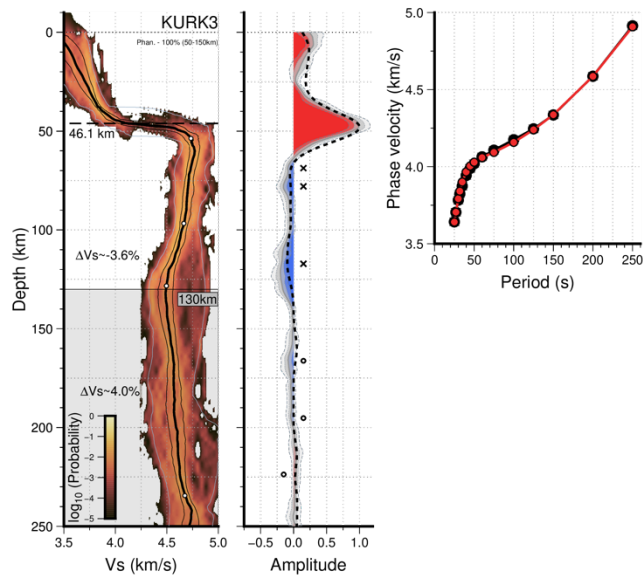


Figure S B-25. Inversion results KURK3 (global)

Inversion results for KURK3 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

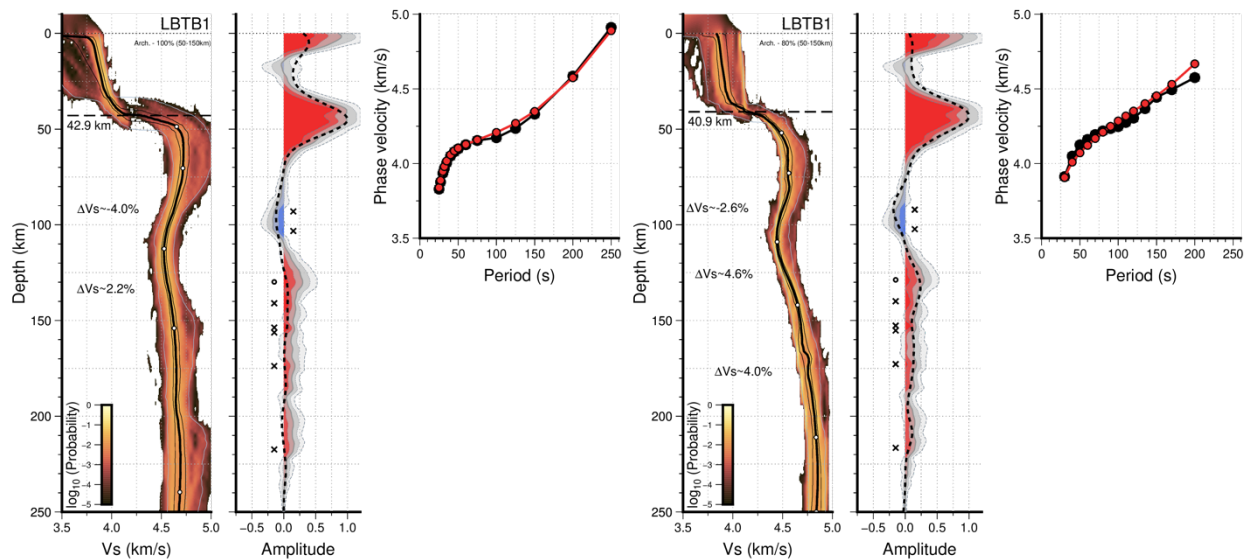


Figure S B-26. Inversion results LBTB1 (global and regional)

Inversion results for LBTB1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

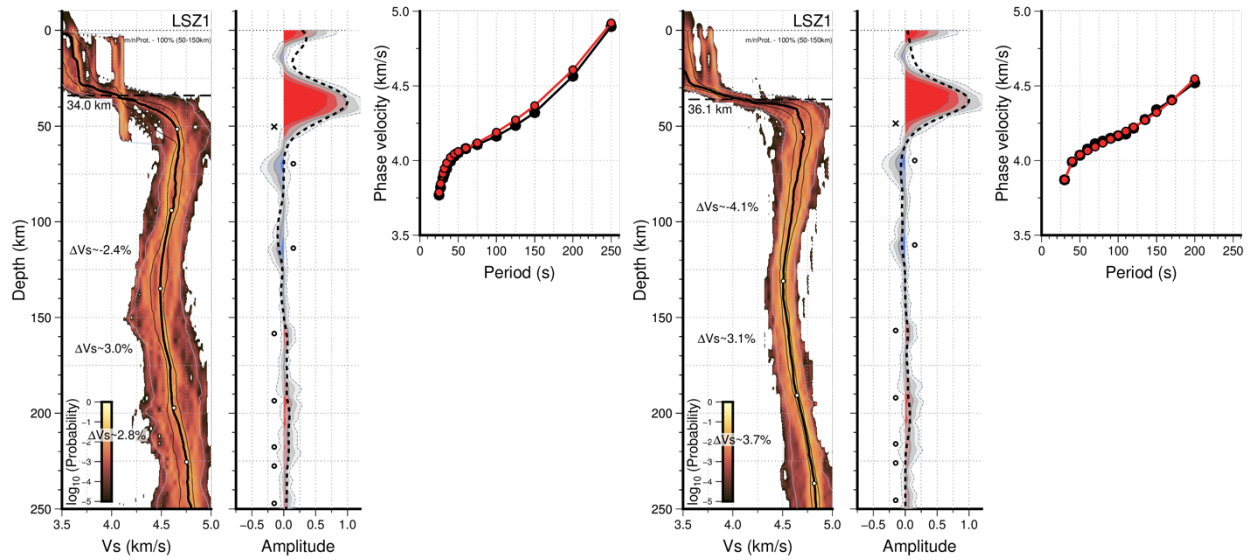


Figure S B-27. Inversion results LSZ1 (global and regional)

Inversion results for LSZ1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

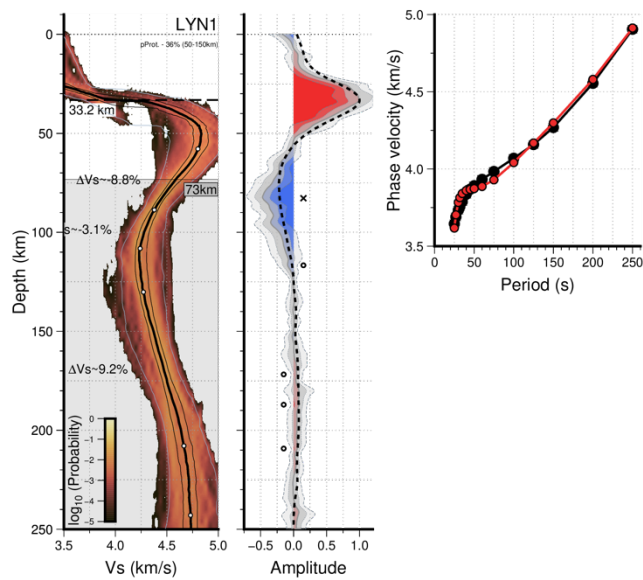


Figure S B-28. Inversion results LYN1 (global)

Inversion results for LYN1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

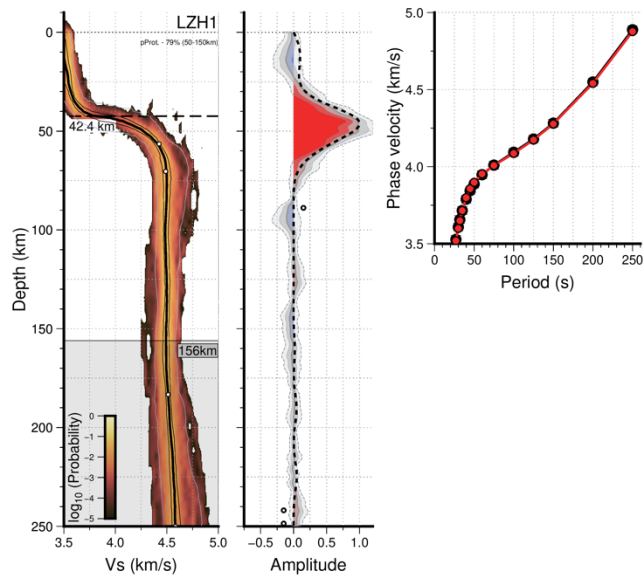


Figure S B-29. Inversion results LZH1 (global)

Inversion results for LZH1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

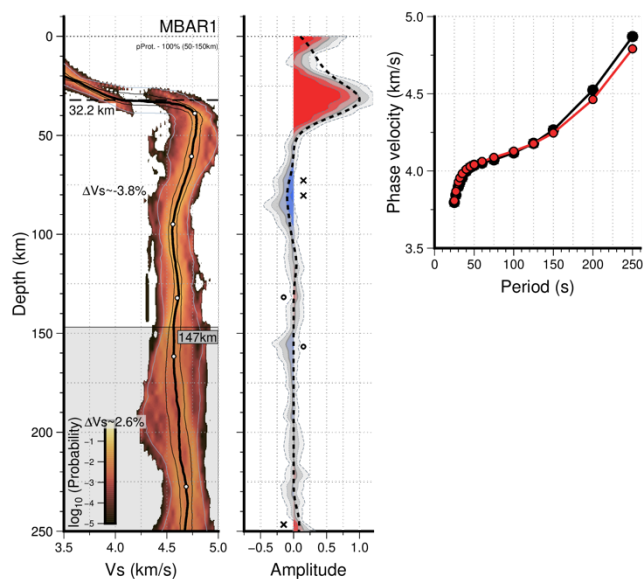


Figure S B-30. Inversion results MBAR1 (global)

Inversion results for MBAR1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

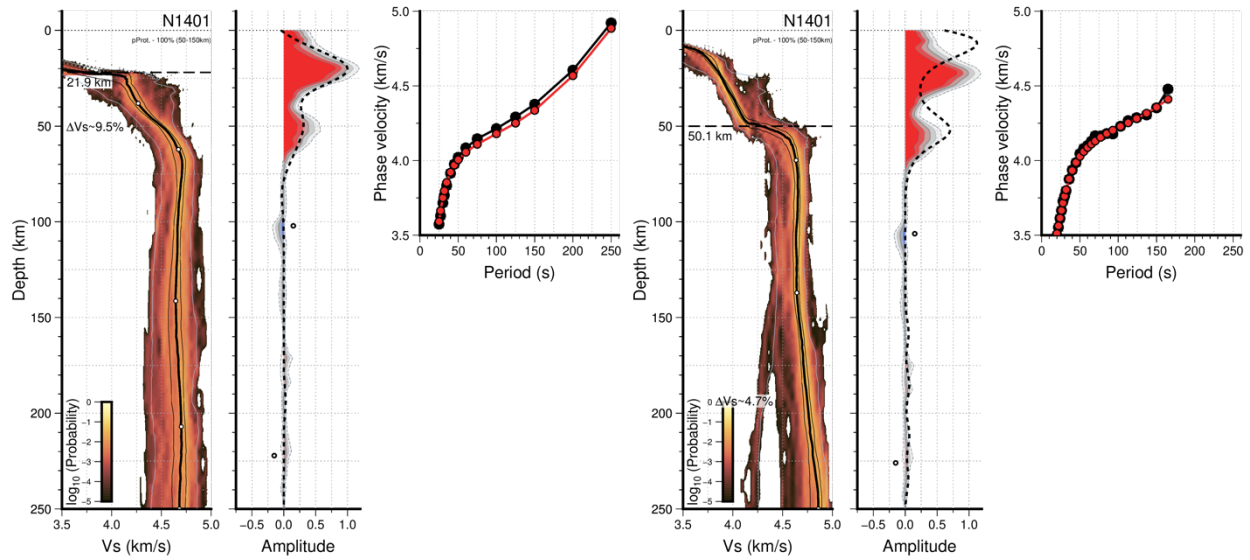


Figure S B-31. Inversion results N1401 (global and regional)

Inversion results for N1401 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

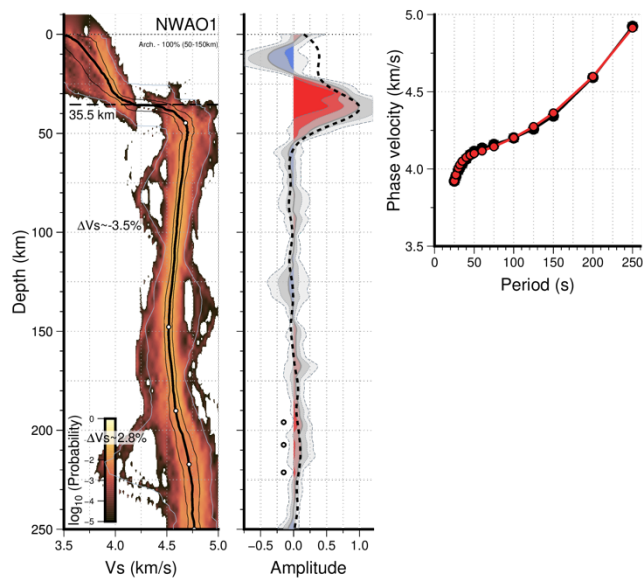


Figure S B-32. Inversion results NWA01 (global)

Inversion results for NWA01 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

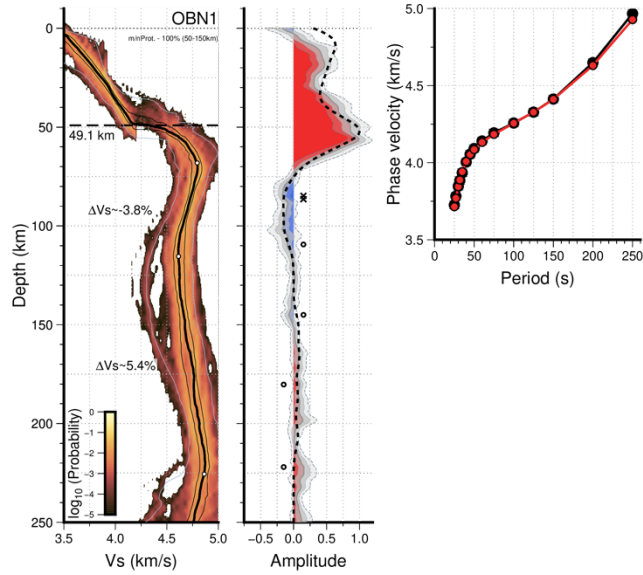


Figure S B-33. Inversion results OBN1 (global)

Inversion results for OBN1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

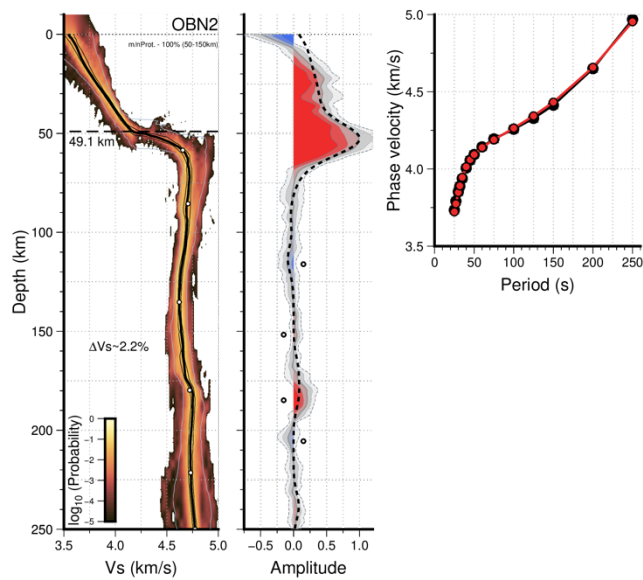


Figure S B-34. Inversion results OBN2 (global)

Inversion results for OBN2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

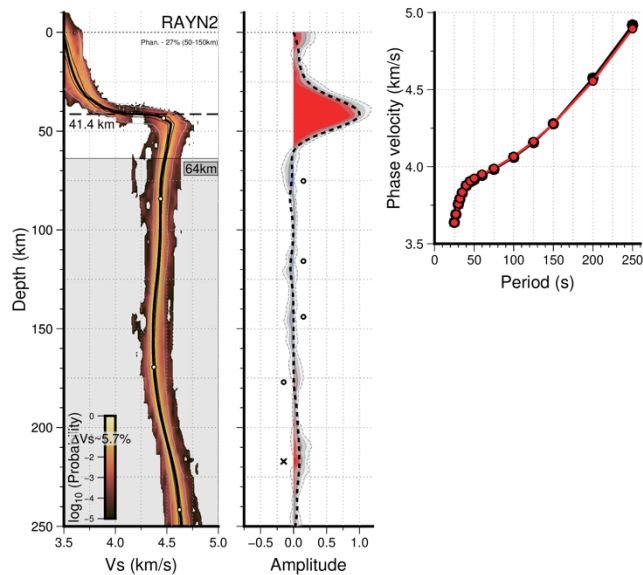


Figure S B-37. Inversion results RAYN2 (global)

Inversion results for RAYN2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

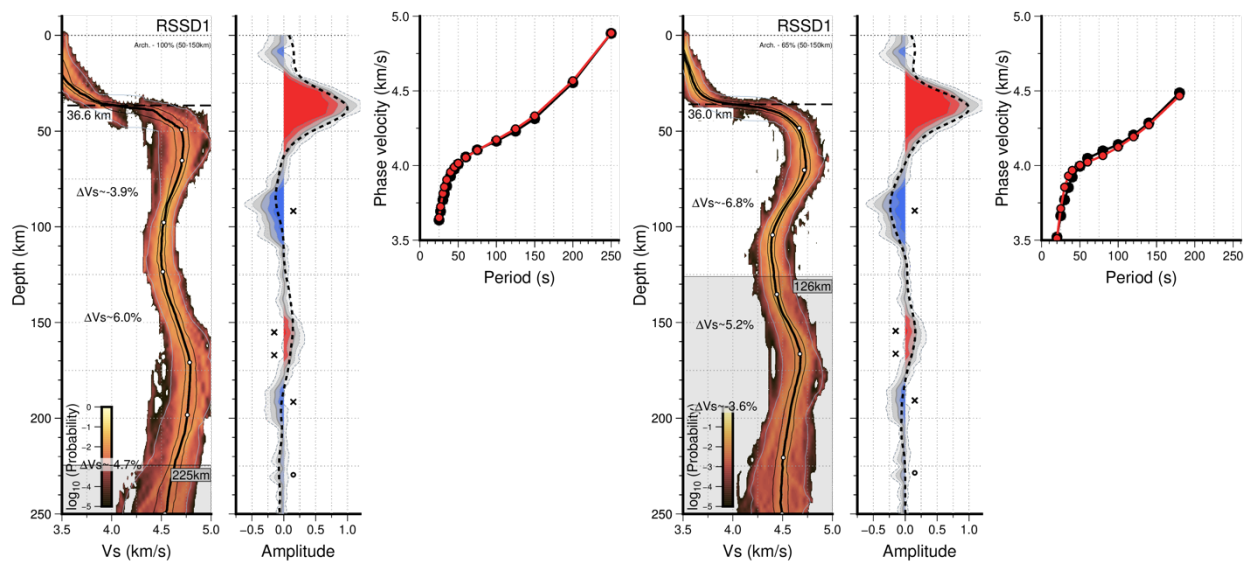


Figure S B-38. Inversion results RSSD1 (global and regional)

Inversion results for RSSD1 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

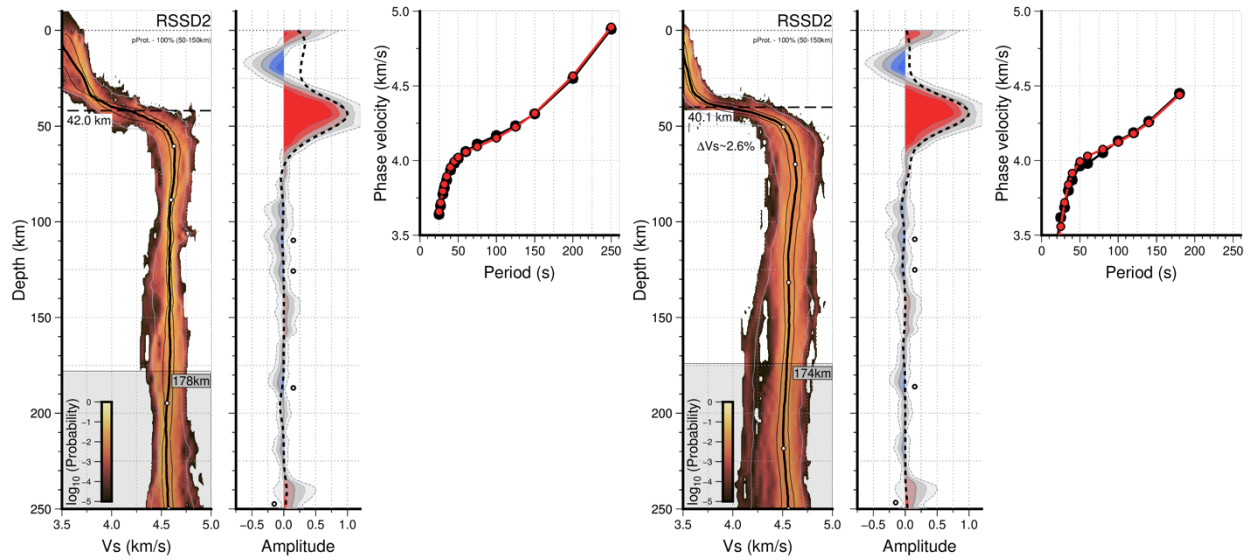


Figure S B-39. Inversion results RSSD2 (global and regional)

Inversion results for RSSD2 using (left) Rayleigh phase velocities from GDM52 and (right) regional phase velocities. Plotted as in Fig. 2-2.

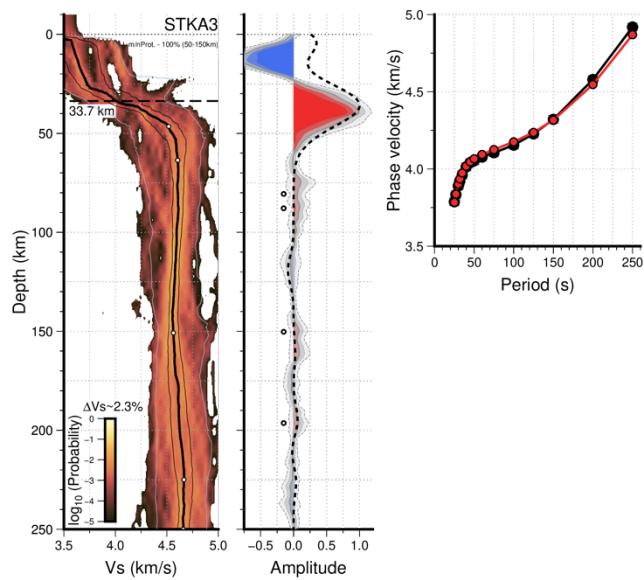


Figure S B-40. Inversion results STKA3 (global)

Inversion results for STKA3 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

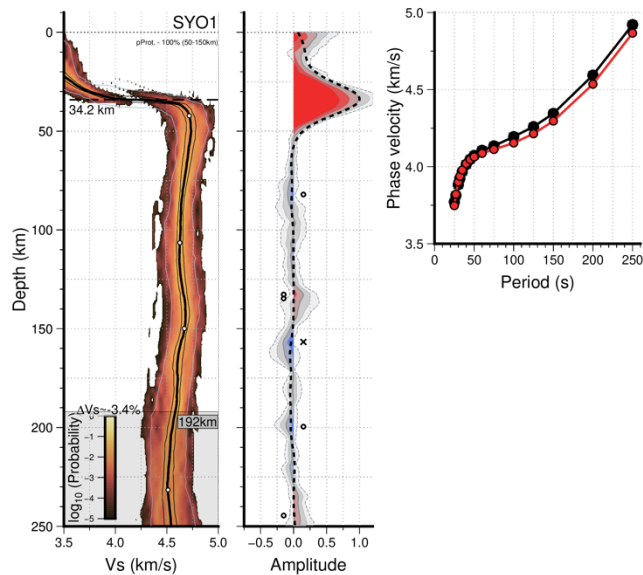


Figure S B-41. Inversion results SYO1 (global)

Inversion results for SYO1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

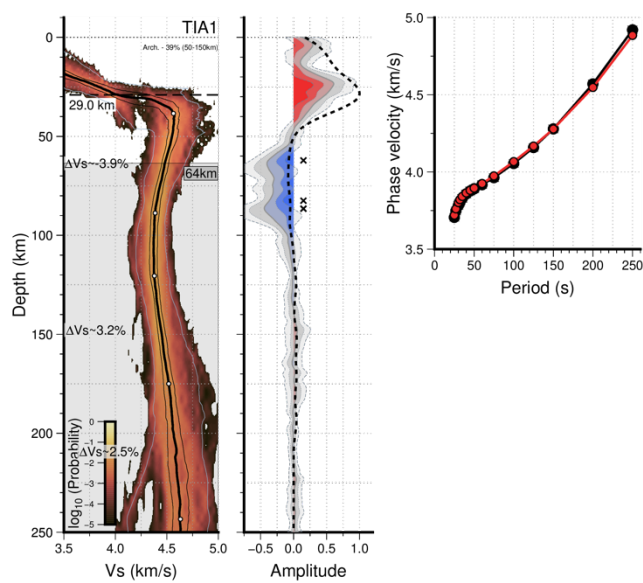


Figure S B-42. Inversion results TIA1 (global)

Inversion results for TIA1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

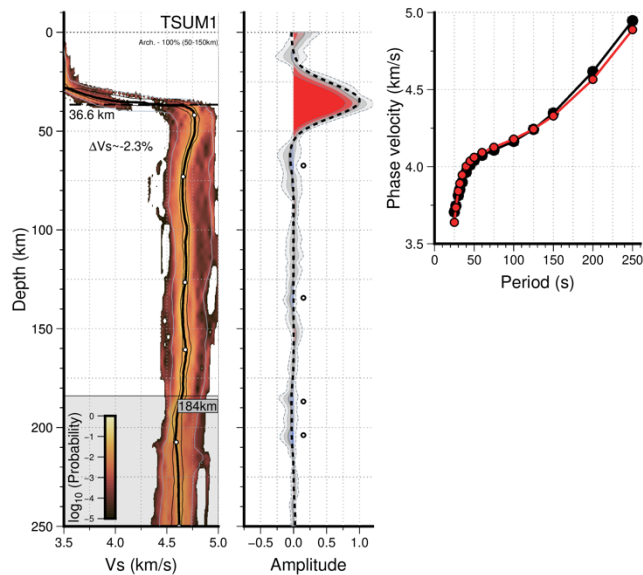


Figure S B-43. Inversion results TSUM1 (global)

Inversion results for TSUM1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

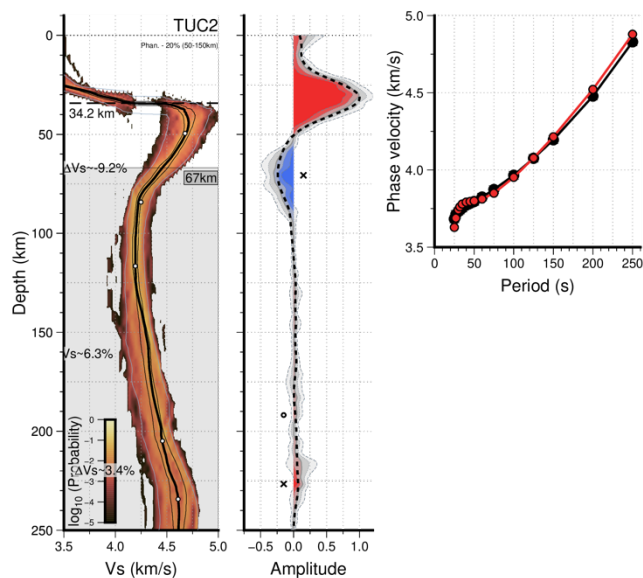


Figure S B-44. Inversion results TUC2 (global)

Inversion results for TUC2 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

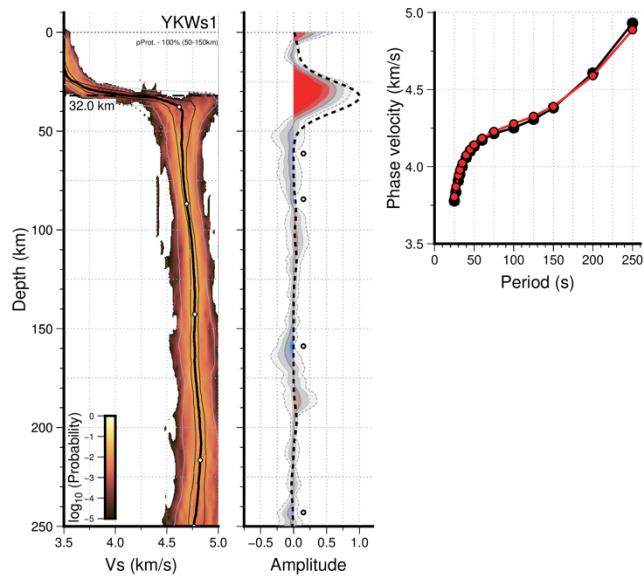


Figure S B-45. Inversion results YKWs1 (global)

Inversion results for YKWs1 using Rayleigh phase velocities from GDM52. Plotted as in Fig. 2-2.

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Appendix C Supplementary Materials: Investigating the Complex Tectonic History of Antarctica with Phase Velocity & Attenuation

S1. Supplementary Figures

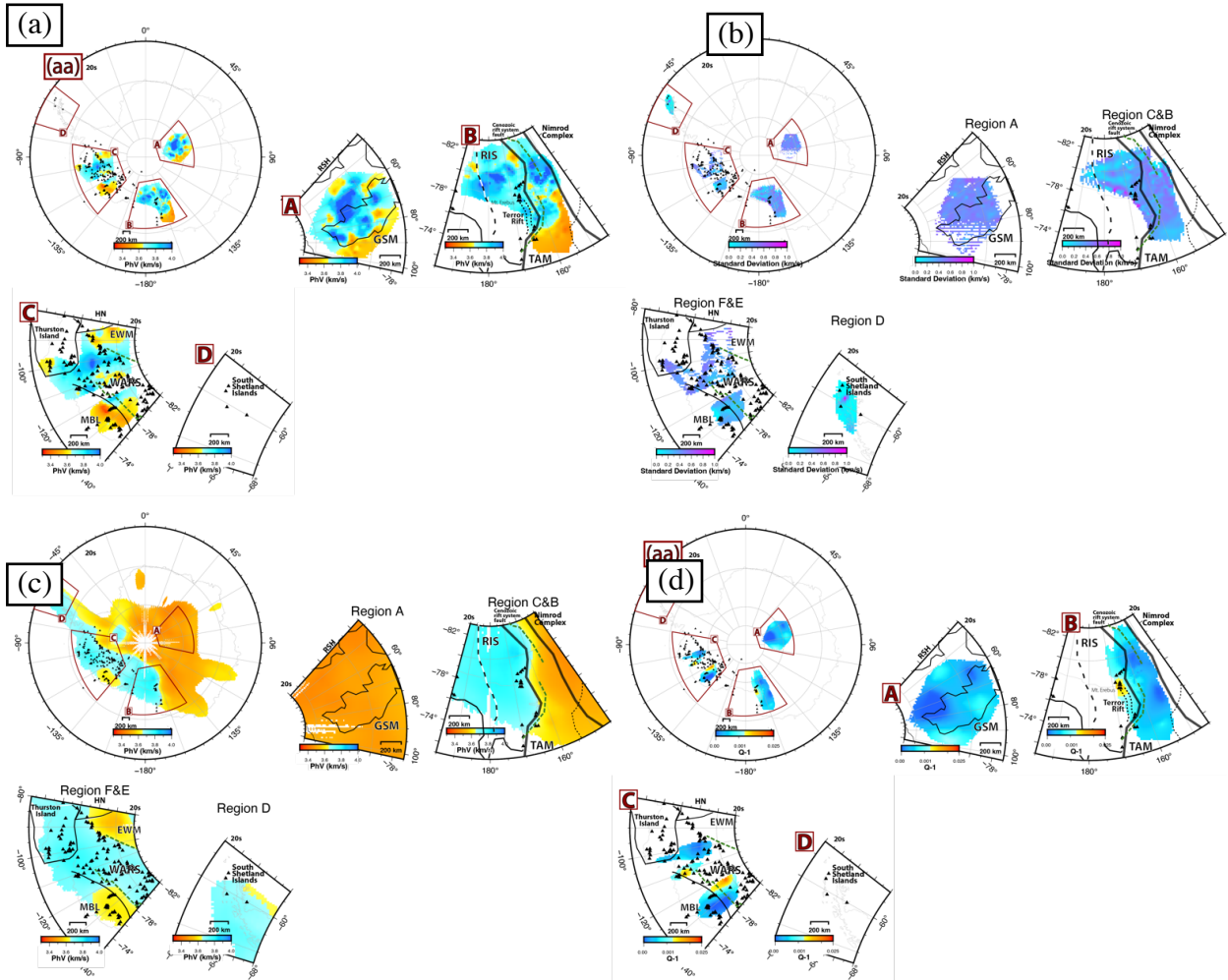


Figure S C-1. Results 20s period

Results for 20s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

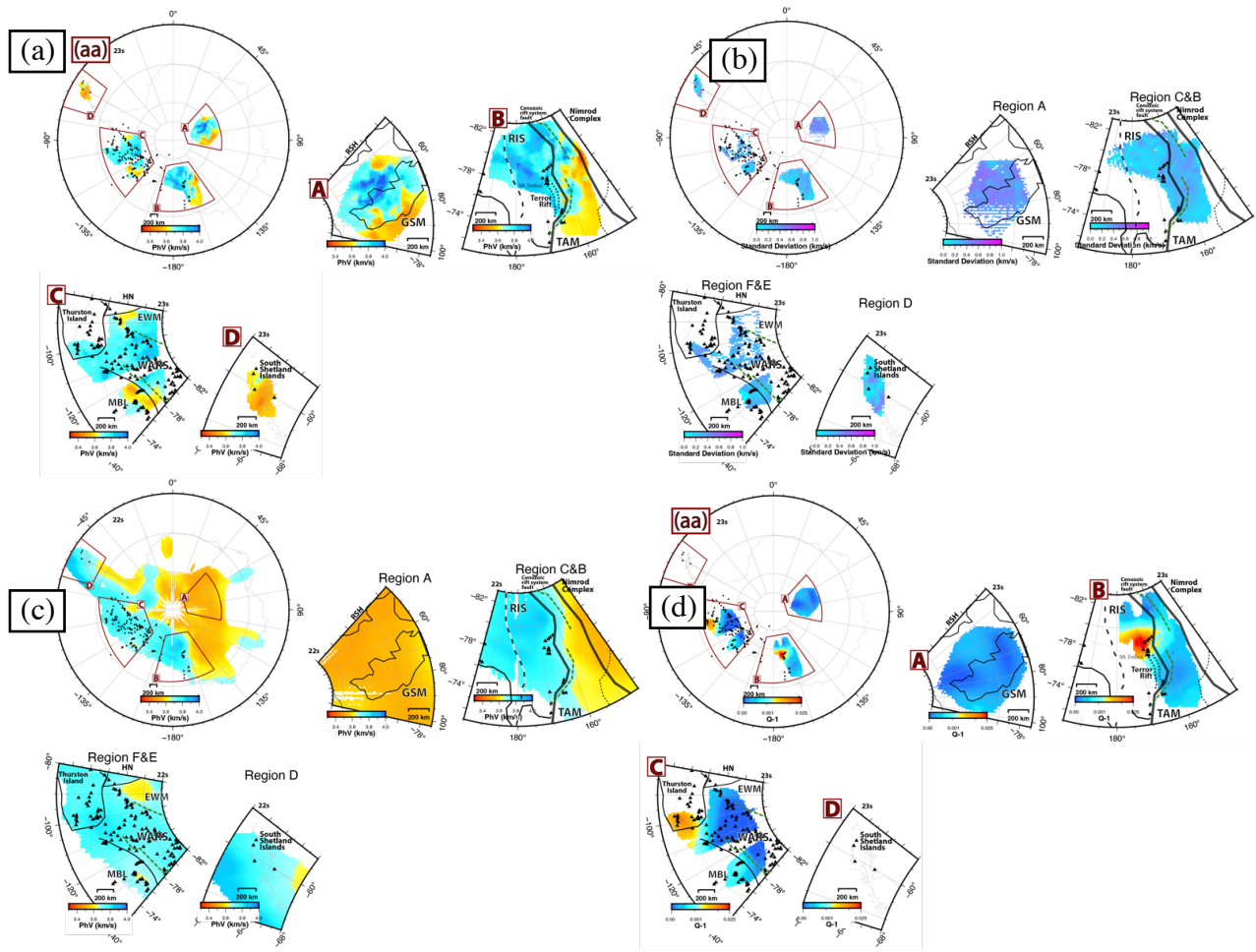


Figure S C-2. Results for 23s period

Results for 23s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

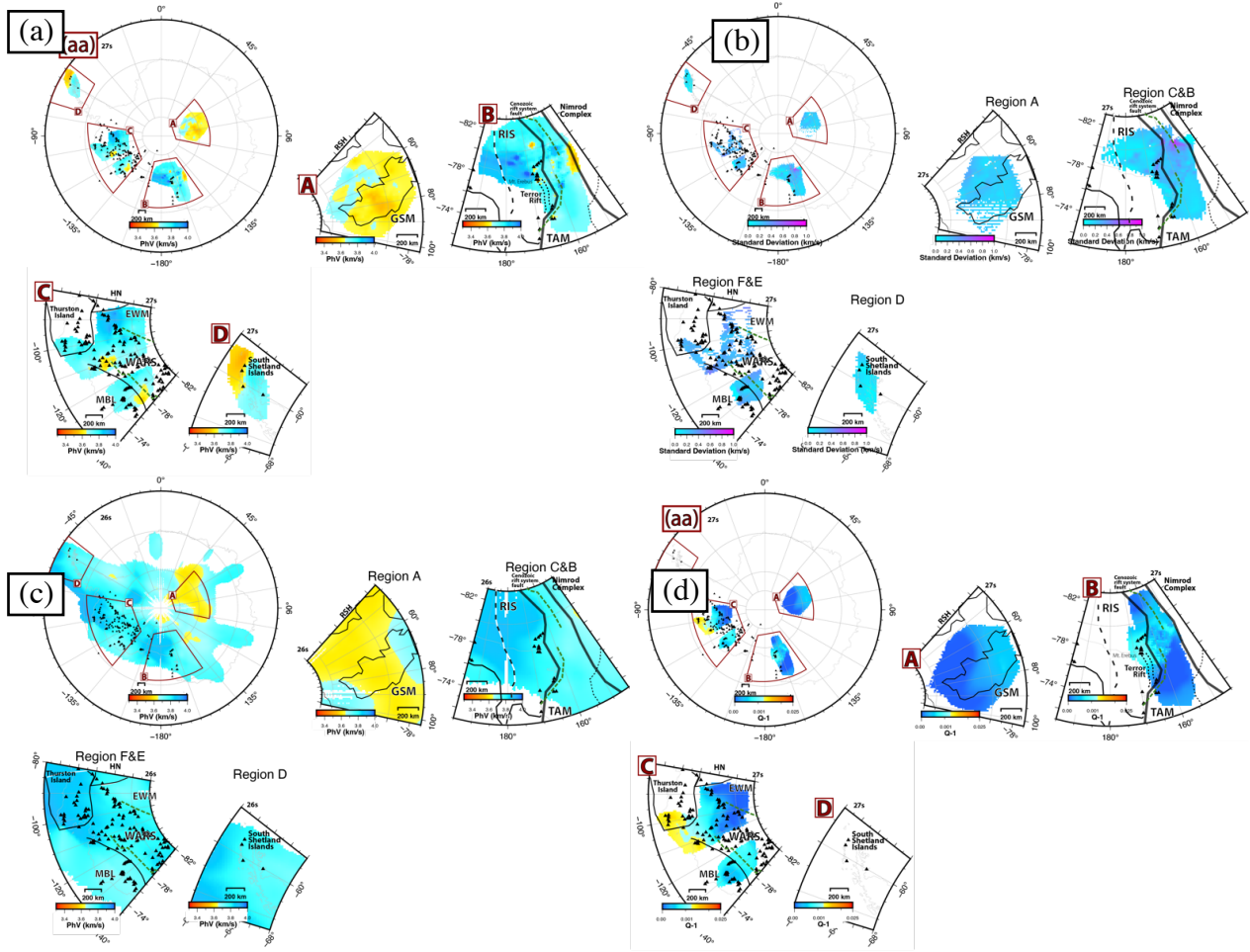


Figure S C-3. Results for 27s period

Results for 27s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

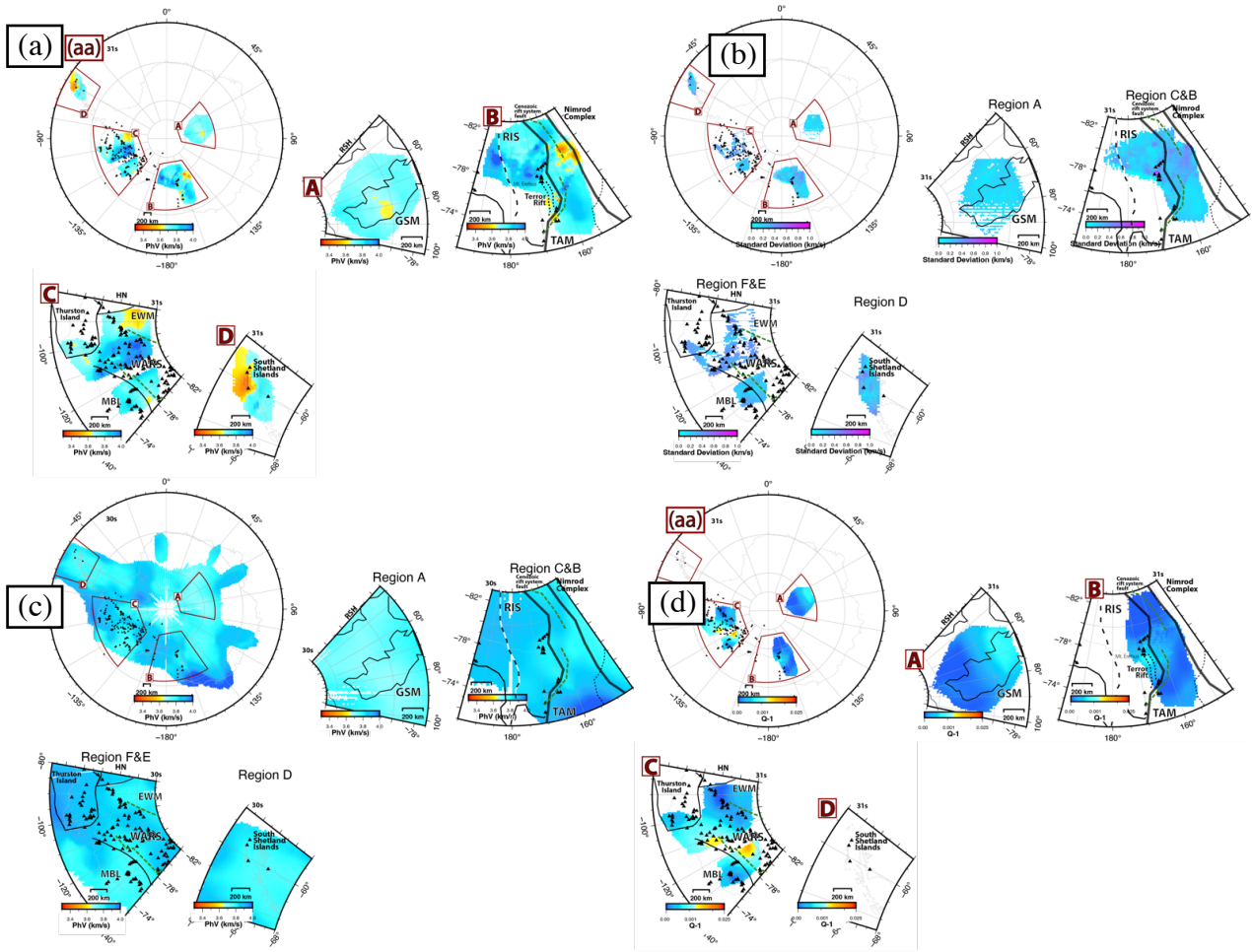


Figure S C-4. Results for 31s period

Results for 31s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

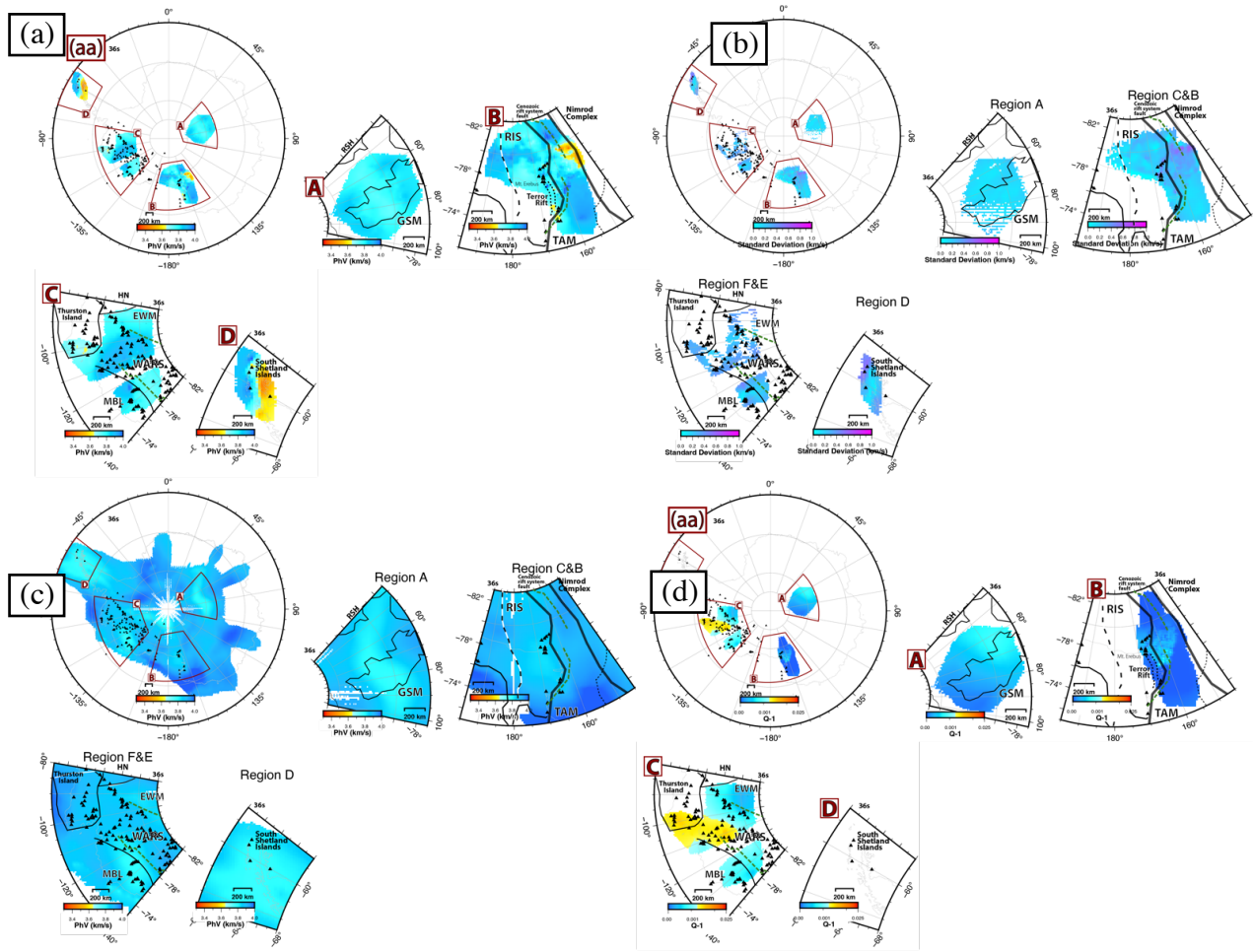


Figure S C-5. Results for 36s period

Results for 36s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

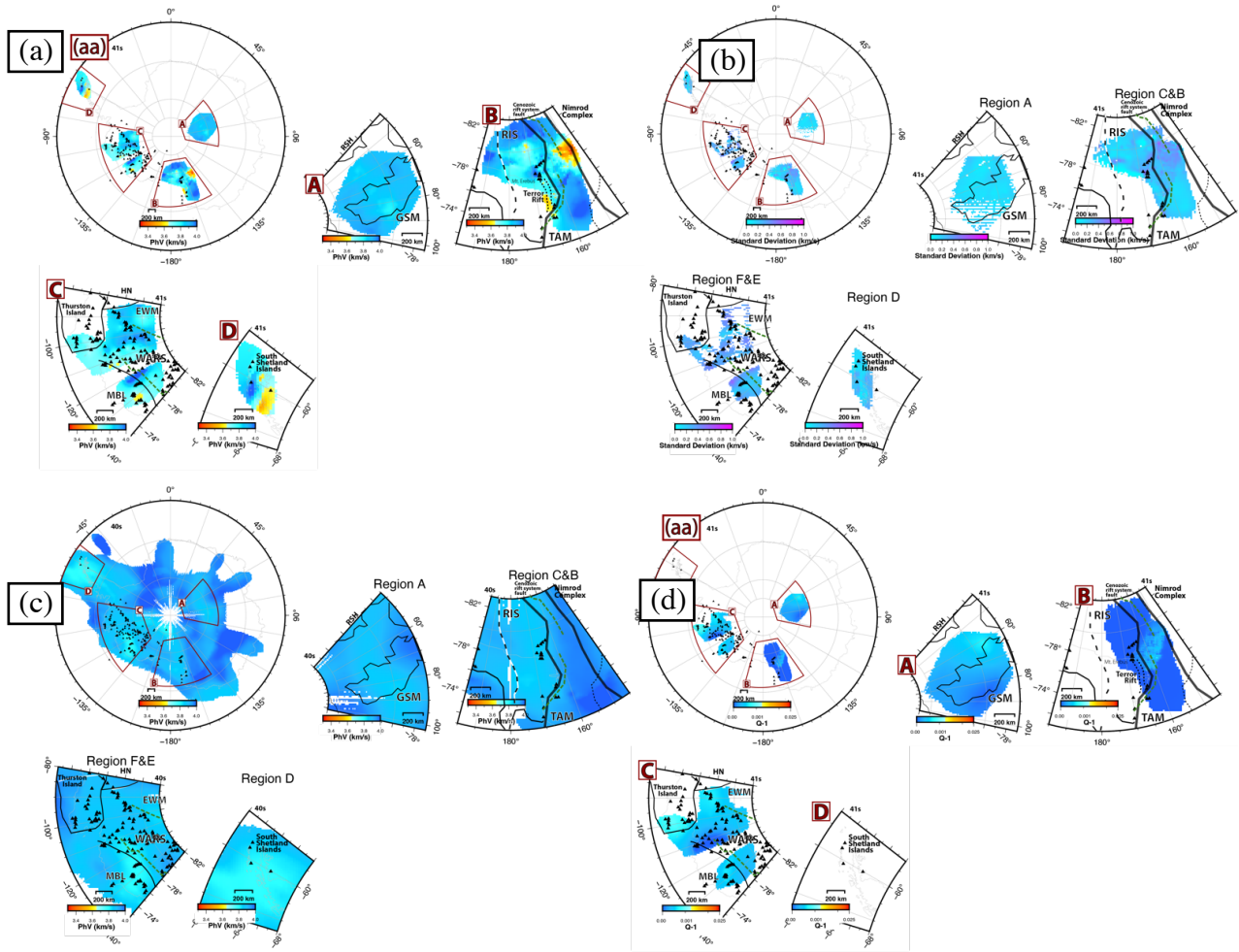


Figure S C-6. Results for 41s period

Results for 41s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

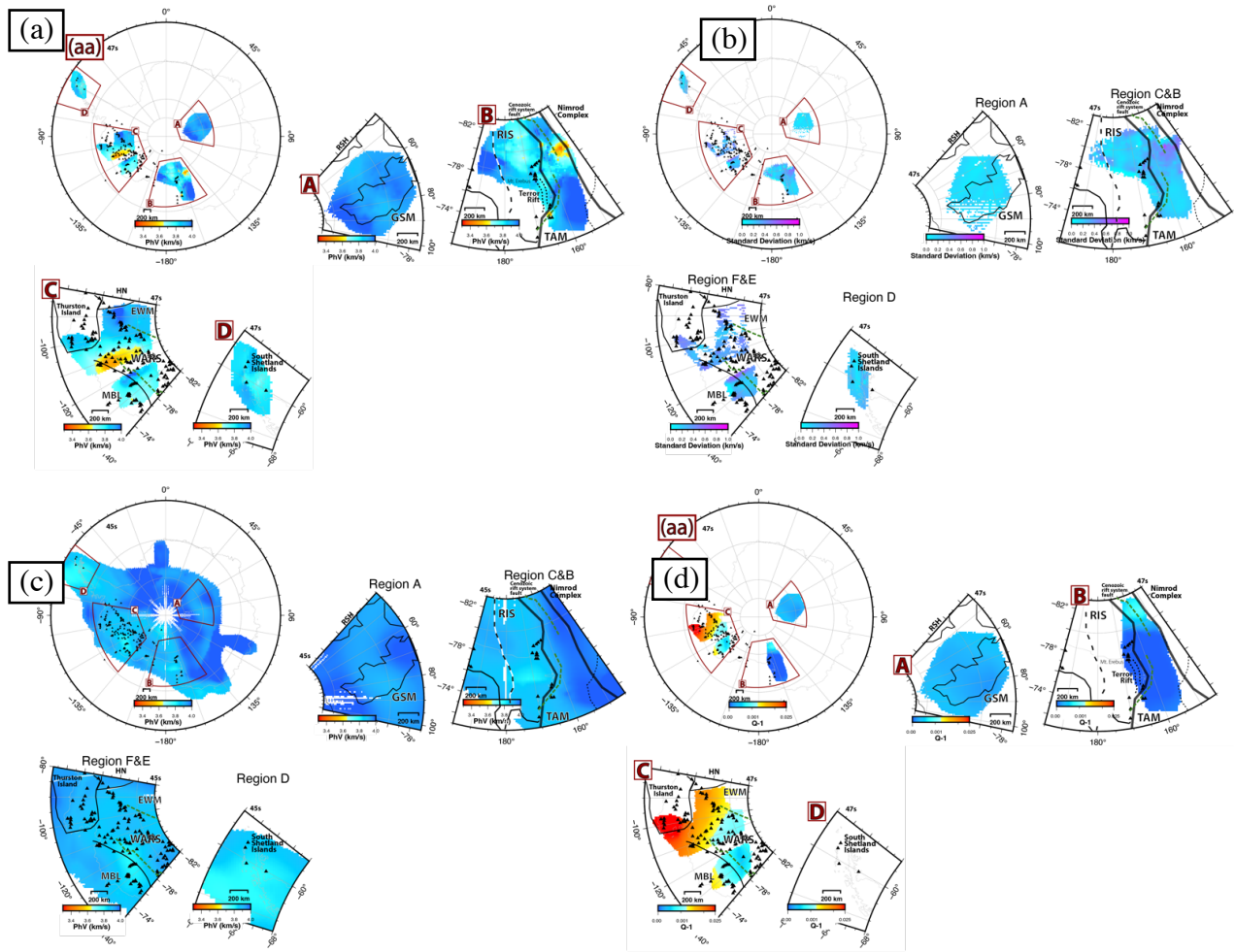


Figure S C-7. Results for 47s period

Results for 47s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

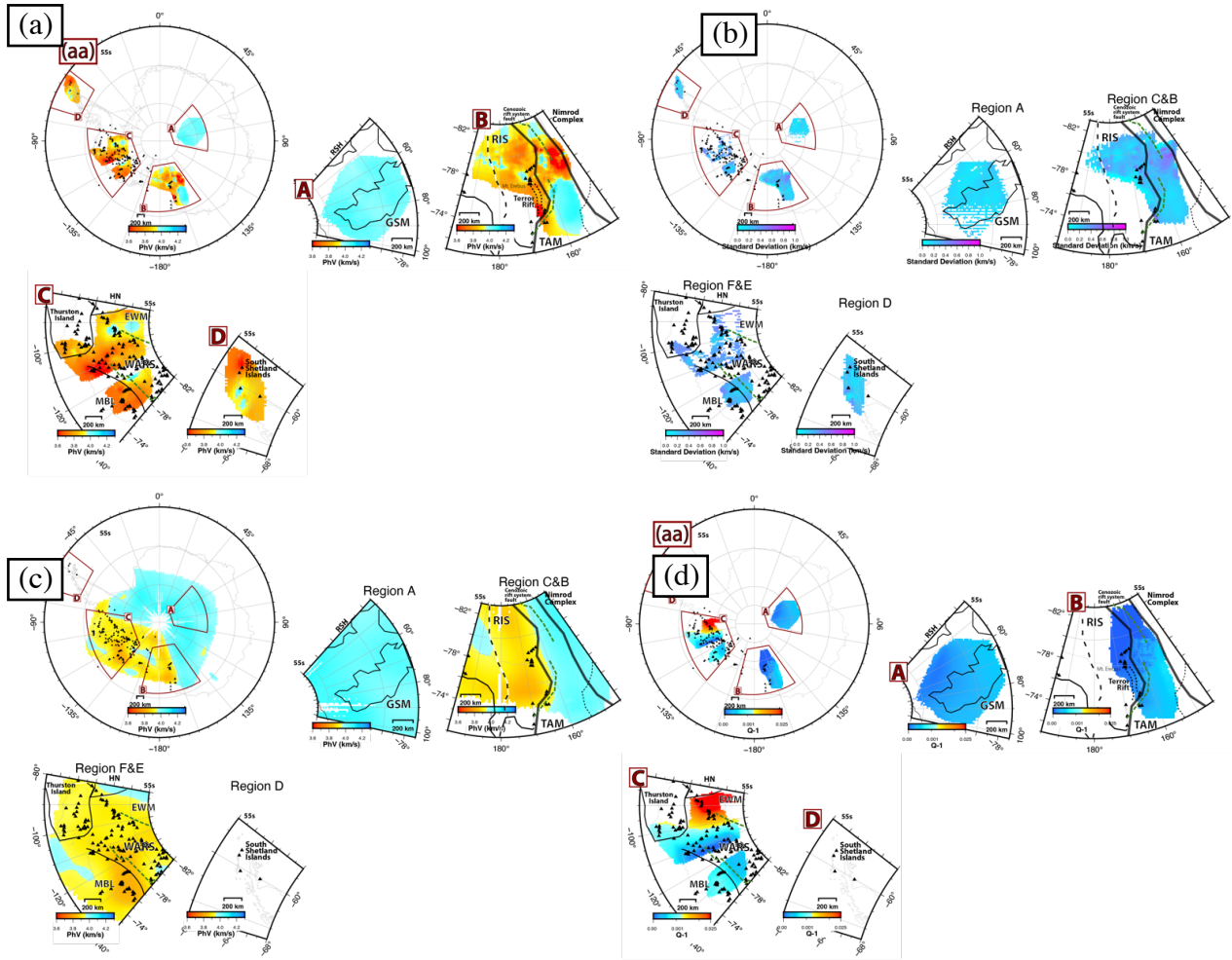


Figure S C-8. Results for 55s period

Results for 55s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

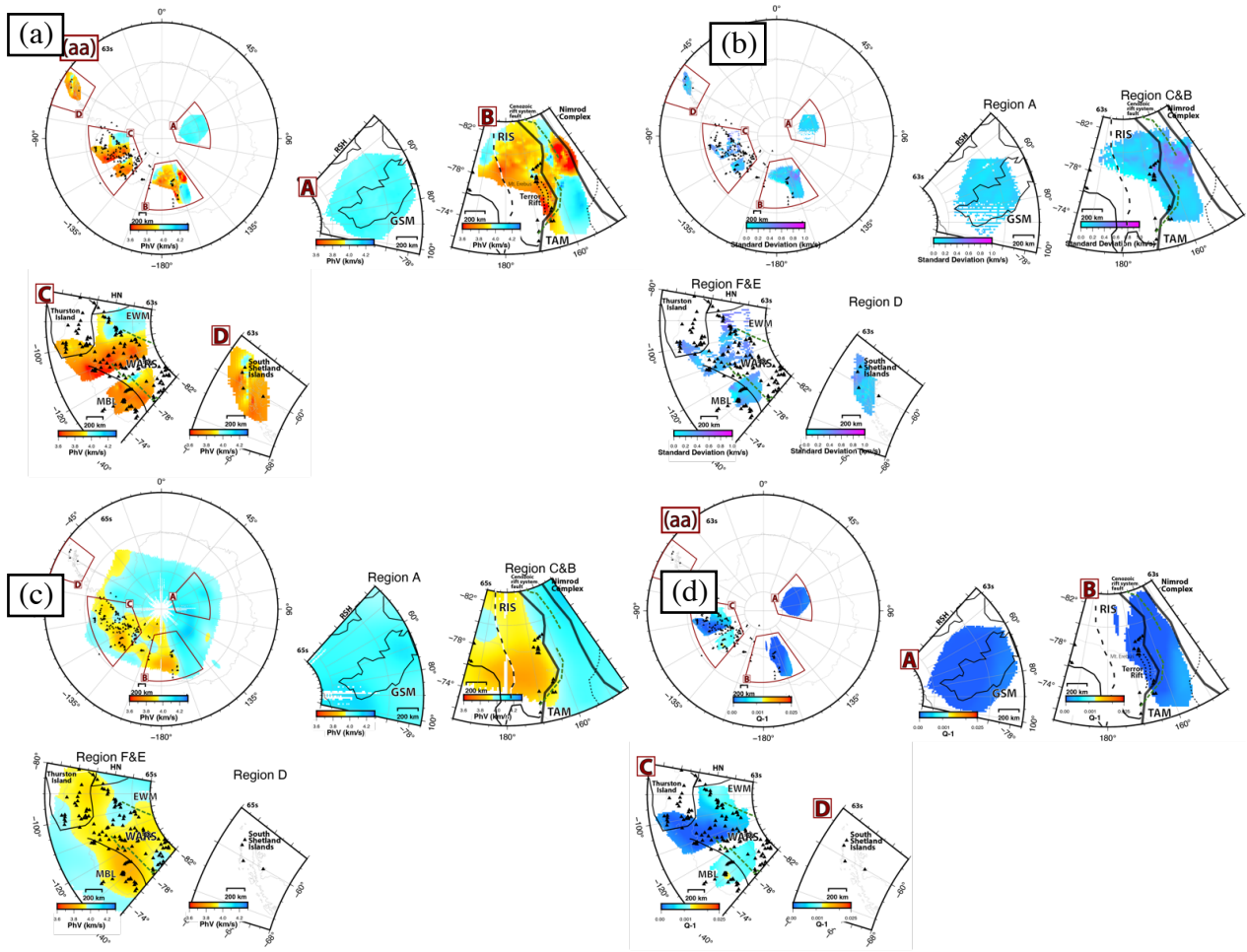


Figure S C-9. Results for 63s period

Results for 63s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

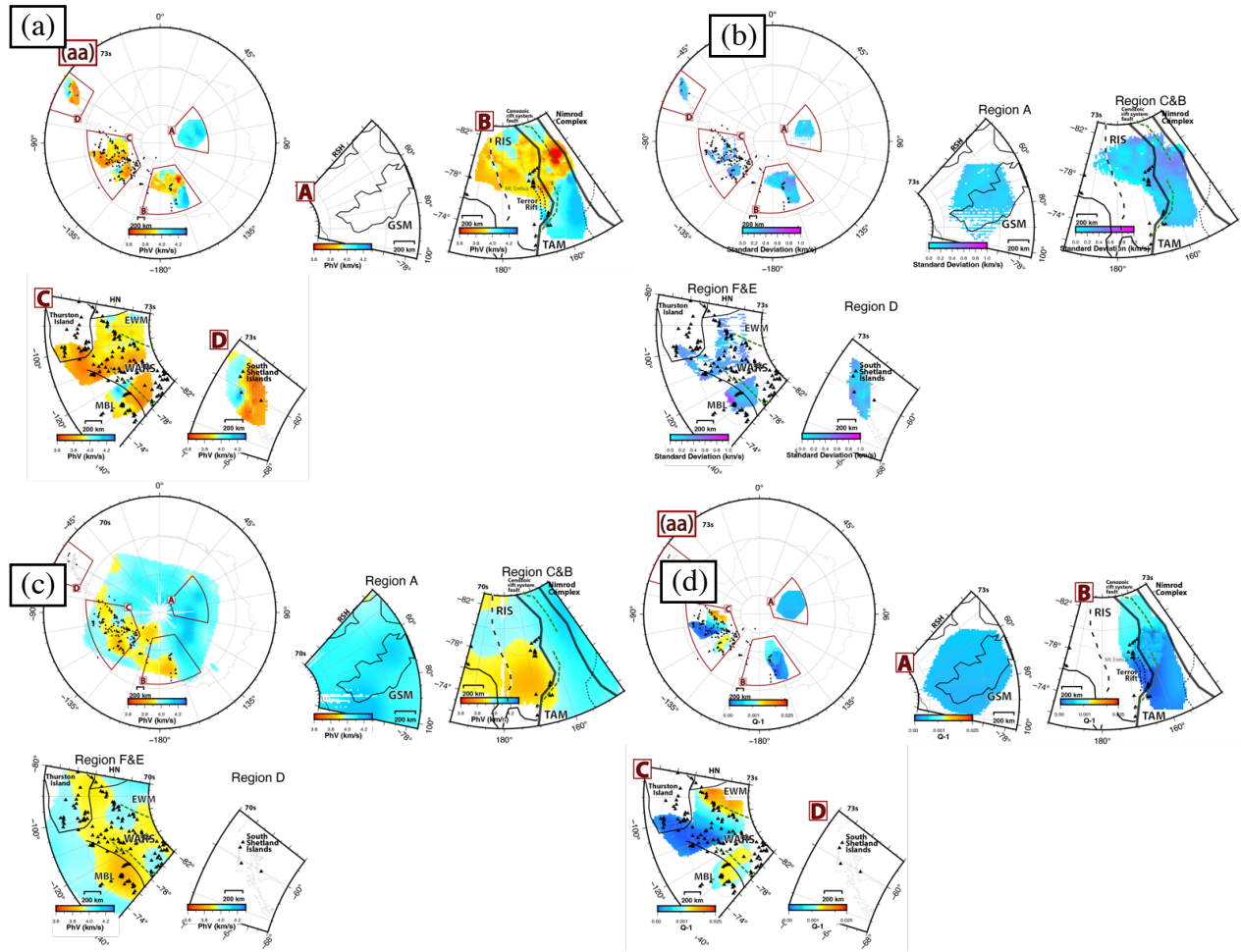


Figure S C-10. Results for 73s period

Results for 73s period: (a) Eikonal phase velocities, (b) standard deviations of phase velocities, (c) Shen (2018) phase velocities, (d) attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

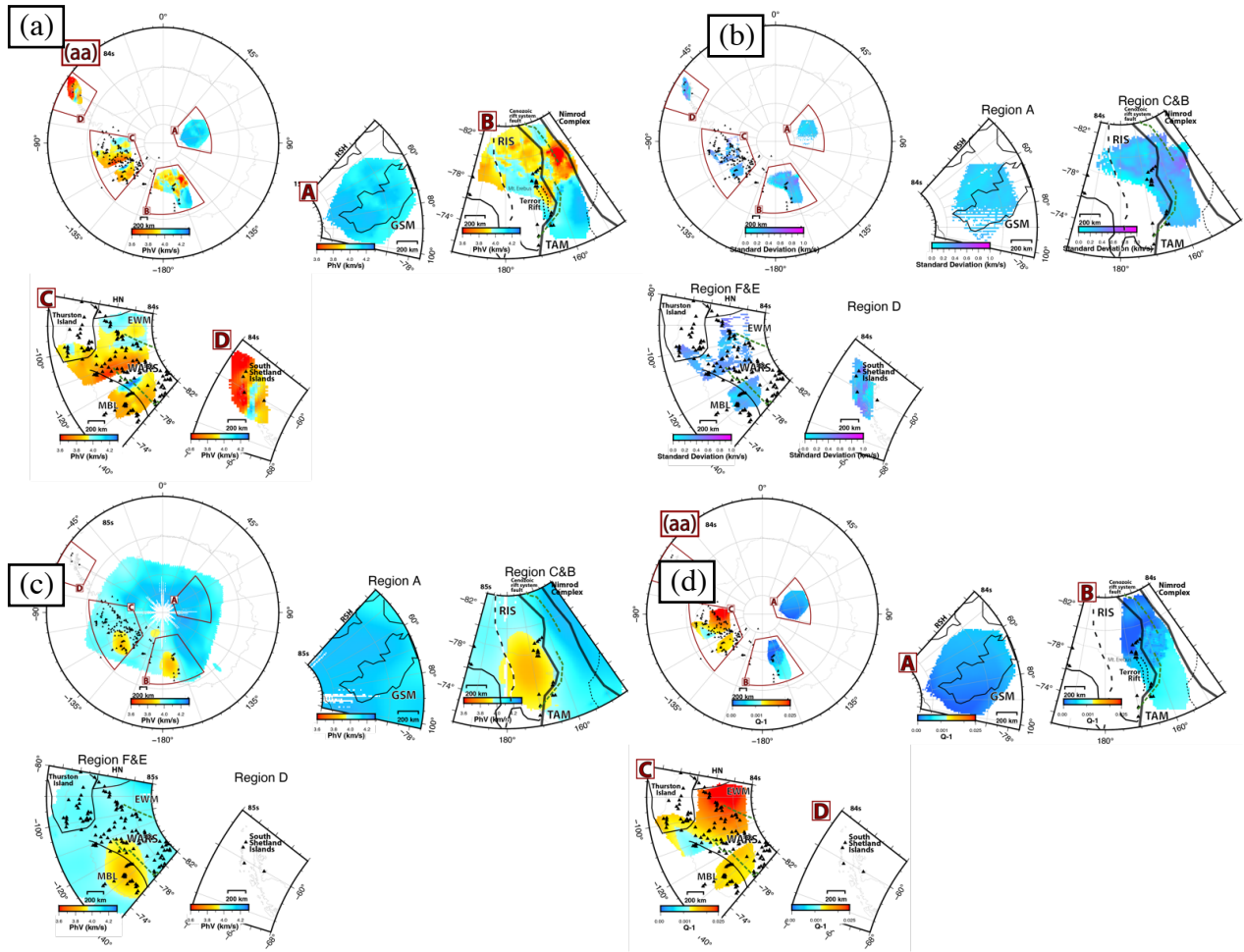


Figure S C-11. Results for 84s period

Results for 84s period: (a) Eikonal phase velocities, (b) standard deviations of phase velocities, (c) Shen (2018) phase velocities, (d) attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

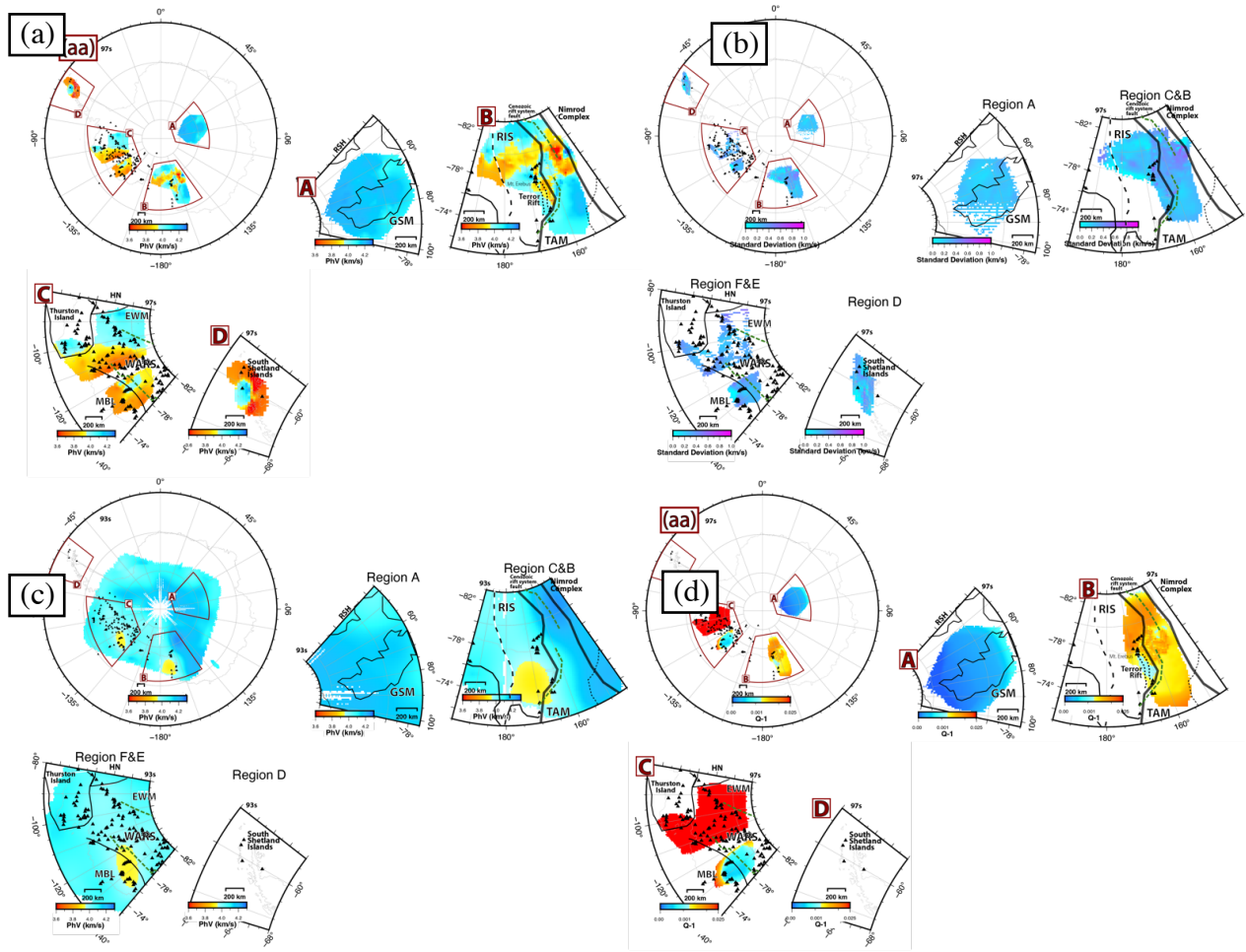


Figure S C-12. Results for 97s period

Results for 97s period: (a) Eikonal phase velocities, (b) standard deviations of phase velocities, (c) Shen (2018) phase velocities, (d) attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

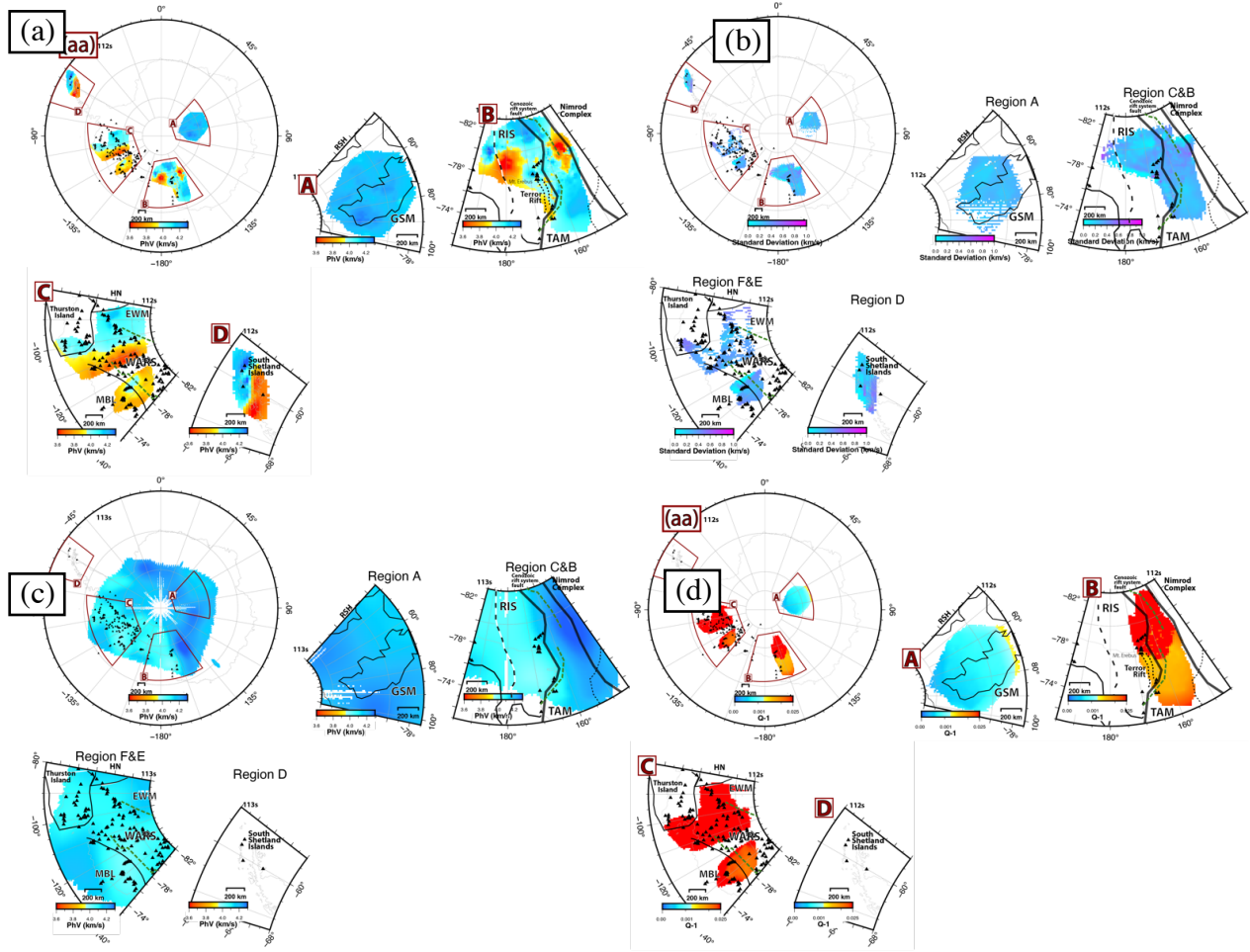


Figure S C-13. Results for 112s period

Results for 112s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

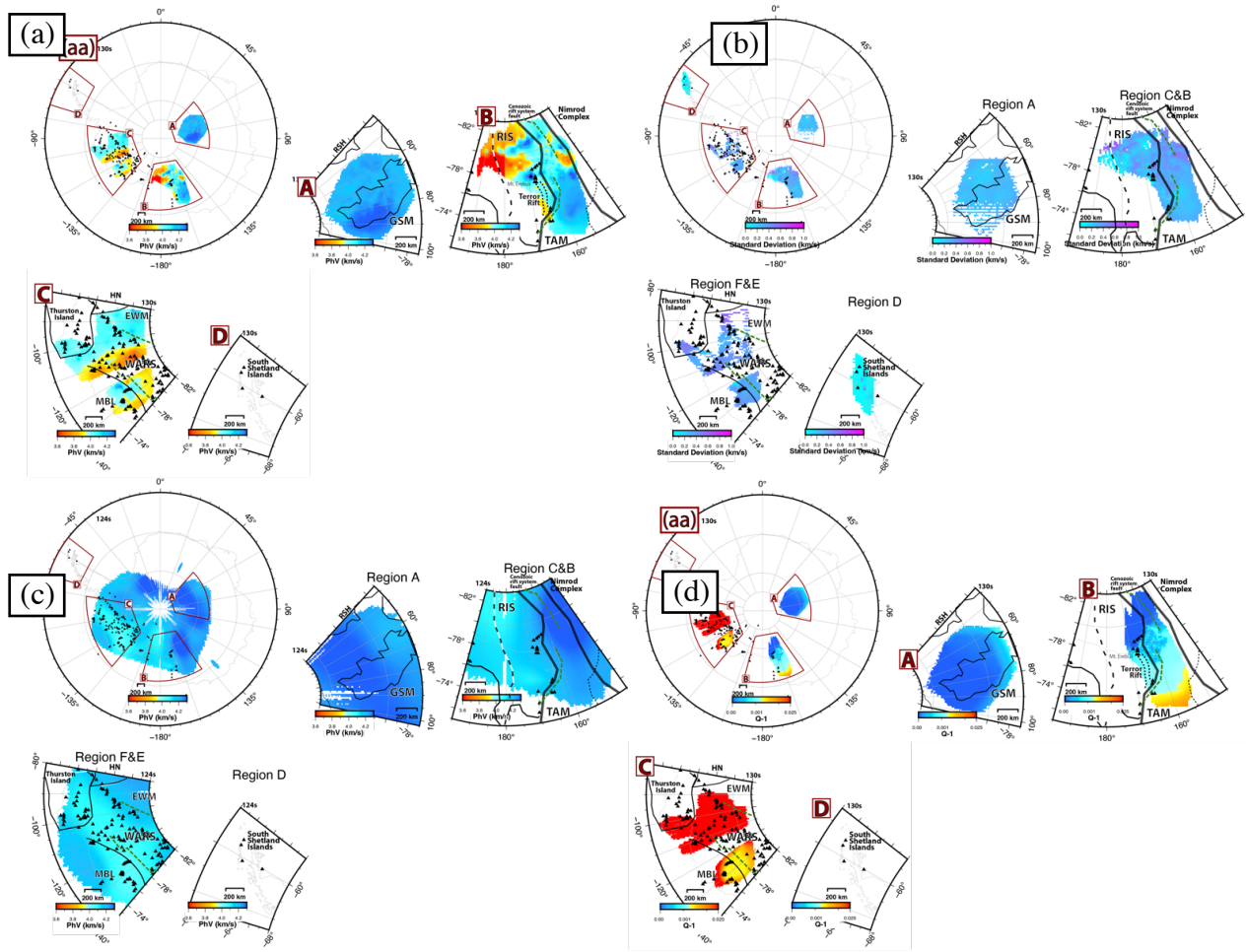


Figure S C-14. Results for 130s period

Results for 130s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D).

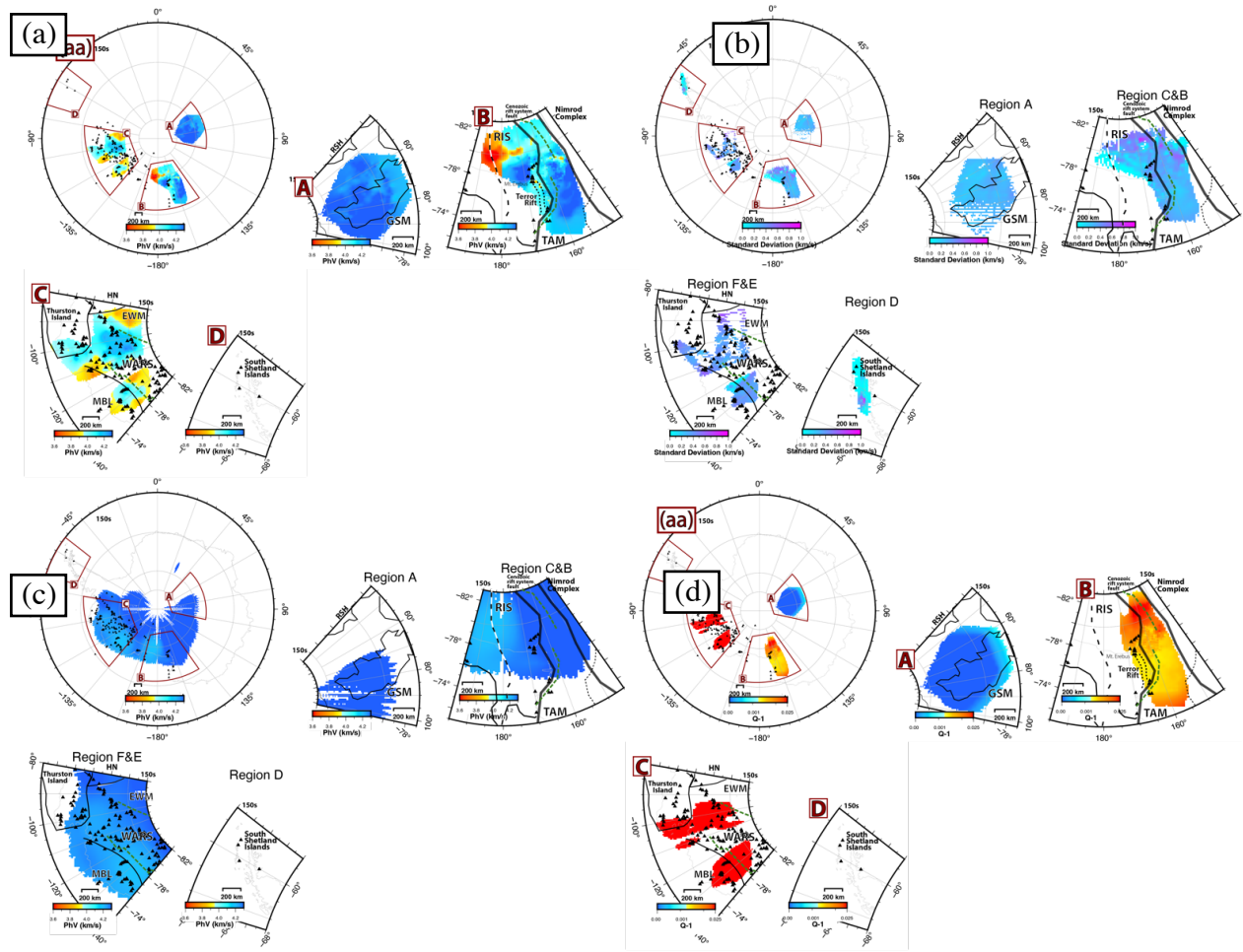


Figure S C-15. Results for 150s period

Results for 150s period: **(a)** Eikonal phase velocities, **(b)** standard deviations of phase velocities, **(c)** Shen (2018) phase velocities, **(d)** attenuation. Results provided with a global perspective (circle) and for all regions discussed in the manuscript (A-D)