

# **Joint Seismic Constraints on the Structure and Evolution of the Continental Lithosphere**

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This dissertation by Yiran Huang 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 April 20, 2026



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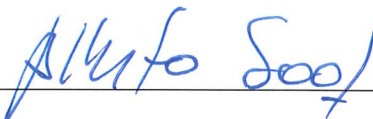
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- Huang, Y., Dalton, C. A. Estimating the local amplification of Rayleigh waves with regional broadband seismic arrays. In prep.
- Huang, Y., Dalton, C. A., & Hariharan, A. (2025). A new approach to constrain crustal Vp/Vs from Rayleigh wave phase velocity and local amplification: Application to the western US. *Geophysical Research Letters*, 52. <https://doi.org/10.1029/2024GL111980>
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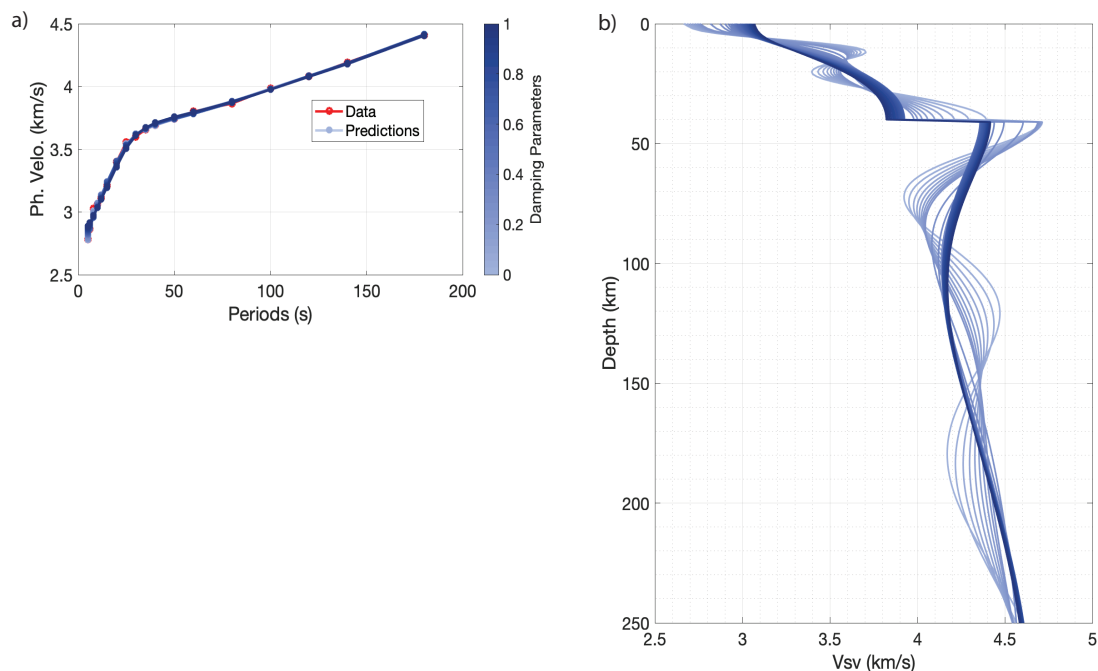
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## CHAPTER 1| Introduction

The thermal and chemical structure of the Earth plays a fundamental role in its formation, evolution, and present-day dynamics, controlling processes from the deep interior to the near-surface. Seismology provides powerful constraints on temperature, composition, and volatile distribution, offering key insights into Earth's internal state and its evolution history. For example, because shear modulus decreases strongly with increasing temperature, shear wave velocity ( $V_s$ ) models can be used to infer thermal structure and therefore lithosphere thickness (e.g., Priestley and McKenzie 2006), with important implications for global plate dynamics. In addition, empirical relationships between seismic velocities and chemical compositions allow  $V_s$  and  $V_p/V_s$ , the ratio of compressional- and shear-wave velocities, to be used to estimate crustal composition ( $\text{SiO}_2$  wt%), with implications for continental evolution (e.g., Sui et al. 2022). Furthermore, since partial melt also reduces shear modulus, sharp decreases in seismic velocity have been interpreted as evidence for melt in the upper mantle, which may influence deformation within the lithosphere–asthenosphere system (Rychert et al. 2020). These examples highlight the critical role of seismic observations in probing Earth's interior, which is otherwise largely inaccessible.

Despite many advances over several decades of study, several challenges limit our ability to interpret seismic observations. Among these, we highlight three that are most relevant to this thesis. First, seismic data measurements contain uncertainties in both phase and amplitude, which propagate into surface-wave tomography. These uncertainties arise from multiple sources, such as instrument calibration, complex wave propagation

through heterogeneous structures, and surface wave overtone interference. Previous studies have made substantial effort to account for these error sources and improve seismic measurement accuracy (e.g., Lin and Ritzwoller 2011; Hariharan et al. 2020).



**Figure 1.3: Figure 1.** An example illustrating the challenges of using single seismic observable to constrain Earth structure. (a) Phase velocity dispersion curves. Red= data. Blue= Predictions from inverted  $V_{sv}$  models with varying damping parameters (color-coded). The predicted phase velocities fit the observations equally well. (b) Corresponding inverted  $V_{sv}$  models, color-coded by damping parameters.

Second, the inversion of seismic observations (e.g., phase velocity dispersion) to obtain a seismic model (e.g., shear wave velocity models) is inherently non-unique and depends on the inversion strategy and model parameterization, as shown in Fig 1. For example, surface wave phase velocities at different frequencies sample different depth ranges, which contributes to their dispersive nature. As shown in Fig 2.1a, shorter-period surface waves are more sensitive to shallower depths, whereas longer-period waves sample deeper structure. But their sensitivity is distributed broadly in depth, and the

resulting depth-dependent  $V_s$  model depends not only on the dispersion data but also on regularization choices and how key interfaces, such as the crust–mantle boundary, are parameterized. Moreover, assumptions such as scaling relationships between  $V_s$ ,  $V_p$  and density further influence the resulting seismic models. Third, the relationship between seismic structure and underlying physical properties is itself non-unique. Variations in temperature, composition, partial melt, and volatile content can all produce similar changes in elastic properties, complicating interpretation. For example, the correlation between how  $V_p/V_s$  varies with  $\text{SiO}_2$  content remains debated (Hacker et al. 2015), highlighting the difficulty of linking seismic observations to specific physical conditions using a single dataset.

To address the challenges described above, many studies have used multiple seismic datasets to better constrain Earth structure. For example, joint inversion of surface waves, which are sensitive to absolute shear wave velocity, and converted body waves, which are sensitive to vertical velocity changes, has been widely used to obtain improved shear-wave velocity models. By combining datasets with complementary sensitivities, these approaches provide stronger constraints on seismic structure than is possible with either data set alone (e.g., Julià et al. 2000; Shen et al. 2013; Eilon et al. 2018; Byrnes et al. 2023). In addition, seismic models derived from different observables can be integrated to infer physical properties of the Earth. For example, combining compressional and/or shear wave velocity with attenuation provides constraints on physical conditions of upper mantle, such as water content, temperature, melt fraction, and grain size (e.g., Ma et al. 2020; Eilon 2025).

Beyond seismic observations, constraints from other geophysical and geochemical disciplines are increasingly incorporated to improve interpretation. For example, gravity (e.g., Lowry and Pérez-Gussinyé 2011), heat flow and crustal heat generation (e.g., Sui et al. 2025), xenolith compositions (e.g., Shinevar et al. 2023), and magnetotelluric data (Moorkamp et al. 2019) are commonly used alongside seismic models to better constrain the physical state of the Earth’s interior, to reduce interpretational ambiguity arising from multiple contributing factors.

The work presented in this thesis builds on these approaches by jointly inverting multiple seismic datasets and integrating constraints from independent disciplines. Specifically, we combine Rayleigh wave phase and amplitude information (Chapter 2), and Rayleigh wave phase velocities with S-to-P receiver functions (Chapters 3 and 4), together with constraints from heat flow and xenolith data to better resolve lithospheric structure. This integrated framework is applied with a particular focus on cratonic regions, aiming to improve constraints on their thermal, compositional, and structural properties.

Cratons are broadly defined as tectonically stable continental regions formed in the Precambrian, characterized by thick, cold lithosphere and chemically depleted mantle compositions that help maintain long-term buoyancy despite low temperatures. In seismic models, cratonic lithosphere is typically represented as a high-velocity and thick “keel” in the uppermost mantle compared to younger continental or oceanic regions, as shown in some global studies (e.g., Dalton et al. 2017; Eaton et al. 2009; Fischer et al. 2020; Lee et al. 2011). However, increasingly detailed seismic observations reveal that the structure of cratonic upper mantle is more complex than this simplified picture.

First, layered seismic velocity structures are widely reported in global cratons using a variety of seismic datasets, often characterized by negative velocity gradients at depths of ~60–110 km (e.g., Chen 2017; Rychert and Shearer 2009; Liu et al. 2023; Wirth and Long 2014; Abt et al. 2010). Although these features are sometimes intermittent (Krueger et al., 2021) and even elusive (Kind et al. 2020), numerous mechanisms have been proposed to explain their origin, including partial melting, compositional heterogeneity, anisotropy, and elastically accommodated grain-boundary sliding (e.g., Rader et al. 2015; Selway et al. 2015; Karato et al. 2015; Thybo and Perchuć 1997; Wirth and Long 2014; Fu et al. 2022). These interpretations highlight the complexity of cratonic lithosphere and its modification history. Second, the commonly assumed model of a cold, compositionally depleted lithosphere based on xenolith and heat flow constraints cannot be fully reconciled with seismic observations, both in terms of velocity gradients and absolute values (e.g., Garber et al. 2018; Hirsch et al. 2015; Bruneton et al. 2004; Pedersen et al. 2009). This discrepancy suggests that either the thermal and/or compositional structure is more complex than previously assumed, or that improved seismic models are required to capture these features. Third, global variability in cratonic lithosphere has been documented not only by seismology (e.g., Eeken et al. 2018; Boyce et al. 2024), but also by xenolith studies (e.g., Lee et al. 2011) and heat flow observations (e.g., Jaupart and Mareschal 2015). These variations reflect diverse formation histories and subsequent tectonic and metasomatic modification processes.

Motivated by these challenges, this thesis aims to better constrain cratonic lithospheric structure and its underlying origin and evolution through the development of improved seismic models. This is achieved by jointly inverting multiple seismic datasets



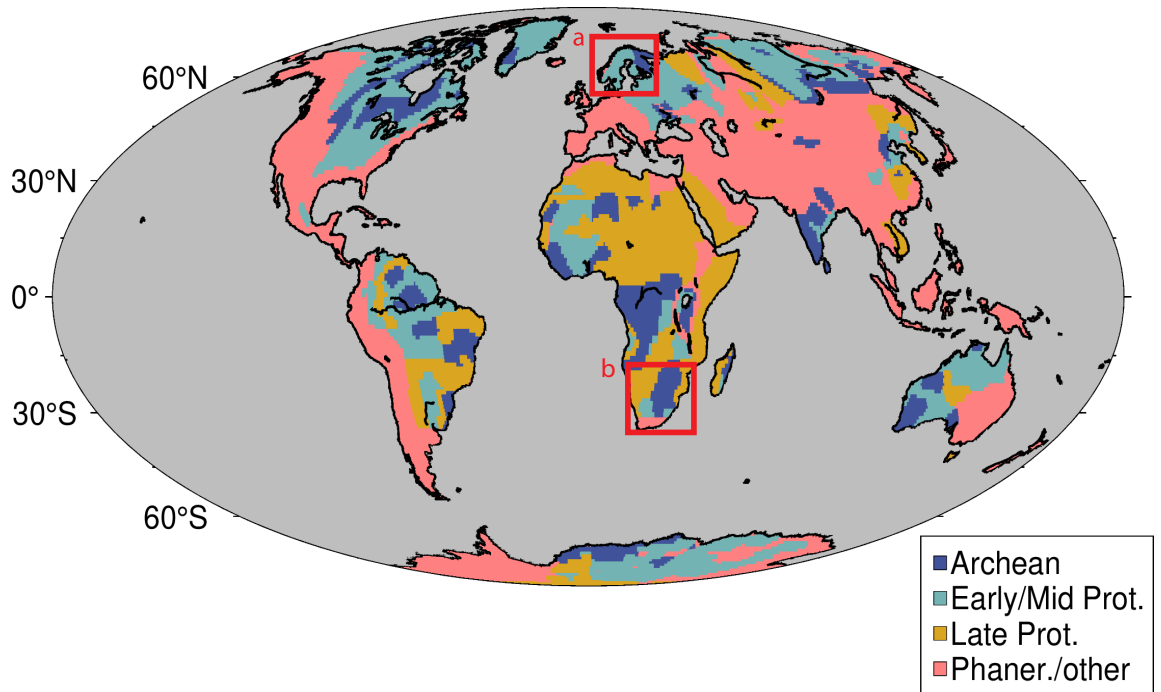
and integrating constraints from other disciplines, enabling a more consistent and robust interpretation of lithospheric thermal, compositional, and structural properties across different cratonic regions.

Chapter 2 presents the first study to infer crustal  $V_p/V_s$  in continental settings only using surface wave observations. By using the complementary sensitivities of Rayleigh wave phase velocity and amplitude, we develop a new joint inversion approach to constrain both shear-wave velocity ( $V_{sv}$ ) in the crust and uppermost mantle and crustal  $V_p/V_s$ . The method is validated through synthetic experiments and applied to real data in the western U.S. Although the absolute crustal  $V_p/V_s$  values depend somewhat on the choice of reference station and assumptions about the elastic structure beneath it, we show that relative variations are robustly recovered. The resulting  $V_p/V_s$  map in the western U.S. is consistent, at first order, with previous studies and correlates well with major tectonic features. These variations provide insights into crustal composition and its tectonic evolution. This method therefore offers strong potential for estimating crustal  $V_p/V_s$  using surface waves, particularly in regions where traditional approaches, such as receiver functions or local earthquake tomography, are not feasible.

However, this study also highlights challenges in measuring Rayleigh wave amplification, which is important because differences in amplitude datasets can lead to variations in the inferred  $V_p/V_s$  models. The unreliable amplification measurement is an issue that may be more significant in regional seismic networks. To address this, the Appendix presents an improved approach for estimating local Rayleigh wave amplification using regionally distributed, non-temporally overlapping broadband seismic arrays, with application to Fennoscandia. Building on the amplitude ratio method of Eddy

and Ekström (2014), we isolate the receiver effects and identify both constant and time-dependent erroneous instrument response through different references. The resulting amplification measurements show strong correlations with known tectonic and geological features and are consistent with predictions from independent regional models in Fennoscandia.

Chapters 3 and 4 present new constraints on the structure and evolution of cratonic lithosphere through developing consistent seismic models and comparative analysis of two Archean cratons: Fennoscandia and southern Africa (locations shown in Fig 2). Both chapters use a transdimensional Bayesian joint inversion of Rayleigh wave phase velocities and S-to-P receiver functions to resolve shear-wave velocity ( $V_{sv}$ ) structure with improved sensitivity to both absolute velocities and vertical gradients.



*Figure 1.4: Figure 2.* Age classification of continental crust from the Crust1.0 model (Laske et al., 2013). Red boxes outline the two study regions in this thesis: (a) Fennoscandia and (b) Southern Africa.

Chapter 3 focuses on the Fennoscandian craton and reveals a layered lithospheric structure for the craton interior, characterized by a high-velocity layer at depths of ~140–210 km. Positive velocity gradient in Sp receiver functions between ~140–190 km pushes the high-velocity lid in the Vsv model to deeper depth compared with a model derived from phase velocity only. To interpret these features, we systematically compare seismic observations with thermodynamic forward models based on xenolith-derived compositions and geotherms. We show that we can reconcile the high velocities observed in the lower lithosphere (~>150 km) with a depleted peridotite composition and xenolith-derived geotherm if that geotherm is corrected for thermometer bias and conductive cooling and a moderate diamond fraction (~0.5 vol.%) is allowed. In contrast, the reduced Vsv in the shallow mantle are best explained by a combination of radial anisotropy and the presence of carbonate or hydrous minerals introduced through metasomatism. This chapter provides evidence for at least a two-layered lithosphere and offers new insights into the thermal and compositional architecture of stable continental interiors through metasomatism.

Chapter 4 extends this framework to the southern Africa craton, enabling a direct comparison with a thermally and tectonically modified cratonic system. While both the Fennoscandia and southern Africa cratons exhibit similar shallow lithospheric structures above ~150 km, southern Africa is distinguished from Fennoscandia by a pronounced low-velocity zone between ~150–200 km. We interpret these low velocities as evidence for thermally perturbed lithosphere, which is also associated with observations of elevated surface topography and Cretaceous kimberlite magmatism and is supported by geochemical constraints and geodynamic models. By integrating these regional-scale

observations with global craton comparisons, we demonstrate that variations in lithospheric thermal state play a key role in governing craton evolution and surface uplift, shaping the craton into its present state.

This thesis demonstrates that the lithosphere is a vertically layered and dynamically evolving system, rather than a simple, homogeneous structure. By integrating complementary seismic observations with constraints from other disciplines, this work provides new insights into lithospheric composition, thermal state, and their implication in craton evolution history.

## 1.1 References

Abt, David L., Karen M. Fischer, Scott W. French, Heather A. Ford, Huaiyu Yuan, and

Barbara Romanowicz. 2010. “North American Lithospheric Discontinuity Structure Imaged by *Ps* and *Sp* Receiver Functions.” *Journal of Geophysical Research: Solid Earth* 115 (B9): 2009JB006914.

<https://doi.org/10.1029/2009JB006914>.

Boyce, Alistair, Thomas Bodin, Stéphanie Durand, Dorian Soergel, and Eric Debayle.

2024. “Seismic Evidence for Craton Formation by Underplating and Development of the MLD.” *Geophysical Research Letters* 51 (4): e2023GL106170.

<https://doi.org/10.1029/2023GL106170>.

Bruneton, Marianne, Helle A. Pedersen, Pierre Vacher, et al. 2004. “Layered Lithospheric Mantle in the Central Baltic Shield from Surface Waves and Xenolith Analysis.”

*Earth and Planetary Science Letters* 226 (1–2): 41–52.

<https://doi.org/10.1016/j.epsl.2004.07.034>.

Byrnes, Joseph, James Gaherty, and Emily Hopper. 2023. “Seismic Architecture of the Lithosphere-Asthenosphere System in the Western United States from a Joint Inversion of Body- and Surface-Wave Observations: Distribution of Partial Melt in the Upper Mantle.” *Seismica* 2 (2). <https://doi.org/10.26443/seismica.v2i2.272>.

Chen, Ling. 2017. “Layering of Subcontinental Lithospheric Mantle.” *Science Bulletin* 62 (14): 1030–34. <https://doi.org/10.1016/j.scib.2017.06.003>.

Dalton, Colleen A., Xueyang Bao, and Zhitu Ma. 2017. “The Thermal Structure of Cratonic Lithosphere from Global Rayleigh Wave Attenuation.” *Earth and Planetary Science Letters* 457 (January): 250–62. <https://doi.org/10.1016/j.epsl.2016.10.014>.

Eaton, David W., Fiona Darbyshire, Rob L. Evans, Herman Grütter, Alan G. Jones, and Xiaohui Yuan. 2009. “The Elusive Lithosphere–Asthenosphere Boundary (LAB) beneath Cratons.” *Lithos* 109 (1–2): 1–22. <https://doi.org/10.1016/j.lithos.2008.05.009>.

Eddy, Celia L., and Göran Ekström. 2014. “Local Amplification of Rayleigh Waves in the Continental United States Observed on the USArray.” *Earth and Planetary Science Letters* 402 (September): 50–57. <https://doi.org/10.1016/j.epsl.2014.01.013>.

- Eeken, Thomas, Saskia Goes, Helle A. Pedersen, Nicholas T. Arndt, and Pierre Bouilhol. 2018. "Seismic Evidence for Depth-Dependent Metasomatism in Cratons." *Earth and Planetary Science Letters* 491 (June): 148–59.  
<https://doi.org/10.1016/j.epsl.2018.03.018>.
- Eilon, Z. C. 2025. "Attenuation and Velocity Tomography of the Northern East African Rift." *Journal of Geophysical Research: Solid Earth* 130 (2): e2024JB030417.  
<https://doi.org/10.1029/2024JB030417>.
- Eilon, Zachary, Karen M. Fischer, and Colleen A. Dalton. 2018. "An Adaptive Bayesian Inversion for Upper-Mantle Structure Using Surface Waves and Scattered Body Waves." *Geophysical Journal International* 214 (1): 232–53.  
<https://doi.org/10.1093/gji/ggy137>.
- Fischer, Karen M., Catherine A. Rychert, Colleen A. Dalton, Meghan S. Miller, Caroline Beghein, and Derek L. Schutt. 2020. "A Comparison of Oceanic and Continental Mantle Lithosphere." *Physics of the Earth and Planetary Interiors* 309 (December): 106600. <https://doi.org/10.1016/j.pepi.2020.106600>.
- Fu, Hui-Ying, Zhong-Hai Li, and Ling Chen. 2022. "Continental Mid-Lithosphere Discontinuity: A Water Collector During Craton Evolution." *Geophysical Research Letters* 49 (23): e2022GL101569.  
<https://doi.org/10.1029/2022GL101569>.
- Garber, Joshua M., Satish Maurya, Jean-Alexis Hernandez, et al. 2018. "Multidisciplinary Constraints on the Abundance of Diamond and Eclogite in the

- Cratonic Lithosphere.” *Geochemistry, Geophysics, Geosystems* 19 (7): 2062–86.  
<https://doi.org/10.1029/2018GC007534>.
- Hacker, Bradley R., Peter B. Kelemen, and Mark D. Behn. 2015. “Continental Lower Crust.” *Annual Review of Earth and Planetary Sciences* 43 (1): 167–205.  
<https://doi.org/10.1146/annurev-earth-050212-124117>.
- Hariharan, Anant, Colleen A. Dalton, Zhitu Ma, and Göran Ekström. 2020. “Evidence of Overtone Interference in Fundamental-Mode Rayleigh Wave Phase and Amplitude Measurements.” *Journal of Geophysical Research: Solid Earth* 125 (1): e2019JB018540. <https://doi.org/10.1029/2019JB018540>.
- Hirsch, Aaron C., Colleen A. Dalton, and Jeroen Ritsema. 2015. “Constraints on Shear Velocity in the Cratonic Upper Mantle from Rayleigh Wave Phase Velocity.” *Geochemistry, Geophysics, Geosystems* 16 (11): 3982–4005.  
<https://doi.org/10.1002/2015GC006066>.
- Jaupart, C., and J. C. Mareschal. 2015. “Post-Orogenic Thermal Evolution of Newborn Archean Continents.” *Earth and Planetary Science Letters* 432 (December): 36–45. <https://doi.org/10.1016/j.epsl.2015.09.047>.
- Julià, J., C. J. Ammon, R. B. Herrmann, and A. M. Correig. 2000. “Joint Inversion of Receiver Function and Surface Wave Dispersion Observations.” *Geophysical Journal International* 143 (1): 99–112. <https://doi.org/10.1046/j.1365-246x.2000.00217.x>.

- Karato, Shun-ichiro, Tolulope Olugboji, and Jeffrey Park. 2015. “Mechanisms and Geologic Significance of the Mid-Lithosphere Discontinuity in the Continents.” *Nature Geoscience* 8 (7): 509–14. <https://doi.org/10.1038/ngeo2462>.
- Laske, Gabi, Guy Masters, Zhitu Ma, and Mike Pasyanos. 2013. “Update on CRUST1.0 - A 1-Degree Global Model of Earth’s Crust.” April 1, EGU2013-2658. <https://ui.adsabs.harvard.edu/abs/2013EGUGA..15.2658L>.
- Lee, Cin-Ty A., Peter Luffi, and Emily J. Chin. 2011. “Building and Destroying Continental Mantle.” *Annual Review of Earth and Planetary Sciences* 39 (1): 59–90. <https://doi.org/10.1146/annurev-earth-040610-133505>.
- Lin, Fan-Chi, and Michael H. Ritzwoller. 2011. “Helmholtz Surface Wave Tomography for Isotropic and Azimuthally Anisotropic Structure: Helmholtz Surface Wave Tomography.” *Geophysical Journal International* 186 (3): 1104–20. <https://doi.org/10.1111/j.1365-246X.2011.05070.x>.
- Liu, Tianze, Emily J. Chin, and Peter Shearer. 2023. “Strong Physical Contrasts Across Two Mid-Lithosphere Discontinuities Beneath the Northwestern United States: Evidence for Cratonic Mantle Metasomatism.” *AGU Advances* 4 (6): e2023AV001014. <https://doi.org/10.1029/2023AV001014>.
- Lowry, Anthony R., and Marta Pérez-Gussinyé. 2011. “The Role of Crustal Quartz in Controlling Cordilleran Deformation.” *Nature* 471 (7338): 353–57. <https://doi.org/10.1038/nature09912>.



- Ma, Zhitu, Colleen A. Dalton, Joshua B. Russell, James B. Gaherty, Greg Hirth, and Donald W. Forsyth. 2020. “Shear Attenuation and Anelastic Mechanisms in the Central Pacific Upper Mantle.” *Earth and Planetary Science Letters* 536 (April): 116148. <https://doi.org/10.1016/j.epsl.2020.116148>.
- Moorkamp, Max, Stewart Fishwick, Richard J. Walker, and Alan G. Jones. 2019. “Geophysical Evidence for Crustal and Mantle Weak Zones Controlling Intra-Plate Seismicity – the 2017 Botswana Earthquake Sequence.” *Earth and Planetary Science Letters* 506 (January): 175–83. <https://doi.org/10.1016/j.epsl.2018.10.048>.
- Pedersen, H. A., S. Fishwick, and D. B. Snyder. 2009. “A Comparison of Cratonic Roots through Consistent Analysis of Seismic Surface Waves.” *Lithos* 109 (1–2): 81–95. <https://doi.org/10.1016/j.lithos.2008.09.016>.
- Priestley, K., and D. Mckenzie. 2006. “The Thermal Structure of the Lithosphere from Shear Wave Velocities.” *Earth and Planetary Science Letters* 244 (1–2): 285–301. <https://doi.org/10.1016/j.epsl.2006.01.008>.
- Rader, Erika, Erica Emry, Nicholas Schmerr, et al. 2015. “Characterization and Petrological Constraints of the Midlithospheric Discontinuity.” *Geochemistry, Geophysics, Geosystems* 16 (10): 3484–504. <https://doi.org/10.1002/2015GC005943>.
- Rychert, Catherine A., Nicholas Harmon, Steven Constable, and Shunguo Wang. 2020. “The Nature of the Lithosphere-Asthenosphere Boundary.” *Journal of*

*Geophysical Research: Solid Earth* 125 (10): e2018JB016463.

<https://doi.org/10.1029/2018JB016463>.

Rychert, Catherine A., and Peter M. Shearer. 2009. “A Global View of the Lithosphere-Asthenosphere Boundary.” *Science* 324 (5926): 495–98.

<https://doi.org/10.1126/science.1169754>.

Selway, Kate, Heather Ford, and Peter Kelemen. 2015. “The Seismic Mid-Lithosphere Discontinuity.” *Earth and Planetary Science Letters* 414 (March): 45–57.

<https://doi.org/10.1016/j.epsl.2014.12.029>.

Shen, Weisen, Michael H. Ritzwoller, and Vera Schulte-Pelkum. 2013. “A 3-D Model of the Crust and Uppermost Mantle beneath the Central and Western US by Joint Inversion of Receiver Functions and Surface Wave Dispersion.” *Journal of Geophysical Research: Solid Earth* 118 (1): 262–76.

<https://doi.org/10.1029/2012JB009602>.

Shinevar, William J., Eva M. Golos, Oliver Jagoutz, Mark D. Behn, and Robert D. Van Der Hilst. 2023. “Mantle Thermochemical Variations beneath the Continental United States through Petrologic Interpretation of Seismic Tomography.” *Earth and Planetary Science Letters* 602 (January): 117965.

<https://doi.org/10.1016/j.epsl.2022.117965>.

Sui, Siyuan, Weisen Shen, and Oliver S. Boyd. 2025. “A Crustal Thermal Model of the Conterminous United States Constrained by Multiple Data Sets: A Monte–Carlo

Approach.” *Geophysical Journal International* 241 (3): 1712–25.

<https://doi.org/10.1093/gji/ggaf118>.

Sui, Siyuan, Weisen Shen, Kevin Mahan, and Vera Schulte-Pelkum. 2022. “Constraining the Crustal Composition of the Continental U.S. Using Seismic Observables.” *GSA Bulletin*, ahead of print, December 5. <https://doi.org/10.1130/B36229.1>.

Thybo, H., and E. Perchuc. 1997. “The Seismic 8° Discontinuity and Partial Melting in Continental Mantle.” *Science* 275 (5306): 1626–29.  
<https://doi.org/10.1126/science.275.5306.1626>.

Wirth, Erin A., and Maureen D. Long. 2014. “A Contrast in Anisotropy across Mid-Lithospheric Discontinuities beneath the Central United States—A Relic of Craton Formation.” *Geology* 42 (10): 851–54. <https://doi.org/10.1130/G35804.1>.

# **CHAPTER 2| A New Approach to Constrain Crustal Vp/Vs From Rayleigh Wave Phase Velocity and Local Amplification: Application to the Western US**

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## Abstract

The ratio of compressional- and shear-wave velocities,  $V_p/V_s$ , in the continental crust is a useful proxy for silica abundance and partial melt. However, constraining crustal  $V_p/V_s$  remains challenging. Rayleigh wave site amplification is a valuable complement to phase velocity and can constrain crustal  $V_p$ . We develop a two-step approach to jointly invert phase velocity and amplification for  $V_p/V_s$  in the crust and  $V_{sv}$  in the crust and uppermost mantle. We apply this joint inversion to construct new crustal  $V_p/V_s$  models for the western US from three different amplification data sets. The models contain  $V_p/V_s$  values in the range 1.6–1.85 and variations that are consistent between the three models and with previous results, including a westward increase of  $V_p/V_s$  across the Central Basin and Range. Crustal  $V_p/V_s$  across three regions of the Basin and Range are explored to gain insight into crustal composition and enhance understanding of crustal evolution.

# 1 Introduction

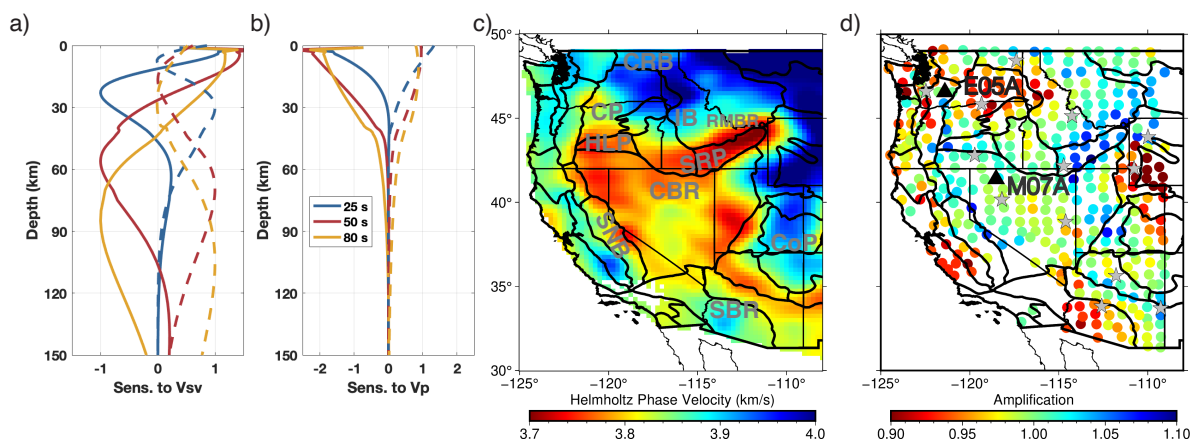
The ratio of compressional-wave and shear-wave velocities,  $V_p/V_s$ , is largely insensitive to temperature yet strongly sensitive to silica abundance and partial melt in the continental crust (e.g., Christensen, 1996; Lowry & Pérez-Gussinyé, 2011; Watanabe, 1993). It has been used to suggest that quartz-rich crust promotes deformation (Lowry & Pérez-Gussinyé, 2011), show how the lower crust of volcanic arcs transforms into lower continental crust (Jagoutz & Behn, 2013), demonstrate that silica enrichment controls the recurrence time of slow earthquakes (Audet & Bürgmann, 2014), and illuminate two unconnected upper-crustal magma reservoirs that feed volcanism at Mount St. Helens (Kiser et al., 2016). The  $V_p/V_s$  ratio has been used to estimate the composition of the continental crust (e.g., Sui et al., 2023; Zandt & Ammon, 1995) and volcanic arcs (Shillington et al., 2013), and sharp reductions in the crustal  $V_p/V_s$  ratio with depth have been employed to estimate crustal thermal gradients (e.g., Diaferia et al., 2019; Kuo-Chen et al., 2012).

However, the  $V_p/V_s$  ratio in the continental crust is challenging to constrain. H-k stacking with teleseismic P-to-S receiver functions (Zhu & Kanamori, 2000) is widely used due to its simplicity and because it requires only a single seismic station (e.g., Piana Agostinetti & Amato, 2009; Sun et al., 2012; Tugume et al., 2012). However, H- k stacking assumes a horizontal, isotropic, single-layer crustal model and a fixed crustal  $V_p$  to determine the average crustal thickness (H) and  $V_p/V_s$  (k), and error in  $V_p/V_s$  can occur where these assumptions fail. Adaptations to H-k stacking, such as incorporating density constraints (Lowry & Pérez-Gussinyé, 2011), correcting bias caused by radial anisotropy (Brunsvik & Eilon, 2023; Hammond, 2014), employing a two-layer stacking

method (Yeck et al., 2013), and incorporating S-to-P receiver functions (Rychert & Harmon, 2016) have helped to overcome these drawbacks. The  $V_p/V_s$  ratio can be estimated from  $P_n$  and  $S_n$  station terms (e.g., Buehler & Shearer, 2014; Haines, 1980), although crustal anisotropy, differences in  $P_n$  and  $S_n$  ray paths, and azimuthal averaging contribute error to such estimates (Buehler & Shearer, 2014). The crustal phases  $P_g$  and  $S_g$  can also be employed to estimate  $V_p/V_s$  in the upper crust where large volumes of data are available to suppress noise (Steck et al., 2011). A comparison of four recent crustal  $V_p/V_s$  models in the western United States (Figure S1 in Supporting Information S1) highlights the challenges of estimating  $V_p/V_s$ ; there are significant disparities in both the patterns and absolute values, even among models developed using similar methods.

Rayleigh wave phase velocities are routinely used to image the speeds of horizontally traveling, vertically polarized shear waves ( $V_{sv}$ ) in the crust and uppermost mantle. They are also sensitive to  $V_p$ , but it is difficult to constrain both  $V_p$  and  $V_{sv}$  with phase velocities alone. Most studies make assumptions about  $V_p$  rather than solve for it, even though incorrect assumptions about  $V_p$  can lead to erroneous  $V_{sv}$  results (Li et al., 2017; Figure S2 in Supporting Information S1). Rayleigh wave local site amplification can be a valuable complement to phase velocity. Its sensitivity to  $V_{sv}$  is focused at slightly shallower depths, and its relative sensitivity to  $V_p$  in the crust is larger than that of phase velocity (Figures 1a and 1b). Moreover, in contrast to phase, amplification is sensitive to elastic structure directly beneath a seismic station rather than integrated along a ray path. However, despite the acknowledged potential of amplification measurements to improve seismic models (e.g., Bowden et al., 2015, 2017; Eddy & Ekström, 2014, 2020; Lin et al., 2012), they are rarely used in inversions for elastic structure, with two

exceptions. Schardong et al. (2019) utilized Rayleigh wave amplification to solve for 3-D  $V_s$  variations in the upper mantle of the western US, and Sturgeon et al. (2023) expanded upon this by incorporating Love wave amplification to further constrain crustal  $V_s$  in the same region.



**Figure 2.1: Figure 1.** (a, b) The sensitivity of perturbations in Rayleigh wave amplification (solid lines) and phase velocity (dashed lines) to perturbations in  $V_{sv}$  and  $V_p$  at three periods, calculated using the one-dimensional velocity profile at station US.WVOR from Eilon et al. (2018) and normalized by the absolute maximum of the  $V_{sv}$  kernel of each data type at each period to convey the relative magnitude of sensitivity to  $V_p$  compared with sensitivity to  $V_{sv}$ . (c, d) Observations of Rayleigh wave. (c) Helmholtz phase velocity and (d) amplification at 60 s. Amplification values are plotted relative to station TA.M07A. The locations of normalization stations TA.M07A and TA.E05A used in Figure 4 are denoted by black triangles, and other stations in Figure 4 are denoted as gray stars. Black lines outline physiographic provinces after Fenneman (1928). CRB is Coast Range Batholith; CP is Columbia Plateau; IB is Idaho Batholith; RMBR is Rocky Mountains Basin and Range; SRP is Snake River Plain; CBR is Central Basin and Range; SBR is Southern Basin and Range; HLP is High Lava Plains; CoP is Colorado Plateau; SNB is Sierra Nevada Batholith.

In this study, we present a two-step approach to jointly invert Rayleigh wave phase velocity and amplification measurements for  $V_{sv}$  in the crust and uppermost mantle and  $V_p/V_s$  in the crust. This approach leverages the complementary sensitivities of amplification and phase velocity, facilitating a simultaneous inversion for the two seismic parameters. We demonstrate its utility in synthetic tests and apply this joint inversion



technique to real data to build a new crustal Vp/Vs model for the western US. We show which aspects of the crustal Vp/Vs model are well resolved, compare to previous studies, and provide a preliminary interpretation in terms of crustal composition.

## 2 Data and Methods

We solve for profiles of seismic velocity throughout the western US at depths extending from the base of sedimentary layers to 300 km, utilizing a combination of two distinct Rayleigh wave data types: phase velocity for periods 5–180 s and local amplification for periods 25–80 s. We use a more limited period range for the amplification data set in order to focus on the periods that we have the highest confidence in and that offer the greatest sensitivity to crustal Vp. For Rayleigh wave phase velocity, we utilize the USANT15 maps (Ekström, 2017) at 5–25 s and determine new maps of phase velocity (40–180 s) and amplification from 643 teleseismic earthquakes with a minimum magnitude  $M_w = 5.8$  that occurred during 2006–2010 and were recorded by EarthScope USArray stations (IRIS Transportable Array, 2003). We employ the Automated Surface Wave Measuring System (ASWMS) software (Jin & Gaherty, 2015) to measure phase delay and amplitude and also to calculate phase velocity maps with Helmholtz tomography (Lin & Ritzwoller, 2011). We also determine new maps of local amplification from Rayleigh wave amplitude measurements, corrected for the source radiation pattern (Hariharan et al., 2022), at USArray stations. We use the amplitude-ratio approach (Eddy & Ekström, 2014, 2020), which isolates station effects and minimizes source and propagation effects by using the ratios of amplitudes recorded at nearby stations for which it can be assumed that the waves have similar take-off angles

and ray paths. Eddy and Ekström (2020) showed, using synthetic seismograms, that with a large earthquake data set ( $>200$  events), these assumptions are valid and the amplitude-ratio approach can reliably measure amplification effects due to local elastic structure. The average number of events for the station pairs in our data set is 228. For each pair of stations, we determine the azimuthally weighted average amplitude ratio in order to account for non-uniform back-azimuthal coverage, and we solve for relative amplification at all stations using a least-squares inversion, fixing amplification to unity for a selected station as discussed below. The resulting Rayleigh wave phase velocity and amplification maps are displayed in Figure 1, Figures S3, and S4 in Supporting Information S1, and their uncertainty is presented in Figure S5 in Supporting Information S1. Text S1 and Figures S6–S8 in Supporting Information S1 contain a description of the amplification estimation method and a comparison with previous studies and phase velocity.

We implement a linear joint inversion method to determine, at each station,  $V_{sv}$  in the crust and uppermost mantle and  $V_p/V_s$  in the crust. Our approach involves a two-step strategy, the key aspects of which are summarized here, with additional details in Text S2 in Supporting Information S1. In the first step, we conduct a 15-iteration linear inversion, using Rayleigh wave phase velocity data alone, to constrain the  $V_{sv}$  structure in 1-km-thick layers. We assume the sedimentary structure from Mooney et al. (2023). In the first step, we force  $V_p$  in the crust to scale with  $V_{sv}$  according to the relationships of Brocher (2005) and crustal density to scale with  $V_{sv}$  using the relationship from Levandowski et al. (2014). In the mantle, we force  $V_p/V_{sv}$  scaling consistent with ak135 (Kennett et al.,

1995) and density/ $V_{sv}$  scaling for lherzolite from Abers and Hacker (2016). In practice, we solve for perturbations to a reference  $V_{sv}$  model that is updated at each iteration and impose a smoothness constraint that is loosened at the Moho, with Moho depth data sourced from Schmandt et al. (2015). In the second step, Rayleigh wave amplification and phase velocity data are jointly inverted to refine the  $V_{sv}$  model and estimate crustal  $V_p/V_s$ . The initial model for this step is the final  $V_{sv}$  model from Step 1, and we invert for a perturbation to the reference crustal  $V_p/V_s$  as a single layer as well as perturbations to  $V_{sv}$ .

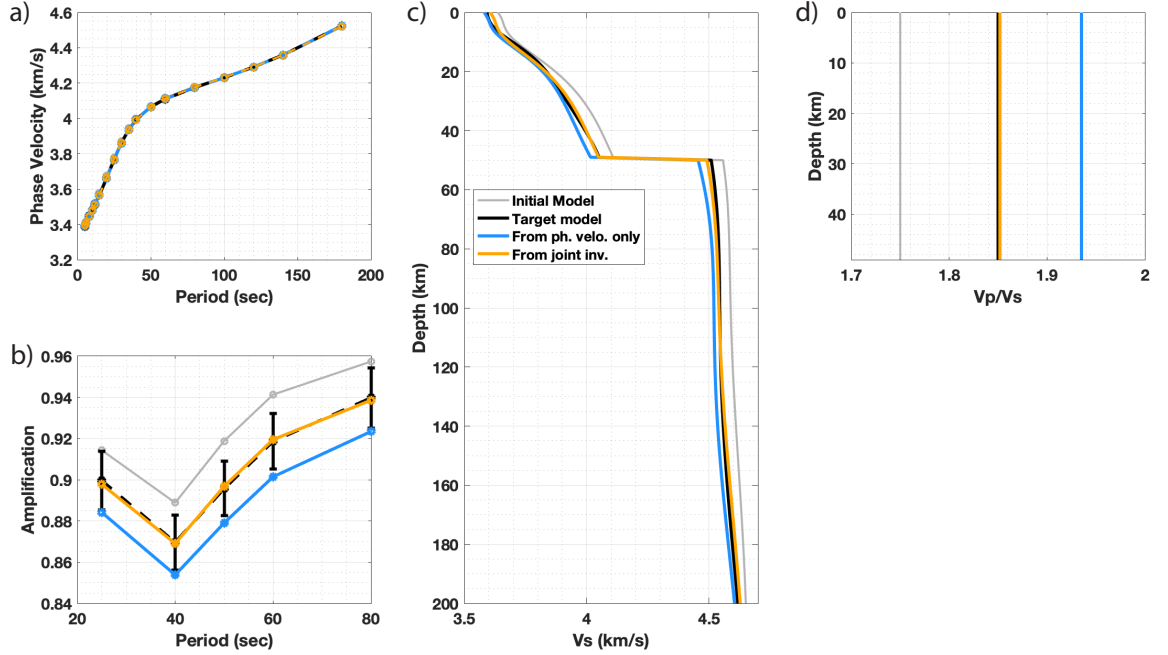
A challenge with inverting the amplification observations is that, unlike phase velocity, they are relative, not absolute, amplification and represent deviations away from the value at a selected station, which we refer to as the normalization station. We approach this challenge in two ways. First, we choose a normalization station where  $V_{sv}$  and  $V_p$  are well constrained by previous studies, especially where crustal  $V_p$  models from refraction/ reflection studies are available. We use the assumed  $V_{sv}$  and  $V_p$  at the normalization station to predict absolute amplification, with which all other relative amplification observations are converted to absolute amplification. Second, we test how our choice of the normalization station and its elastic structure affects the recovered  $V_{sv}$  and  $V_p/V_s$  models. As discussed in Section 4, we conclude from these tests that lateral variations in  $V_p/V_s$  are robustly recovered, but absolute  $V_p/V_s$  depends somewhat on choices about the normalization station.

### 3 Results

In this section, we first present a series of synthetic experiments to illustrate the capability of amplification to resolve crustal  $V_p/V_s$ . We then apply the joint inversion of Rayleigh wave phase velocity and amplification to the western US.

#### 3.1. Synthetic Experiments

In Figure 2 we show how the joint inversion better recovers crustal  $V_p/V_s$  than inverting phase velocity alone. Here, the target model has crustal  $V_p/V_s = 1.85$ , and the inversion solves for updates to  $V_{sv}$  at depths  $\leq 300$  km and a single value of  $V_p/V_s$  in the crust. When only phase velocity data are used, the recovered  $V_{sv}$  is slower than the input model in the lower crust and shallow mantle, and the recovered  $V_p/V_s$  is too high. When phase velocities and amplification are jointly inverted, both parameters are well recovered because amplification has a higher sensitivity to crustal  $V_p$  and a sensitivity to  $V_{sv}$  that complements that of phase velocity, for example, by having zero crossings in the sign of sensitivity (Figure 1a). Additional tests are described in Text S3 in Supporting Information S1. We explore the effects of noise and of allowing for a two-layer crustal  $V_p/V_s$  model (Figure S9 in Supporting Information S1), although we ultimately favor a one-layer  $V_p/V_s$  model. The largest amplification uncertainty results in a standard deviation of just 0.014 for crustal  $V_p/V_s$  in the one-layer crustal  $V_p/V_s$  model. We also show the potential for jointly inverting Rayleigh wave phase velocity and amplification to better constrain shallow sedimentary layers than phase velocity alone (Figure S10 in Supporting Information S1).



**Figure 2.2: Figure 2.** Results of synthetic experiment to investigate recovery of Vsv and crustal Vp/Vs simultaneously. Black = target (input) model, gray = initial model, blue = inverting only phase velocity, yellow = inverting phase velocity and amplification. (a) Rayleigh wave phase velocity. (b) Rayleigh wave amplification. Black bars show uncertainty on the amplification measurements. (c) Depth - dependent Vsv. (d) Crustal Vp/Vs.

### 3.2. Application to Western US

We apply the two-step joint inversion method to the phase velocity and amplification maps described in Section 2 (hereafter “This Study”). In order to explore how the resulting Vsv and Vp/Vs models depend on the amplification data sets, we repeat the joint inversion using two additional amplification data sets from the western US: Eddy and Ekström (2020) (hereafter “EE”) and a hybrid data set that combines the measurements of Lin et al. (2012) at 30 s and Schardong et al. (2019) at 40–78 s (hereafter “LS”) in order to span the necessary period range. The three data sets are compared in Figures S6–S7 in Supporting Information S1. While agreement between them is generally good, there is some scatter, owing both to differences in the underlying

measurements of Rayleigh wave amplitude and challenges in isolating local amplification from the other factors that influence amplitude. We do not favor the results from a particular data set and instead discuss all three results.

In the second step, we iterate until the largest  $V_{sv}$  change at any depth between iterations is  $<0.002$  km/s, and we always perform at least 10 iterations. Station TA.M07A is set as the normalization station; the effect of this choice is discussed in Section 4.1.  $V_{sv}$  structure at this station is from Shen and Ritzwoller (2016) for depths of 0–150 km and from Xie et al. (2018) for depths of 150–300 km. The crustal  $V_p$  structure is sourced from a local refraction/ reflection profile (Lerch et al., 2007). We retain and interpret only models where both phase velocity and amplification are well fit by the Step 2 final model and mask stations with sedimentary thickness  $\geq 2$  km, since amplification is sensitive to uncertainties in assumed sedimentary structure (Figure S10 in Supporting Information S1). Future work will consider whether some stations with large misfits require a multi-layer  $V_p/V_s$  model as well as the potential to solve for the properties of thick sedimentary layers by including higher-frequency amplification data.

Results are shown in Figure 3 for  $V_p/V_s$  and Figure S11 in Supporting Information S1 for  $V_{sv}$ . For the models derived with amplification values from this study and LS, 90% of the stations have crustal  $V_p/V_s$  within the range 1.58–1.8, with a slightly larger range (1.59–1.95) for the model obtained with the EE amplification. We find consistent variations in crustal  $V_p/V_s$  values between the three models (Figures 3a–3c and Figure S12 in Supporting Information S1), in many places corresponding to surface features and agreeing with previous studies (Figure 3d and Figure S1 in Supporting Information S1). All three models identify high  $V_p/V_s$  in the northwestern Basin and Range and High

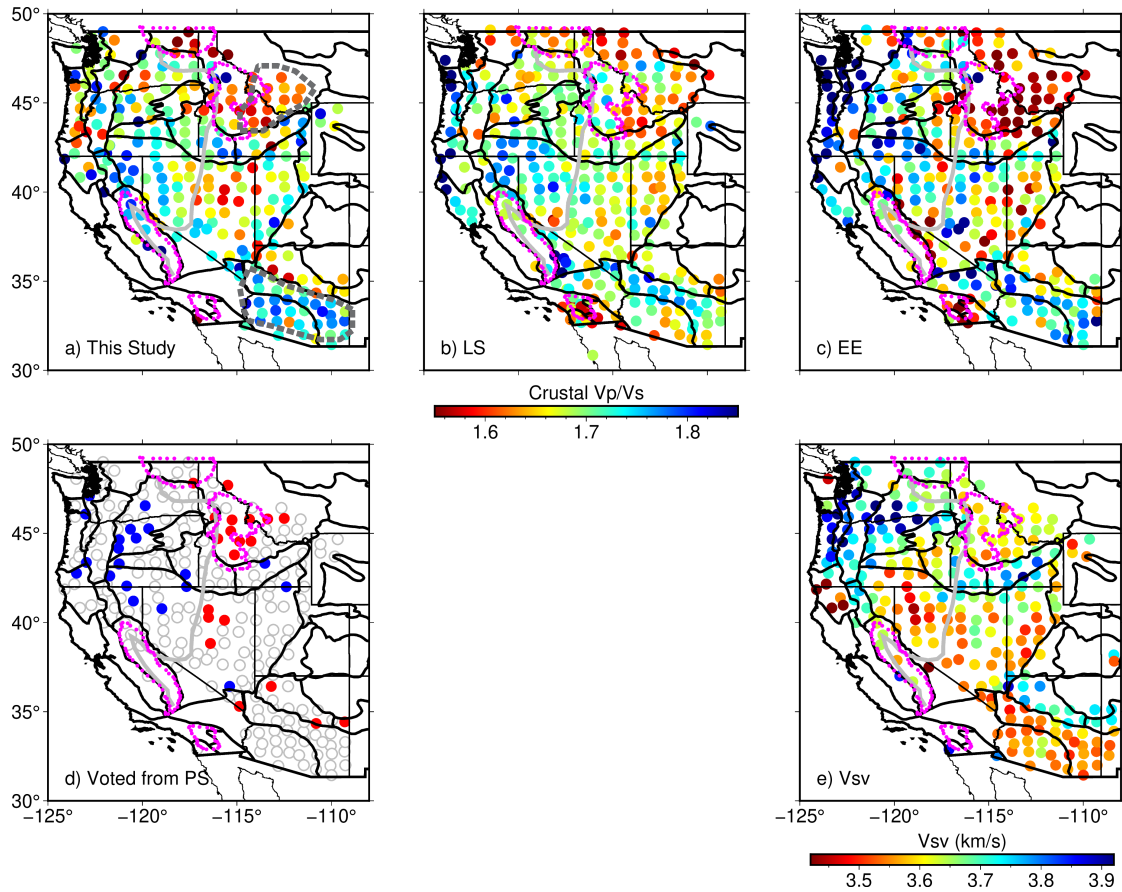
Lava Plain, consistent with previous regional studies that attribute it to mafic crust containing partial melt (Eagar et al., 2011). Like Lowry and Pérez-Gussinyé (2011) and Sui et al. (2023), all three models find lower  $V_p/V_s$  in the extensional Basin and Range and associated with granitic batholiths such as the Coast Range (three models), Idaho (this study and EE), Baja (LS and EE), and Sierra Nevada (EE) batholiths. As identified by Lowry and Pérez-Gussinyé (2011), there is a westward increase in  $V_p/V_s$  across the 0.706 isopleth of initial  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio (ISr) in the Central Basin and Range in all three models, considered to represent the western extent of Precambrian crust.

In a few places, such as westernmost Oregon and Washington, Yellowstone and the Snake River Plain, the three models lack consensus due to disagreements between the amplification measurements. High  $V_p/V_s$  in Yellowstone and the Snake River Plain is observed in the model derived from this study and may reflect a combination of partial melt and basaltic material in the crust (DeNosaquo et al., 2009; Lowry & Pérez-Gussinyé, 2011; Sui et al., 2023). However, it is not a persistent feature in the LS and EE models.

It is not straightforward to compare the new crustal  $V_p/V_s$  results with earlier studies that estimated  $V_p/V_s$  from body waves, due to limited agreement between those earlier studies (Figure S1 in Supporting Information S1). Figure 3d summarizes where the earlier studies agree on high and low  $V_p/V_s$  values. Agreement with the results obtained from joint inversion of phase velocity and amplification can be seen in the Idaho Batholith and Rocky Mountains Basin and Range to the north of the Snake River Plain and in the westward increase of  $V_p/V_s$  across the Central Basin and Range.

The  $V_{sv}$  model agrees well with previous studies (e.g., Shen & Ritzwoller, 2016). Changes to it between Steps 1 and 2 are generally slight but can be as large as 4% in the

crust and shallowest mantle. The largest changes occur where  $V_p/V_s$  from Step 2 deviates from the assumed value in Step 1, with a shift to higher  $V_{sv}$  corresponding to a shift to lower  $V_p/V_s$  and vice versa. Figure S2 in Supporting Information S1 illustrates how incorrect assumptions about crustal  $V_p/V_s$  can introduce error into the recovered  $V_{sv}$  model, providing an explanation for the adjustments to the  $V_{sv}$  model.



**Figure 2.3: Figure 3.** Crustal  $V_p/V_s$  from Step 2 joint inversion of phase velocity and amplification data from (a) this study, (b) a hybrid of Lin et al. (2012) and Schardong et al. (2019), and (c) Eddy and Ekström (2020). Gray dashed circles in panel (a) indicate the Rocky Mountains Basin and Range and Southern Basin and Range. Magenta - dotted lines enclose large batholiths. Solid gray line is the approximate  $^{87}\text{Sr}/^{86}\text{Sr} = 0.706$  isotopic isopleth. Notice (c) is plotted with the same median value as panel (a) for comparison. (d) “Voted” relative crustal  $V_p/V_s$  values from four different models in previous studies (PS) that did not use surface-wave data (Figure S1 in Supporting Information S1). Blue circles indicate that at least three of the previous studies agree on relatively high  $V_p/V_s$  values, and red circles indicate that at least three of the previous studies agree on relatively low values. White circles indicate either lack of agreement or agreement on moderate  $V_p/V_s$  values. (e)  $V_{sv}$  in the middle crust (characterized as the upper half of the depth range between 11 km and the Moho).



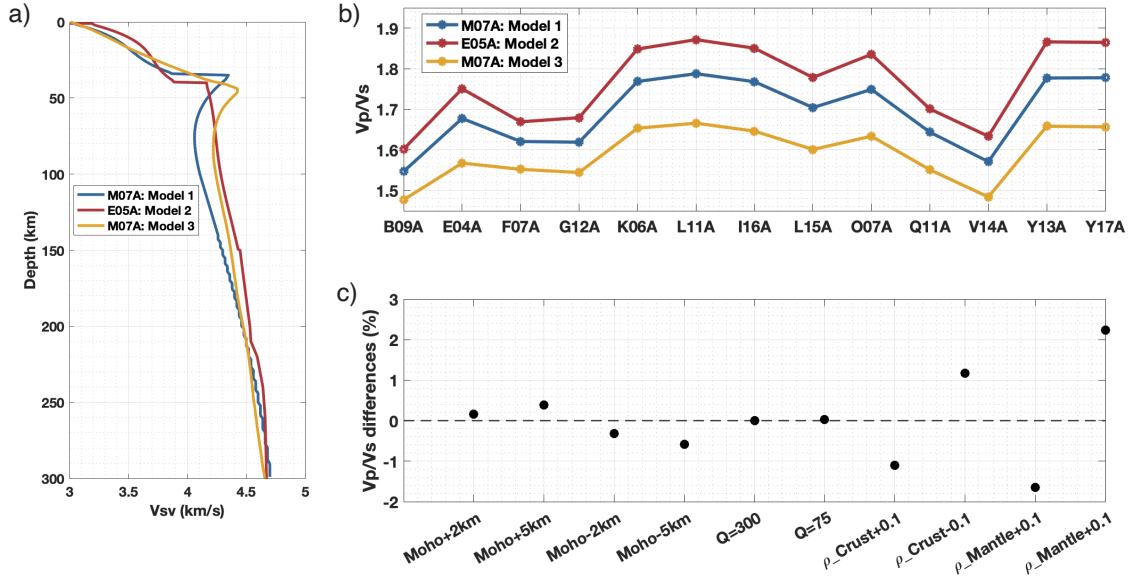
## 4 Discussion

### 4.1. Sources of Uncertainty

One source of uncertainty is choosing a normalization station, where we assume the seismic structure is well known, to convert relative amplification into absolute amplification. In Figures 4a and 4b we test how this choice affects the crustal  $V_p/V_s$  values. We try two EarthScope USArray stations as normalization stations: TA.M07A near an active-source profile that determined crustal  $V_p$  (Lerch et al., 2007) and TA.E05A near the iMUSH experiment (Kiser et al., 2016). We try two different  $V_{sv}$  models for TA.M07A and one for TA.E05A (Figure 4a; Shen & Ritzwoller, 2016; Xie et al., 2018; Eilon et al., 2018; Schaeffer & Lebedev, 2014). We repeat the joint inversion at 13 stations with these three combinations of normalization station and seismic model. Figure 4b shows that while absolute  $V_p/V_s$  depends somewhat on choices about the normalization station and structure, variations between the stations are robustly recovered.  $V_{sv}$  is even less sensitive to these choices (Figure S13 in Supporting Information S1). Thus, we can be confident in the relative crustal  $V_p/V_s$  values, somewhat confident in the absolute crustal  $V_p/V_s$  values, and confident in the absolute  $V_{sv}$  values.

In Figure 4c we investigate other potential sources of error in the recovered crustal  $V_p/V_s$  using real data from station TA.L15A. An incorrect assumption of Moho depth by 5 km will cause a  $V_p/V_s$  error less than 1%. Changing mantle shear attenuation ( $1/Q$ ) from 1/143 to 1/75 or 1/300 changes  $V_p/V_s$  by less than 0.05%. Changing density by 0.1 g/cm<sup>3</sup> in the entire crust or the mantle shallower than 300 km can bias crustal  $V_p/V_s$  by about 1% and 2%, respectively. Therefore, while model biases from these parameters are

not entirely negligible, they are generally much smaller than the variations in crustal  $V_p/V_s$  observed in our result, which exceed  $\pm 8\%$ .



**Figure 2.4: Figure 4.** (a, b) How crustal  $V_p/V_s$  depends on the normalization station and velocity model at 13 example stations and for: normalization station TA.M07A with two different elastic models and normalization station TA.E05A with a third model. (c) How  $V_p/V_s$  is affected by assumptions in the inversion: deepening or shallowing of Moho by 2 or 5 km (Moho+2 km; Moho+5 km; Moho- 2 km; Moho- 5 km), changing mantle  $Q$  from 143 to  $Q = 300$  or  $Q = 75$ , increasing ( $\rho_{Crust}+0.1$ ) or decreasing ( $\rho_{Crust}-0.1$ ) crustal density by  $0.1 \text{ g/cm}^3$ , increasing ( $\rho_{Mantle}+0.1$ ) or decreasing ( $\rho_{Mantle}-0.1$ ) density by  $0.1 \text{ g/cm}^3$  in the mantle shallower than 300 km.

## 4.2. Implications for Crustal Composition and Evolution

To first order, crustal  $V_p/V_s$  is considered to be mostly insensitive to temperature and somewhat sensitive to  $\text{SiO}_2$  content (Christensen, 1996; Lowry & Pérez-Gussinyé, 2011; Sui et al., 2023), with some studies using crustal  $V_p/V_s$  as a proxy for quartz abundance (Savard et al., 2018; Sui et al., 2023). Christensen (1996) showed that an increase in  $\text{SiO}_2$  from 55 to 75 wt% leads to a nearly linear decrease in  $V_p/V_s$ . It is clear that an unambiguous mapping between  $V_p/V_s$  and  $\text{SiO}_2$  content does not exist, but very low

Vp/Vs is best explained by high SiO<sub>2</sub> (e.g., Sui et al., 2023), and most explanations for low crustal Vp/Vs invoke silica enrichment. While mafic rocks are generally characterized by average-to-high Vp/Vs values, their variation in Vp/Vs is not correlated with SiO<sub>2</sub> (Christensen, 1996; Hacker et al., 2015).

In this section, we highlight several results that may bear on the composition of the crust. One, we observe an eastward decrease in crustal Vp/Vs across central Basin and Range in all three models, particularly across the ISr = 0.706 isopleth. Much of the North American Cordillera formed through the accretion of allochthonous terranes onto the west margin of the North American cratonic basement during the Mesozoic and early Cenozoic time (Coney et al., 1980), with the ISr = 0.706 isopleth regarded as the approximate western boundary of the North American cratonic basement (e.g., Armstrong et al., 1977; Kistler & Peterman, 1973; Whitmeyer & Karlstrom, 2007). The eastward decrease in Vp/Vs in this study supports the interpretation of young mafic oceanic terrane in the west and the Precambrian crust in the east.

One of the most striking agreements among crustal Vp/Vs models derived from the three amplification data sets is very low Vp/Vs in the Rocky Mountains Basin and Range in eastern Idaho and western Montana, which suggests high SiO<sub>2</sub> in the crust. This same area was investigated by Berry et al. (2022), who sought explanations for its anomalously high elevation and low Moho temperature. Berry et al. (2022) used thermodynamic calculations to show that hydration of the crust could reduce the Moho temperature through endothermic reactions that promote the growth of quartz at the expense of plagioclase, orthopyroxene, and garnet (Ma & Lowry, 2017). The prevalence of quartz, with its low density (Christensen, 1996; Lowry & Pérez-Gussinyé, 2011), supports the

high elevation and also explains the low crustal  $V_p/V_s$ . Hydration of the crust in the western US by fluids expelled from the Laramide flat slab (Humphreys et al., 2003) is consistent with xenolith compositions (Jones et al., 2015) and evidence for low-density lower crust from isostatic calculations (Schmandt et al., 2015). We speculate that this area is underlain by thicker and colder lithosphere, as suggested by high  $V_{sv}$  in the shallow mantle (Figure S11 in Supporting Information S1), which favors hydration reactions (Porter et al., 2017) that result in quartz enrichment and low crustal  $V_p/V_s$ .

High  $V_p/V_s$  values are found in the Southern Basin and Range, another feature consistently revealed by all three models. Buehler and Shearer (2014) also observed very high  $V_p/V_s$  in this region from  $P_n$  and  $S_n$  data, whereas moderate crustal  $V_p/V_s$  estimates were obtained from H-k stacking of receiver functions (Ma & Lowry, 2017; Sui et al., 2023), possibly due to the strong azimuthal dependence of  $V_p/V_s$  in this area (Frassetto et al., 2006). Sammon et al. (2020) found that this area is characterized by more mafic crust than that of the Colorado Plateau and Northern Basin and Range. They proposed that the Southern Basin and Range crust has been thinned by removal of felsic material from the top, rather than by delamination at the bottom, allowing ductile uplift of the lower crust. Our finding of high  $V_p/V_s$  in this region aligns with this interpretation, substantiating the predominantly mafic component in the crust, and providing seismological support for the top-down thinning model rather than delamination.

## 5 Conclusions

We develop a new two-step approach to jointly invert Rayleigh wave phase velocity and local amplification to constrain  $V_{sv}$  in the crust and uppermost mantle and  $V_p/V_s$  in the crust. We demonstrate its efficacy with synthetic experiments and apply it to real data in the western US, using three amplification data sets to explore the most robust features. Tests indicate that although the absolute crustal  $V_p/V_s$  values depend somewhat on the choice of normalization station and its assumed seismic model, variations are consistently recovered. We find  $V_p/V_s$  variations that correspond to tectonic structures and agree across models derived from the three data sets, such as high  $V_p/V_s$  in the northern Great Basin and High Lava Plain. An eastward decrease in  $V_p/V_s$  across the central Basin and Range highlights differences in crustal composition and origin near western edge of Precambrian crust. We infer quartz enrichment from anomalously low  $V_p/V_s$  in the Rocky Mountains Basin and Range, driven by fluids released from the Laramide flat slab, and associate it with high elevation. High  $V_p/V_s$  in the Southern Basin and Range support a top-down crustal thinning model. In summary, Rayleigh wave amplification offers great promise for estimating crustal  $V_p/V_s$ , which will be especially valuable in locations where traditional approaches, such as receiver functions and local earthquake tomography, are not viable.

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## 7 References

- Abers, G. A., & Hacker, B. R. (2016). A MATLAB toolbox and Excel workbook for calculating the densities, seismic wave speeds, and major element composition of minerals and rocks at pressure and temperature. *Geochemistry, Geophysics, Geosystems*, 17(2), 616–624. <https://doi.org/10.1002/2015GC006171>
- Armstrong, R. L., Taubeneck, W. H., & Hales, P. O. (1977). Rb-Sr and K-Ar geochronometry of Mesozoic granitic rocks and their Sr isotopic composition, Oregon, Washington, and Idaho. *Geological Society of America Bulletin*, 88(3), 397–411. [https://doi.org/10.1130/0016-7606\(1977\)88<397:rakgom>2.0.co;2](https://doi.org/10.1130/0016-7606(1977)88<397:rakgom>2.0.co;2)
- Audet, P., & Bürgmann, R. (2014). Possible control of subduction zone slow-earthquake periodicity by silica enrichment. *Nature*, 510(7505), 389–392. <https://doi.org/10.1038/nature13391>

- Berry, M. A., Lowry, A. R., Ma, X., Kanda, R. V., & Schutt, D. L. (2022). Wet roots of high elevation in the western United States. *Earth and Planetary Science Letters*, 584, 117483. <https://doi.org/10.1016/j.epsl.2022.117483>
- Bowden, D. C., Tsai, V. C., & Lin, F. C. (2015). Site amplification, attenuation, and scattering from noise correlation amplitudes across a dense array in Long Beach, CA. *Geophysical Research Letters*, 42(5), 1360–1367. <https://doi.org/10.1002/2014GL062662>
- Bowden, D. C., Tsai, V. C., & Lin, F. C. (2017). Amplification and attenuation across USArray using ambient noise wavefront tracking. *Journal of Geophysical Research: Solid Earth*, 122(12), 10–086. <https://doi.org/10.1002/2017JB014804>
- Brocher, T. M. (2005). Empirical relations between elastic wavespeeds and density in the Earth's crust. *Bulletin of the Seismological Society of America*, 95(6), 2081–2092. <https://doi.org/10.1785/0120050077>
- Brunsvik, B., & Eilon, Z. (2023). Radial Anisotropy in Receiver Function H- $\kappa$  Stacks. *Seismological Research Letters*, 95(1), 479–487. <https://doi.org/10.1785/0220230114>
- Buehler, J. S., & Shearer, P. M. (2014). Anisotropy and Vp/Vs in the uppermost mantle beneath the western United States from joint analysis of Pn and Sn phases. *Journal of Geophysical Research: Solid Earth*, 119(2), 1200–1219. <https://doi.org/10.1002/2013JB010559>

- Christensen, N. I. (1996). Poisson's ratio and crustal seismology. *Journal of Geophysical Research*, 101(B2), 3139–3156. <https://doi.org/10.1029/95JB03446>
- Coney, P. J., Jones, D. L., & Monger, J. W. (1980). Cordilleran suspect terranes. *Nature*, 288(5789), 329–333. <https://doi.org/10.1038/288329a0>
- DeNosaquo, K. R., Smith, R. B., & Lowry, A. R. (2009). Density and lithospheric strength models of the Yellowstone–Snake River Plain volcanic system from gravity and heat flow data. *Journal of Volcanology and Geothermal Research*, 188(1–3), 108–127. <https://doi.org/10.1016/j.jvolgeores.2009.08.006>
- Diaferia, G., Cammarano, F., Piana Agostinetti, N., Gao, C., Lekic, V., Molinari, I., & Boschi, L. (2019). Inferring crustal temperatures beneath Italy from joint inversion of receiver functions and surface waves. *Journal of Geophysical Research: Solid Earth*, 124(8), 8771–8785. <https://doi.org/10.1029/2019JB018340>
- Eagar, K. C., Fouch, M. J., James, D. E., & Carlson, R. W. (2011). Crustal structure beneath the High Lava Plains of eastern Oregon and surrounding regions from receiver function analysis. *Journal of Geophysical Research*, 116(B2), B02313. <https://doi.org/10.1029/2010JB007795>
- Eddy, C. L., & Ekström, G. (2014). Local amplification of Rayleigh waves in the continental United States observed on the USArray. *Earth and Planetary Science Letters*, 402, 50–57. <https://doi.org/10.1016/j.epsl.2014.01.013>



- Eddy, C. L., & Ekström, G. (2020). Comparisons between measurements and predictions of Rayleigh wave amplification across the contiguous United States. *Physics of the Earth and Planetary Interiors*, 299, 106407.  
<https://doi.org/10.1016/j.pepi.2019.106407>
- Eilon, Z., Fischer, K. M., & Dalton, C. A. (2018). An adaptive Bayesian inversion for upper-mantle structure using surface waves and scattered body waves. *Geophysical Journal International*, 214(1), 232–253.  
<https://doi.org/10.1093/gji/ggy137>
- Ekström, G. (2017). Short-period surface-wave phase velocities across the conterminous United States. *Physics of the Earth and Planetary Interiors*, 270, 168–175.  
<https://doi.org/10.1016/j.pepi.2017.07.010>
- Fenneman, N. M. (1928). Physiographic divisions of the United States. *Annals of the Association of American Geographers*, 18(4), 261–353.  
<https://doi.org/10.1080/00045602809357034>
- Frassetto, A., Gilbert, H., Zandt, G., Beck, S., & Fouch, M. J. (2006). Support of high elevation in the southern Basin and Range based on the composition and architecture of the crust in the Basin and Range and Colorado Plateau. *Earth and Planetary Science Letters*, 249(1–2), 62–73.  
<https://doi.org/10.1016/j.epsl.2006.06.040>

- Hacker, B. R., Kelemen, P. B., & Behn, M. D. (2015). Continental lower crust. *Annual Review of Earth and Planetary Sciences*, 43(1), 167–205.  
<https://doi.org/10.1146/annurev-earth-050212-124117>
- Haines, A. J. (1980). Differences in time terms between New Zealand seismograph stations and implications for crustal structure. *New Zealand Journal of Geology and Geophysics*, 23(5–6), 541–549.  
<https://doi.org/10.1080/00288306.1980.10424126>
- Hammond, J. O. S. (2014). Constraining melt geometries beneath the Afar Depression, Ethiopia from teleseismic receiver functions: The anisotropic H- $\kappa$  stacking technique. *Geochemistry, Geophysics, Geosystems*, 15(4), 1316–1332.  
<https://doi.org/10.1002/2013gc005186>
- Hariharan, A., Dalton, C. A., Babikoff, J., & Ekström, G. (2022). Controls on surface wave overtone interference. *Geophysical Journal International*, 228(3), 1665–1683. <https://doi.org/10.1093/gji/ggab424>
- Huang, Y., Dalton, C. A., & Hariharan, A. (2025). Data for A new approach to constrain crustal Vp/Vs from Rayleigh wave phase velocity and local amplification: Application to the western US [Dataset]. Brown University Open Data Collection. Brown Digital Repository. Brown University Library.  
<https://doi.org/10.26300/1yec-1e81>

- Humphreys, E., Hessler, E., Dueker, K., Farmer, G. L., Erslev, E., & Atwater, T. (2003). How Laramide-age hydration of North American lithosphere by the Farallon slab controlled subsequent activity in the western United States. *International Geology Review*, 45(7), 575–595. <https://doi.org/10.2747/0020-6814.45.7.575>
- IRIS Transportable Array. (2003). USArray transportable array [Dataset]. International Federation of Digital Seismograph Networks. <https://doi.org/10.7914/SN/TA>
- Jagoutz, O., & Behn, M. D. (2013). Foundering of lower island-arc crust as an explanation for the origin of the continental Moho. *Nature*, 504(7478), 131–134. <https://doi.org/10.1038/nature12758>
- Jin, G., & Gaherty, J. B. (2015). Surface wave phase-velocity tomography based on multichannel cross-correlation. *Geophysical Journal International*, 201(3), 1383–1398. <https://doi.org/10.1093/gji/ggv079>
- Jones, C. H., Mahan, K. H., Butcher, L. A., Levandowski, W. B., & Farmer, G. L. (2015). Continental uplift through crustal hydration. *Geology*, 43(4), 355–358. <https://doi.org/10.1130/g36509.1>
- Kennett, B. L., Engdahl, E. R., & Buland, R. (1995). Constraints on seismic velocities in the Earth from traveltimes. *Geophysical Journal International*, 122(1), 108–124. <https://doi.org/10.1111/j.1365-246X.1995.tb03540.x>

- Kiser, E., Palomeras, I., Levander, A., Zelt, C., Harder, S., Schmandt, B., et al. (2016). Magma reservoirs from the upper crust to the Moho inferred from high-resolution Vp and Vs models beneath Mount St. Helens, Washington State, USA. *Geology*, 44(6), 411–414. <https://doi.org/10.1130/G37591.1>
- Kistler, R. W., & Peterman, Z. E. (1973). Variations in Sr, Rb, K, Na, and initial Sr87/Sr86 in Mesozoic granitic rocks and intruded wall rocks in central California. *Geological Society of America Bulletin*, 84(11), 3489–3512. [https://doi.org/10.1130/0016-7606\(1973\)84<3489:visrkn>2.0.co;2](https://doi.org/10.1130/0016-7606(1973)84<3489:visrkn>2.0.co;2)
- Kuo-Chen, H., Wu, F. T., Jenkins, D. M., Mechie, J., Roecker, S. W., Wang, C. Y., & Huang, B. S. (2012). Seismic evidence for the  $\alpha$ - $\beta$  quartz transition beneath Taiwan from Vp/Vs tomography. *Geophysical Research Letters*, 39(22), L22302. <https://doi.org/10.1029/2012GL053649>
- Lerch, D. W., Klemperer, S. L., Glen, J. M. G., Ponce, D. A., Miller, E. L., & Colgan, J. P. (2007). Crustal structure of the northwestern Basin and Range Province and its transition to unextended volcanic plateaus. *Geochemistry, Geophysics, Geosystems*, 8(2), Q02011. <https://doi.org/10.1029/2006GC001429>
- Levandowski, W., Jones, C. H., Shen, W., Ritzwoller, M. H., & Schulte-Pelkum, V. (2014). Origins of topography in the western US: Mapping crustal and upper mantle density variations using a uniform seismic velocity model. *Journal of Geophysical Research: Solid Earth*, 119(3), 2375–2396. <https://doi.org/10.1002/2013JB010607>

- Li, J., Song, X., Zhu, L., & Deng, Y. (2017). Joint inversion of surface wave dispersions and receiver functions with P velocity constraints: Application to Southeastern Tibet. *Journal of Geophysical Research: Solid Earth*, 122(9), 7291–7310. <https://doi.org/10.1002/2017JB014135>
- Lin, F. C., & Ritzwoller, M. H. (2011). Helmholtz surface wave tomography for isotropic and azimuthally anisotropic structure. *Geophysical Journal International*, 186(3), 1104–1120. <https://doi.org/10.1111/j.1365-246X.2011.05070.x>
- Lin, F. C., Tsai, V. C., & Ritzwoller, M. H. (2012). The local amplification of surface waves: A new observable to constrain elastic velocities, density, and anelastic attenuation. *Journal of Geophysical Research*, 117(B6), B06302. <https://doi.org/10.1029/2012JB009208>
- Lowry, A. R., & Pérez-Gussinyé, M. (2011). The role of crustal quartz in controlling Cordilleran deformation. *Nature*, 471(7338), 353–357. <https://doi.org/10.1038/nature09912>
- Ma, X., & Lowry, A. R. (2017). USArray imaging of continental crust in the conterminous United States. *Tectonics*, 36(12), 2882–2902. <https://doi.org/10.1002/2017TC004540>
- Mooney, W. D., Barrera-Lopez, C., Suárez, M. G., & Castelblanco, M. A. (2023). Earth crustal model 1 (ECM1): A 1 x 1 global seismic and density model. *Earth-Science Reviews*, 243, 104493. <https://doi.org/10.1016/j.earscirev.2023.104493>

- Piana Agostinetti, N., & Amato, A. (2009). Moho depth and Vp/Vs ratio in peninsular Italy from teleseismic receiver functions. *Journal of Geophysical Research*, 114(B6), B06303. <https://doi.org/10.1029/2008JB005899>
- Porter, R., Hoisch, T., & Holt, W. E. (2017). The role of lower-crustal hydration in the tectonic evolution of the Colorado Plateau. *Tectonophysics*, 712, 221–231. <https://doi.org/10.1016/j.tecto.2017.05.025>
- Rychert, C. A., & Harmon, N. (2016). Stacked P-to-S and S-to-P receiver functions determination of crustal thickness, Vp, and Vs: The H-V stacking method. *Geophysical Research Letters*, 43(4), 1487–1494. <https://doi.org/10.1002/2015GL067010>
- Sammon, L. G., Gao, C., & McDonough, W. F. (2020). Lower crustal composition in the southwestern United States. *Journal of Geophysical Research: Solid Earth*, 125(3), e2019JB019011. <https://doi.org/10.1002/essoar.10501208.1>
- Savard, G., Bostock, M. G., & Christensen, N. I. (2018). Seismicity, metamorphism, and fluid evolution across the northern Cascadia fore arc. *Geochemistry, Geophysics, Geosystems*, 19(6), 1881–1897. <https://doi.org/10.1029/2017GC007417>
- Schaeffer, A. J., & Lebedev, S. (2014). Imaging the North American continent using waveform inversion of global and USArray data. *Earth and Planetary Science Letters*, 402, 26–41. <https://doi.org/10.1016/j.epsl.2014.05.014>

- Schardong, L., Ferreira, A. M., Berbellini, A., & Sturgeon, W. (2019). The anatomy of uppermost mantle shear-wave speed anomalies in the western US from surface-wave amplification. *Earth and Planetary Science Letters*, 528, 115822. <https://doi.org/10.1016/j.epsl.2019.115822>
- Schmandt, B., Lin, F. C., & Karlstrom, K. E. (2015). Distinct crustal isostasy trends east and west of the Rocky Mountain Front. *Geophysical Research Letters*, 42(23), 10–290. <https://doi.org/10.1002/2015GL066593>
- Shen, W., & Ritzwoller, M. H. (2016). Crustal and uppermost mantle structure beneath the United States. *Journal of Geophysical Research: Solid Earth*, 121(6), 4306–4342. <https://doi.org/10.1002/2016JB012887>
- Shillington, D. J., Van Avendonk, H. J., Behn, M. D., Kelemen, P. B., & Jagoutz, O. (2013). Constraints on the composition of the Aleutian arc lower crust from VP/VS. *Geophysical Research Letters*, 40(11), 2579–2584. <https://doi.org/10.1002/grl.50375>
- Steck, L. K., Begnaud, M. L., Phillips, S., & Stead, R. (2011). Tomography of crustal P and S travel times across the western United States. *Journal of Geophysical Research*, 116(B11), B11304. <https://doi.org/10.1029/2011JB008260>
- Sturgeon, W., Ferreira, A. M., Schardong, L., & Marignier, A. (2023). Crustal structure of the Western US from Rayleigh and Love wave amplification data. *Journal of Geophysical Research: Solid Earth*, 128(8), e2022JB026148. <https://doi.org/10.1029/2022JB026148>

- Sui, S., Shen, W., Mahan, K., & Schulte-Pelkum, V. (2023). Constraining the crustal composition of the continental US using seismic observables. *Bulletin*, 135(7–8), 2038–2056. <https://doi.org/10.1130/b36229.1>
- Sun, Y., Niu, F., Liu, H., Chen, Y., & Liu, J. (2012). Crustal structure and deformation of the SE Tibetan plateau revealed by receiver function data. *Earth and Planetary Science Letters*, 349, 186–197. <https://doi.org/10.1016/j.epsl.2012.07.007>
- Tugume, F., Nyblade, A., & Julià, J. (2012). Moho depths and Poisson's ratios of Precambrian crust in East Africa: Evidence for similarities in Archean and Proterozoic crustal structure. *Earth and Planetary Science Letters*, 355, 73–81. <https://doi.org/10.1016/j.epsl.2012.08.041>
- Watanabe, T. (1993). Effects of water and melt on seismic velocities and their application to characterization of seismic reflectors. *Geophysical Research Letters*, 20(24), 2933–2936. <https://doi.org/10.1029/93GL03170>
- Wessel, P., Luis, J., Uieda, L., Scharroo, R., Wobbe, F., Smith, W. H., & Tian, D. (2019). The generic mapping tools version 6. *Geochemistry, Geophysics, Geosystems*, 20(11), 5556–5564. <https://doi.org/10.1029/2019gc008515>
- Whitmeyer, S. J., & Karlstrom, K. E. (2007). Tectonic model for the Proterozoic growth of North America. *Geosphere*, 3(4), 220–259. <https://doi.org/10.1130/ges00055.1>
- Xie, J., Chu, R., & Yang, Y. (2018). 3-D upper-mantle shear velocity model beneath the contiguous United States based on broadband surface wave from ambient seismic



noise. *Pure and Applied Geophysics*, 175(10), 3403–3418.

<https://doi.org/10.1007/s00024-018-1881-2>

Yeck, W. L., Sheehan, A. F., & Schulte-Pelkum, V. (2013). Sequential H- $\kappa$  stacking to obtain accurate crustal thicknesses beneath sedimentary basins. *Bulletin of the Seismological Society of America*, 103(3), 2142–2150.

<https://doi.org/10.1785/0120120290>

Zandt, G., & Ammon, C. J. (1995). Continental crust composition constrained by measurements of crustal Poisson's ratio. *Nature*, 374(6518), 152–154.

<https://doi.org/10.1038/374152a0>

Zhu, L., & Kanamori, H. (2000). Moho depth variation in southern California from teleseismic receiver functions. *Journal of Geophysical Research*, 105(B2), 2969–2980. <https://doi.org/10.1029/1999JB900322>

### **References From the Supporting Information:**

Begnaud, M. L., Anderson, D. N., Phillips, W. S., Myers, S. C., & Ballard, S. (2016). Developing and validating path-dependent uncertainty estimates for use with the regional seismic travel time (RSTT) model. In *AGU fall meeting abstracts* (Vol. 2016, pp. S51A–S2745).

Dahlen, F., & Tromp, J. (1998). Theoretical global seismology. In *Theoretical global seismology*. Princeton university press. <https://doi.org/10.1515/9780691216157>

Dziewonski, A. M., & Anderson, D. L. (1981). Preliminary reference Earth model.

Physics of the Earth and Planetary Interiors, 25(4), 297–356.

[https://doi.org/10.1016/0031-9201\(81\)90046-7](https://doi.org/10.1016/0031-9201(81)90046-7)

Ferreira, A. M., & Woodhouse, J. H. (2007). Source, path and receiver effects on seismic surface waves. Geophysical Journal International, 168(1), 109–132.

<https://doi.org/10.1111/j.1365-246X.2006.03092.x>

Golos, E. M., Fang, H., & van der Hilst, R. D. (2020). Variations in seismic wave speed and VP/VS ratio in the North American lithosphere. Journal of Geophysical Research: Solid Earth, 125(12), e2020JB020574.

<https://doi.org/10.1029/2020JB020574>

Masters, G., Woodhouse, J. H., & Freeman, G. (2011). Mineos v1.0.2. Retrieved from

<https://geodynamics.org/cig/software/mineos/>

## 8. Supplemental Material

### 8.1 Introduction

This supporting information includes three texts, one supplementary table, 13 supplementary figures, and a description of the data set format generated by this study. Text S1 provides detailed information on the amplification measurements and their uncertainty. Text S2 describes details of the forward and inverse problems in the two-step inversion strategy. Text S3 demonstrates additional synthetic tests of the joint inversion incorporating realistic noise levels, allowing for a two-layer crustal Vp/Vs model configuration, and constraining sedimentary layers. Table S1 shows the evaluation of amplification measurements as a function of period. Figure S1 compares crustal Vp/Vs models across different studies. Figure S2 demonstrates how the assumed crustal Vp/Vs affects Vsv recovery in inversion. Figures S3 and S4 display the Rayleigh wave phase velocity and amplification maps used in this study. Figure S5 illustrates the estimated amplification uncertainty. Figures S6-8 compare the amplification measurements from this study with previous studies and phase velocity. Figure S9 shows the effects of noise and of allowing for a two-layer crustal Vp/Vs model in synthetic tests. Figure S10 presents the potential of the joint inversion to recover sedimentary layers, given the strong sensitivity of amplification to shallow structure. Figure S11 shows Vsv results at several depths from the Step 1 and Step 2 inversions. Figure S12 compares the crustal Vp/Vs results obtained using different amplification measurements. Figure S13 presents the impact of the choice of normalization station and velocity models on Vsv recovery.

A zipped file accompanying this document contains the Rayleigh wave phase velocity and amplification data set generated in this study, the crustal Vp/Vs model, and Vsv model for the crust and uppermost mantle for western U.S. developed in this study.

## 8.2 Text S1.

### Amplification measurements

Following Eddy and Ekström (2014, 2020), we measure the logarithmic amplitude ratio of two nearby stations within 2 degrees:

$$d_{ij}^k = \ln\left(\frac{A_i^k}{A_j^k}\right) = \ln(A_i^k) - \ln(A_j^k), \quad (1)$$

where  $A_i^k$  and  $A_j^k$  are the Rayleigh wave amplitude measured by station  $i$  and  $j$  for earthquake  $k$  at a given frequency, after removing source radiation pattern. For each pair of stations with  $N_e$  available events, the average amplitude ratio is calculated using azimuthally weighting to account for non-uniform back-azimuthal coverage, as follows:

$$\bar{d}_{ij} = \frac{\sum_{k=1}^{N_e} d_{ij}^k w(k)}{\sum_{k=1}^{N_e} w(k)}, \quad (2)$$

where  $w(k)$  is a weighting factor for every back-azimuthal bin of  $15^\circ$ , determined by the number of events  $n_e$  in each bin according to  $w(k) = \frac{1}{n_e}$ , which downweights azimuthal bins with more events. The uncertainty of the station pair's logarithmic amplitude ratio is defined as  $\bar{\sigma}_{ij} = \sigma_{ij}/\sqrt{N_e}$ , where  $\sigma_{ij}$  is the standard deviation of the  $d_{ij}^k$  values. The event-specific log ratios  $d_{ij}^k$  with absolute values greater than 2.3 are identified as outliers, and if more than 10% of the ratios associated with an event are bad, all ratios for that event are removed. We apply additional quality control by retaining only those  $\bar{d}_{ij}$  values that fall within the 10% - 90% quantile range of all  $\bar{d}_{ij}$ . After these

quality-control steps, there are 8819 station-pair  $\bar{d}_{ij}$  values, each determined from an average of 228 events. The local station amplification factor  $A_{R,i}$  is then obtained using a least-squares inversion. The root-mean-square (RMS) of the residual between the observed  $\bar{d}_{ij}$  values and the values predicted by the result of the least-squares inversion is reported in Table S1.

We estimate the uncertainty of amplification factors  $A_{R,i}$ ,  $\xi_{A,i}$ , by propagating the uncertainty of the station-pair logarithmic amplitude ratio uncertainty  $\bar{\sigma}_{ij}$ :

$$\bar{\sigma}_{ij} = \sqrt{\left(\frac{\xi_{A,i}}{A_{R,i}}\right)^2 + \left(\frac{\xi_{A,j}}{A_{R,j}}\right)^2}. \quad (3)$$

Fig. S5 shows amplification uncertainty  $\xi_A$  for every station, which is estimated from Equation 3 using a least-squares inversion. More than 90% of stations have an uncertainty smaller than 0.04 across periods. Table S1 shows the average amplification uncertainties across the data set.

### Comparison to Previous Amplification Measurements

Eddy and Ekström (2020) and Schardong et al. (2019) use the same amplitude-ratio method as we use to measure local station amplification in the same study region. Lin et al. (2012) use a different method to measure Rayleigh wave amplification, based on wavefront tracking and Helmholtz tomography. Their amplification,  $\beta$ , cannot be directly compared to ours because it is based on surface wave potential rather than surface wave amplitudes. To facilitate comparison, we convert their  $\beta$  values into amplification measurements similar to ours using the conversion factor from Bowden et al. (2017). In order to have a data set that spans the necessary period range, we combine the converted Lin et al. (2012) data at 30 s with amplification values from Schardong et al. (2019) data

at 40-78 s into a hybrid data set we label LS. We present a comparison between the three data sets (this study, LS, and EE (Eddy and Ekström, 2020) ) in Figs. S6 and S7. Overall, the comparisons show good agreement between the different data sets. The results indicate strong agreement among data sets, with most correlation coefficients between each two data sets exceeding 0.7, except for the first period, where differences between data sets are larger. The comparison demonstrates consistency across studies and indicates the robustness of our measurements. It is important to note that no two amplification data sets agree perfectly with each other; each amplification estimate carries some uncertainty owing to the underlying amplitude measurements and challenges in isolating the station terms from the other contributions to amplitude. These differences ultimately map into slight differences in the crustal  $V_p/V_s$  models obtained from each amplification data set.

We further compare the three amplification data sets with measured phase velocity from this study (Fig. S8). Because both phase velocity and amplification have strong sensitivity to  $V_{sv}$  but with opposite sign (at most depths), a negative correlation is expected and generally observed. Perfect negative correlations are not expected, since there are differences in the sensitivities of amplification and phase velocity to elastic structure, as shown for example in Figure 1.

### 8.3 Text S2.

#### The forward problem

We assume that relative perturbations in frequency-dependent Rayleigh wave phase velocity  $c(\omega)$  away from a reference phase velocity  $c_0(\omega)$ ,  $\frac{\delta c}{c_0}(\omega)$ , depend on relative perturbations in isotropic P-wave velocity,  $Vp(z)$ , the speed of horizontally traveling, vertically polarized shear waves,  $Vsv(z)$ , and density,  $\rho(z)$  as a function of depth  $z$ :

$$\frac{\delta c(\omega)}{c_0(\omega)} = \int [K_{Vsv}^c(\omega, z) \frac{\delta Vsv(z)}{Vsv_0(z)} + K_{Vp}^c(\omega, z) \frac{\delta Vp(z)}{Vp_0(z)} + K_{\rho}^c(\omega, z) \frac{\delta \rho(z)}{\rho_0(z)}] dz. \quad (4)$$

Here  $K_{Vsv}^c$ ,  $K_{Vp}^c$ , and  $K_{\rho}^c$  are the Fréchet sensitivity kernels that describe how phase-velocity perturbations are related to perturbations in  $Vsv(z)$ ,  $Vp(z)$ , and  $\rho(z)$ . In practice, we use Mineos (Masters et al., 2011) to calculate the eigenfunctions and eigenfrequencies for a 1-D radially anisotropic reference Earth model and the equations for the Fréchet derivatives (e.g., Dahlen & Tromp, 1998) to construct the sensitivity kernels. Since Rayleigh wave  $\frac{\delta c}{c_0}(\omega)$  is sensitive to the speeds of horizontally and vertically traveling P waves ( $Vph(z)$  and  $Vpv(z)$ ), we calculate  $K_{Vp}^c$  as the sum of sensitivity kernels for  $Vph(z)$  and  $Vpv(z)$  (i.e., we assume  $\frac{\delta Vp(z)}{Vp(z)} = \frac{\delta Vph(z)}{Vph(z)} = \frac{\delta Vpv(z)}{Vpv(z)}$ ).

In a similar fashion, observed relative perturbations in amplification,  $\frac{\delta A(\omega)}{A_0(\omega)}$ , are assumed to vary linearly with relative perturbations in  $Vsv(z)$ ,  $Vp(z)$ , and  $\rho(z)$ :

$$\frac{\delta A(\omega)}{A_0(\omega)} = \int [K_{Vsv}^A(\omega, z) \frac{\delta Vsv(z)}{Vsv_0(z)} + K_{Vp}^A(\omega, z) \frac{\delta Vp(z)}{Vp_0(z)} + K_{\rho}^A(\omega, z) \frac{\delta \rho(z)}{\rho_0(z)}] dz. \quad (5)$$

Here  $K_{Vsv}^A$ ,  $K_{Vp}^A$ , and  $K_{\rho}^A$  are sensitivity kernels that describe the relative change in amplification due to a perturbation in  $Vsv(z)$ ,  $Vp(z)$ , or  $\rho(z)$ . Since these kernels cannot be calculated analytically, we compute them numerically by perturbing a single parameter, one depth layer at a time, and calculating how amplification differs from that of

the reference model, where amplification depends on the radial eigenfunction

$U_r$ , evaluated at the surface ( $z=0$ ) and group velocity,  $G$ :  $A(\omega) = \frac{U_r(\omega)}{\sqrt{G(\omega)}}$  (e.g., Ferreira & Woodhouse, 2007).

Our tests have shown that jointly inverting phase velocity and amplification should allow us to recover  $Vsv(z)$  in the crust and uppermost mantle and  $Vp(z)$  in the crust. We wish to solve for perturbations to the crustal  $Vp/Vs$  ratio  $\kappa(z) = \frac{Vp(z)}{Vsv(z)}$  rather than  $Vp(z)$ , and we substitute into Equations 4 and 5 using a chain rule expansion (e.g., Golos et al., 2020):

$$\frac{\delta Vp(z)}{Vp(z)} = \frac{\delta \kappa(z)}{\kappa(z)} + \frac{\delta Vsv(z)}{Vsv(z)}. \quad (6)$$

Since the joint inversion cannot constrain density, we prescribe a scaling factor  $\beta(z)$  relating  $\rho(z)$  perturbations to  $Vsv(z)$  perturbations. With these assumptions, Equations 4 and 5 become:

$$\frac{\delta c(\omega)}{c_0(\omega)} = \int \{ [K_{Vsv}^c(\omega, z) + K_{Vp}^c(\omega, z) + \beta(z)K_\rho^c(\omega, z)] \frac{\delta Vsv(z)}{Vsv_0(z)} + K_{Vp}^c(\omega, z) \frac{\delta \kappa(z)}{\kappa_0(z)} \} dz, \quad (7)$$

$$\frac{\delta A(\omega)}{A_0(\omega)} = \int \{ [K_{Vsv}^A(\omega, z) + K_{Vp}^A(\omega, z) + \beta(z)K_\rho^A(\omega, z)] \frac{\delta Vsv(z)}{Vsv_0(z)} + K_{Vp}^A(\omega, z) \frac{\delta \kappa(z)}{\kappa_0(z)} \} dz. \quad (8)$$

### The inverse problem

We determine the residual phase velocity  $\frac{\delta c(\omega)}{c_0(\omega)}$  and residual amplification  $\frac{\delta A(\omega)}{A_0(\omega)}$  by differencing the observed values from those predicted with Mineos for the reference model, which is updated after each iteration. Sensitivity kernels for  $Vsv(z)$  and  $Vp(z)$  are also updated after each iteration. We solve for the relative perturbations in  $Vsv(z)$  and  $\kappa(z)$  with Equations 7 and 8 using a LSQR iterative solver to minimize the cost function:



$$\epsilon = \|\mathbf{G}\mathbf{m} - \mathbf{d}\|^2 + \eta\|\mathbf{L}\mathbf{m}\|^2, \quad (9)$$

where  $\mathbf{G}$  is a matrix containing the integrated Rayleigh wave phase velocity and amplification sensitivity kernels,  $\mathbf{m}$  is the model perturbation containing  $\frac{\delta V_{sv}(z)}{V_{sv}(z)}$  and  $\frac{\delta \kappa(z)}{\kappa(z)}$ , and  $\mathbf{d}$  contains the observed  $\frac{\delta c(\omega)}{c_0(\omega)}$  and  $\frac{\delta A(\omega)}{A_0(\omega)}$ . The matrix  $\mathbf{L}$  is a first-derivative smoothing operator with a parameter  $\eta$  chosen to balance data misfit and model regularization.

The Vsv model is parameterized by 300 layers, each 1 km thick, and crustal Vp/Vs is treated as a single value for the entire crust. At depths  $> 400$  km, the model is identical to the Preliminary Reference Earth Model (PREM; Dziewonski & Anderson, 1981), and our model's parameters linearly transition into those of the PREM over the depth range 300-400 km. The initial reference model at the start of Step 1 has a constant Vsv of 4 km/s in all layers and is updated at every iteration.

#### 8.4 Text S3.

We perform tests with synthetic data to investigate how data noise and allowing for a two-layer crustal  $V_p/V_s$  model impact model recovery (Fig. S9). To assess the impact of noise, the Step 2 joint inversion is repeated 100 times, each time adding random noise to the synthetic amplification data with an uncertainty of 0.04 for all periods. We choose an uncertainty of 0.04 for these tests because it is greater than the amplification uncertainties for over 90% of stations, as calculated using equation (3) and shown in Figure S5. We find that even with data noise, the single-layer crustal  $V_p/V_s$  is nearly identical to the target value, and the bootstrapping-derived standard deviation is small (0.014). For the two-layer crustal  $V_p/V_s$  scenarios, similar values are recovered regardless of data noise, with  $V_p/V_s$  that is slightly higher than the target in the upper crust and lower than the target in the lower crust. The standard deviation is larger in the lower crust than the upper crust due to the lower sensitivity of amplification to deeper crustal  $V_p$ . Given the results of these tests, we focus on resolving one-layer crustal  $V_p/V_s$  models in the western U.S. with real data rather than attempting to constrain more complex two-layer configurations.

In Fig. S10 we show how the joint inversion of Rayleigh wave phase velocity and amplification better recovers the low  $V_{sv}$  of shallow sedimentary layers than inverting phase velocity alone. The target (input) model contains a 4-km-thick layer of  $V_{sv} = 2$  km/s underlain by a crust and upper mantle characterized by smoothly varying  $V_{sv}$ . The same  $V_p$  is used for the input and output models so that we can investigate the recovery of  $V_{sv}$  only. Inverting only phase velocities in the period range 25-180 s produces a  $V_{sv}$  model that overestimates  $V_{sv}$  in the sedimentary layer and underestimates  $V_{sv}$  in the crust. It also overpredicts amplification in the period range 25-80 s. When phase velocity

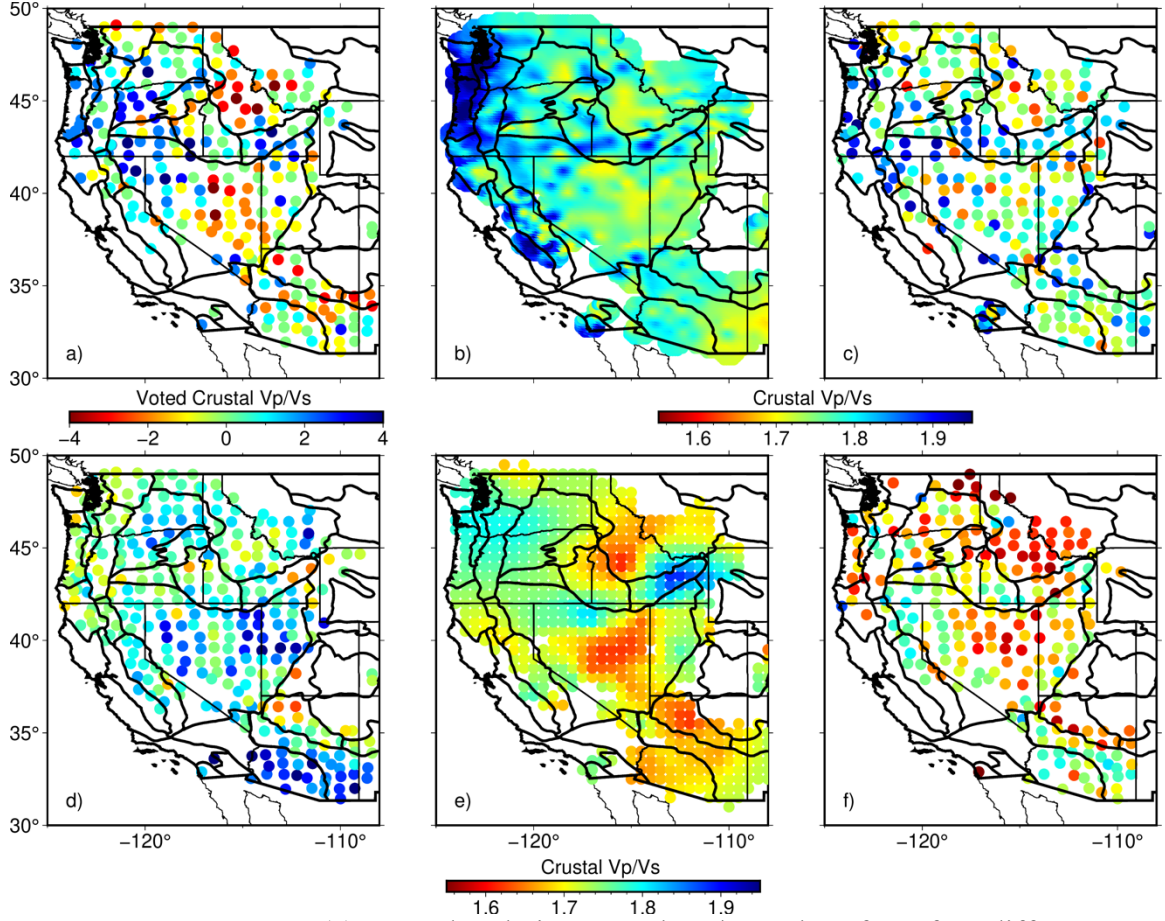
and amplification are jointly inverted, both data sets are well fit by the resulting  $V_{sv}$  model, and recovery of  $V_{sv}$  in the sediments and crust is much better than the phase-velocity-only inversion as a result of the higher sensitivity of amplification to shallower structure compared to phase velocity at the same periods (Fig. 1a,b). The recovery of mantle structure is similar in the two inversions. Although we do not solve for sedimentary structure in this study, we present the results of this test to show that jointly inverting both short-period phase velocities (5-20 s), which are well known to be sensitive to shallow structure, and intermediate-period amplification is a promising approach for improving models of sedimentary layers on continents.

## **8.5 Data Set S1.**

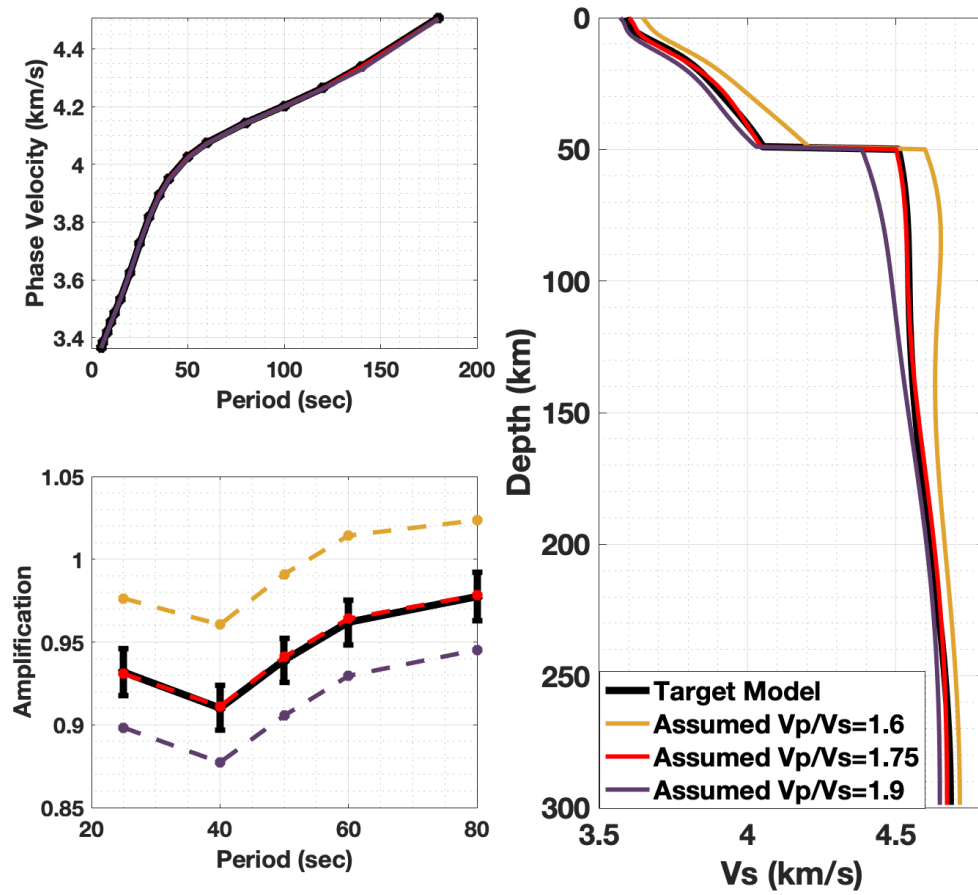
The accompanying .zip file contains 15 files. The files named with ‘Rphvel\_period’, where the variable periods range from 40 to 180 s, include the Rayleigh wave phase velocity maps generated in this study using Helmholtz tomography. The files named with ‘Rampl\_period’, where the variable periods range from 25 s to 80 s, include the Rayleigh wave amplification data, normalized to station TA.M07A. The file ‘Crustal\_Vp\_Vs\_ratio’ contains the crustal  $V_p/V_s$  model derived from the amplification measured in this study. Each file consists of three columns. The first two columns are longitude and latitude, and the third column represents phase velocity, amplification, or crustal  $V_p/V_s$ . The file ‘Vsv.nc’ is the  $V_{sv}$  model generated in this study, providing  $V_{sv}$  values for every station from the surface down to a depth of 300 km.

| Period (s) | RMS of residuals | $\xi_A$ |
|------------|------------------|---------|
| 25         | 0.0227           | 0.0131  |
| 40         | 0.0178           | 0.0109  |
| 50         | 0.0155           | 0.0102  |
| 60         | 0.0139           | 0.0099  |
| 80         | 0.0132           | 0.0103  |

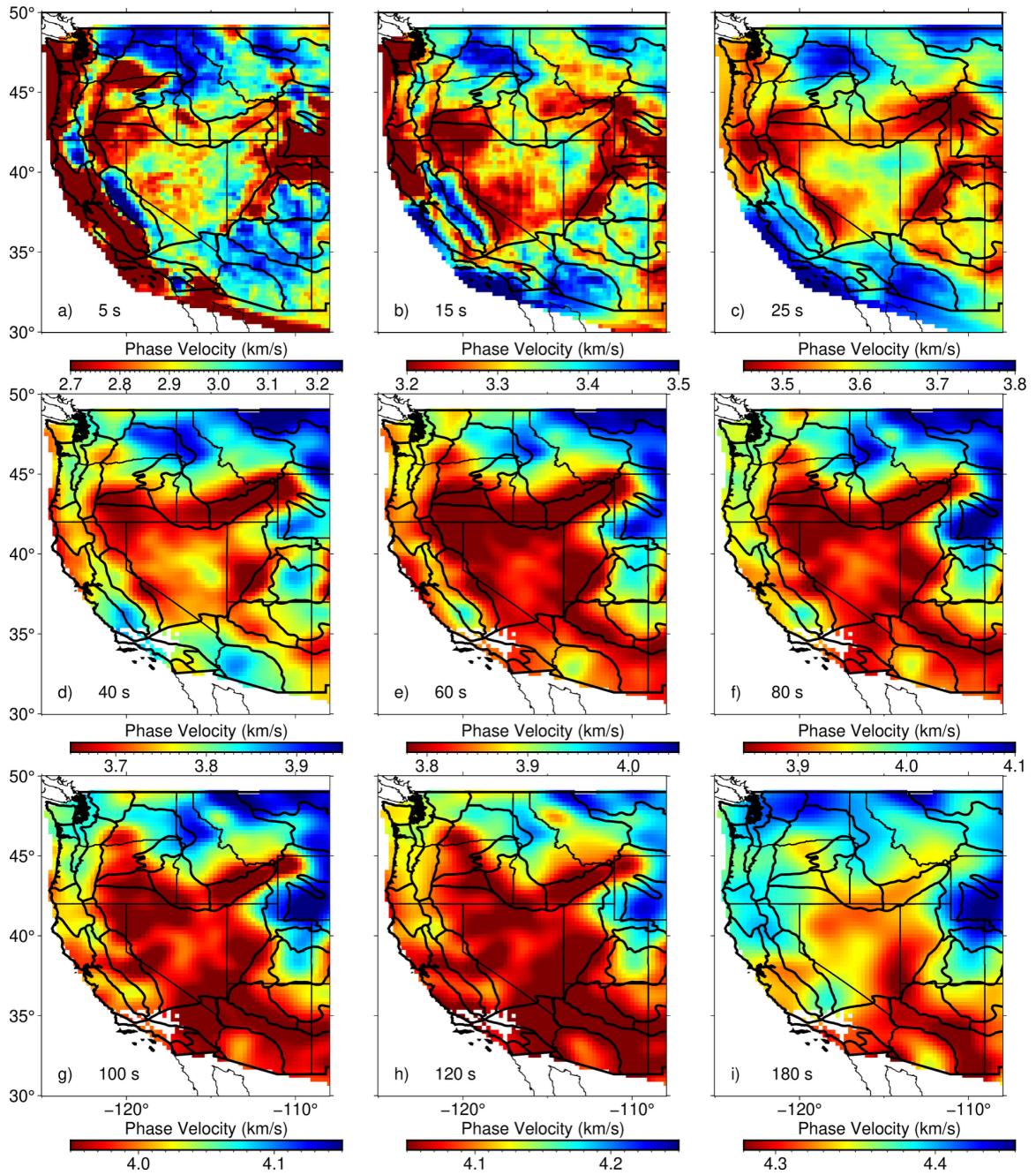
*Table 2.1: **Table S1.** Evaluation of amplification measurements as a function of period. From left to right, columns include: wave period; the root-mean-square (RMS) of the residual between the observed  $\bar{d}_{ij}$  values and the values predicted by the least-squares inversion for local station amplification factors  $A_{R,i}$  and  $A_{R,j}$ ; average amplification uncertainty derived from equation (3).*



**Figure 2.5: Figure S1.** (a) “Voted” relative crustal Vp/Vs values from four different models: (b) Ma & Lowry (2017) from joint inversion of Bouguer gravity anomalies and seismic receiver functions, (c) Sui et al. (2023) from two-layer receiver function H-k stacking method, (d) Buehler & Shearer (2014) from the Pn and Sn station terms, and (e) RSTT201404um using Pg and Sg phases (Begnaud et al., 2016). The “voted” map in (a) coarsely summarizes the level of consensus in the Vp/Vs variations among the four previous studies. The values from a given study are sorted into three bins (the lowest 35%, middle 30%, and highest 35%) and assigned values of -1, 0, or 1, respectively. The “voted” map is a sum of the four studies. Values of 3-4 (blue) indicate that the previous studies agree on relatively high Vp/Vs values and values of -4--3 (red) indicate agreement on relatively low values. (f) Crustal Vp/Vs results from this study.

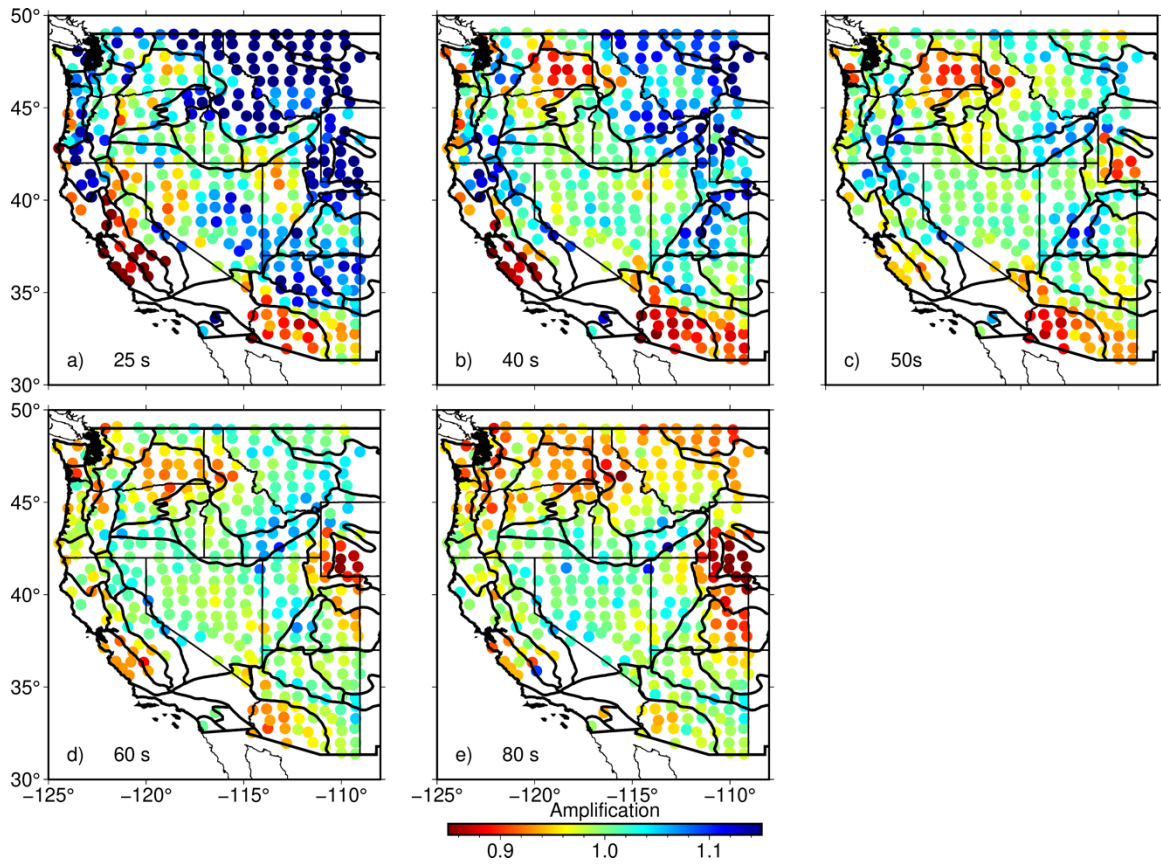


*Figure 2.6: **Figure S2.** Demonstration of how the assumed crustal  $V_p/V_s$  ratio affects recovery of  $V_{sv}$  in the crust and upper mantle. (a) Rayleigh wave phase velocity curves from the target (input) model (black), which has crustal  $V_p/V_s$  equal to 1.75. Other curves show predictions of models obtained from inversion of phase velocity only assuming crustal  $V_p/V_s=1.6$  (yellow), crustal  $V_p/V_s=1.75$  (red), and crustal  $V_p/V_s=1.9$  (purple). (b) As in (a) but for Rayleigh wave amplification. (c) The input and recovered  $V_{sv}$  models. All three recovered models fit the phase velocity data equally well. But the amplification data can discriminate between the three models, and the wrong assumed crustal  $V_p/V_s$  value introduces error into the  $V_{sv}$  recovery.*



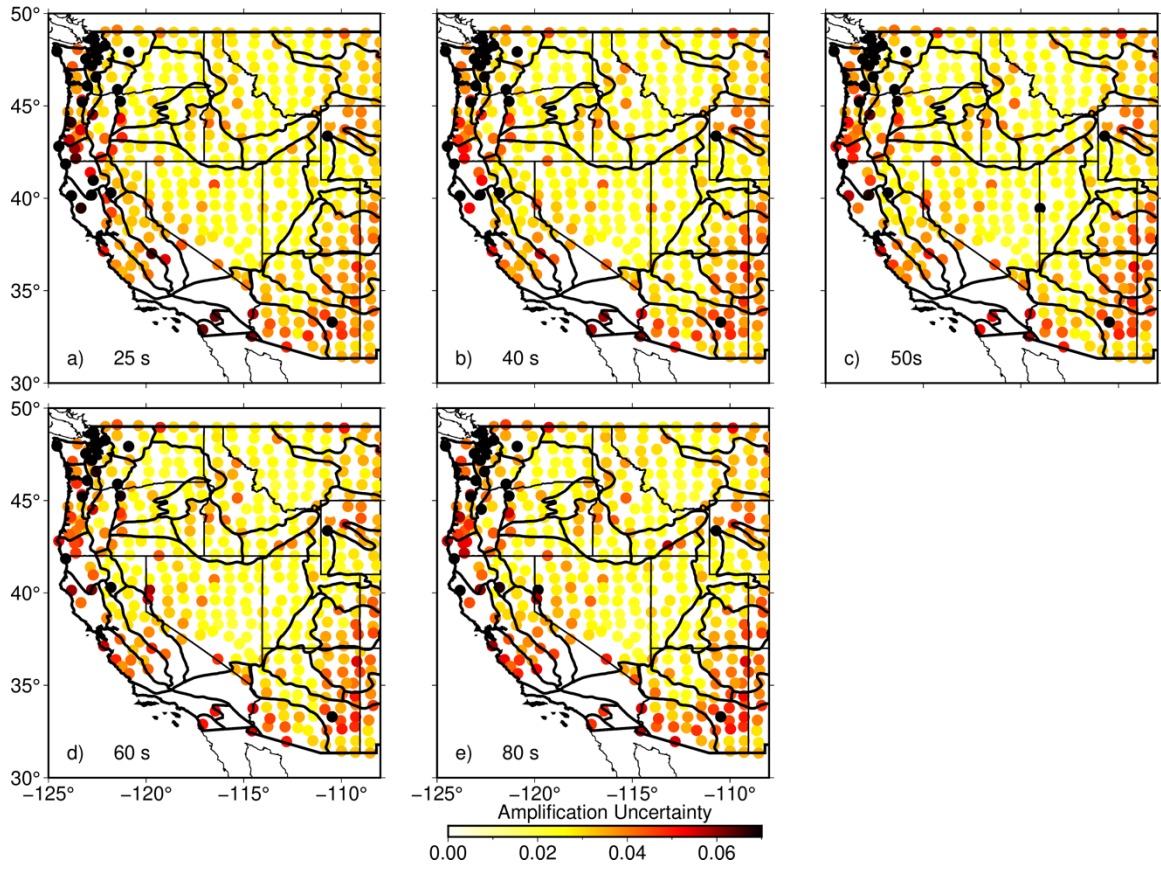
**Figure 2.7: Figure S3.** Rayleigh wave phase velocity maps from USANT15 (Ekström, 2017) at (a) 5 s, (b) 15 s, (c) 25 s. Rayleigh wave Helmholtz phase velocity maps at (d) 40 s, (e) 60 s, (f) 80 s, (g) 100 s, (h) 120 s, and (i) 180 s from this study.



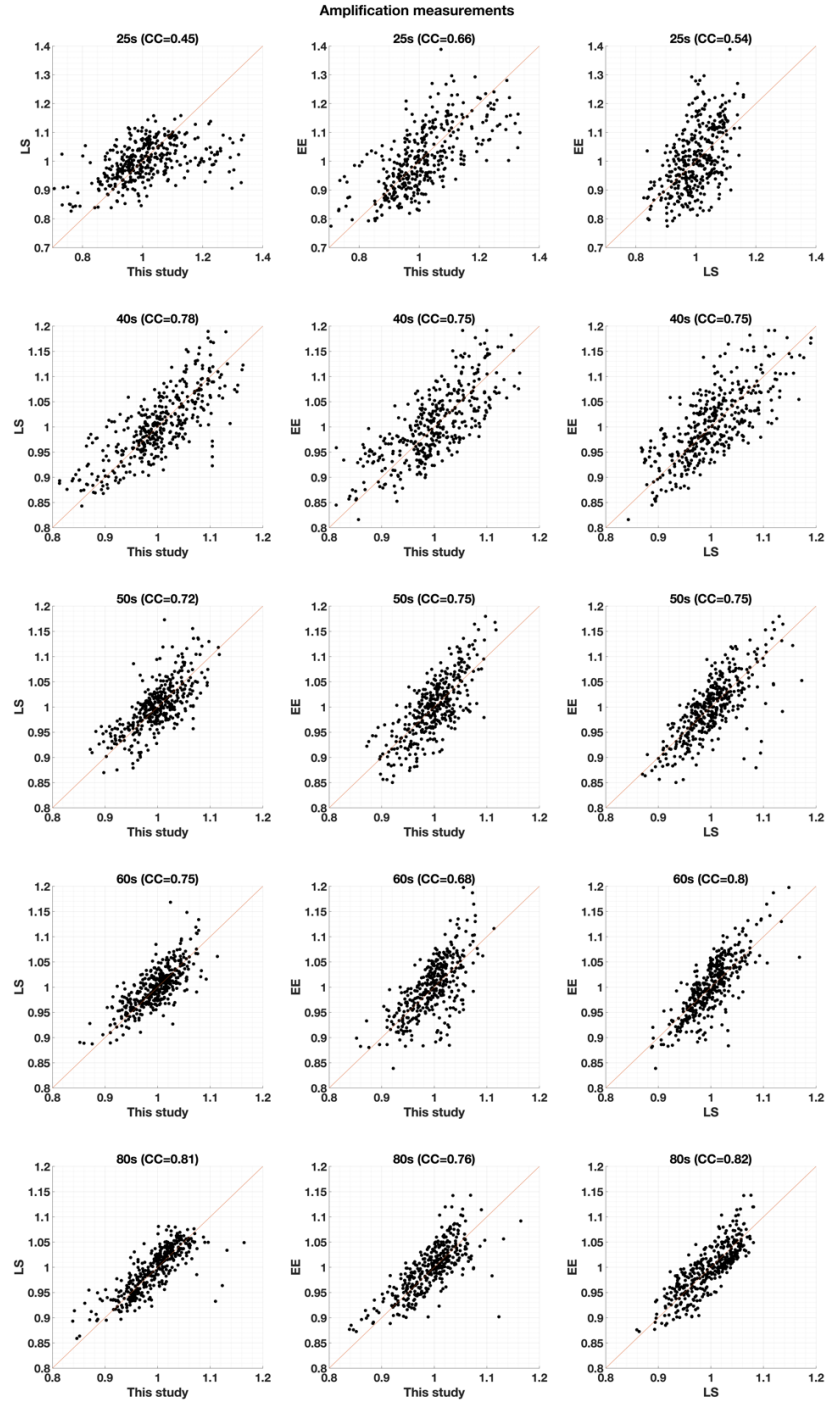


*Figure 2.8: Figure S4.* Measured Rayleigh wave amplification maps at (a) 25 s, (b) 40 s, (c) 50 s, (d) 60 s, (e) 80 s used in this study. All are shown relative to the value at station TA.M07A.

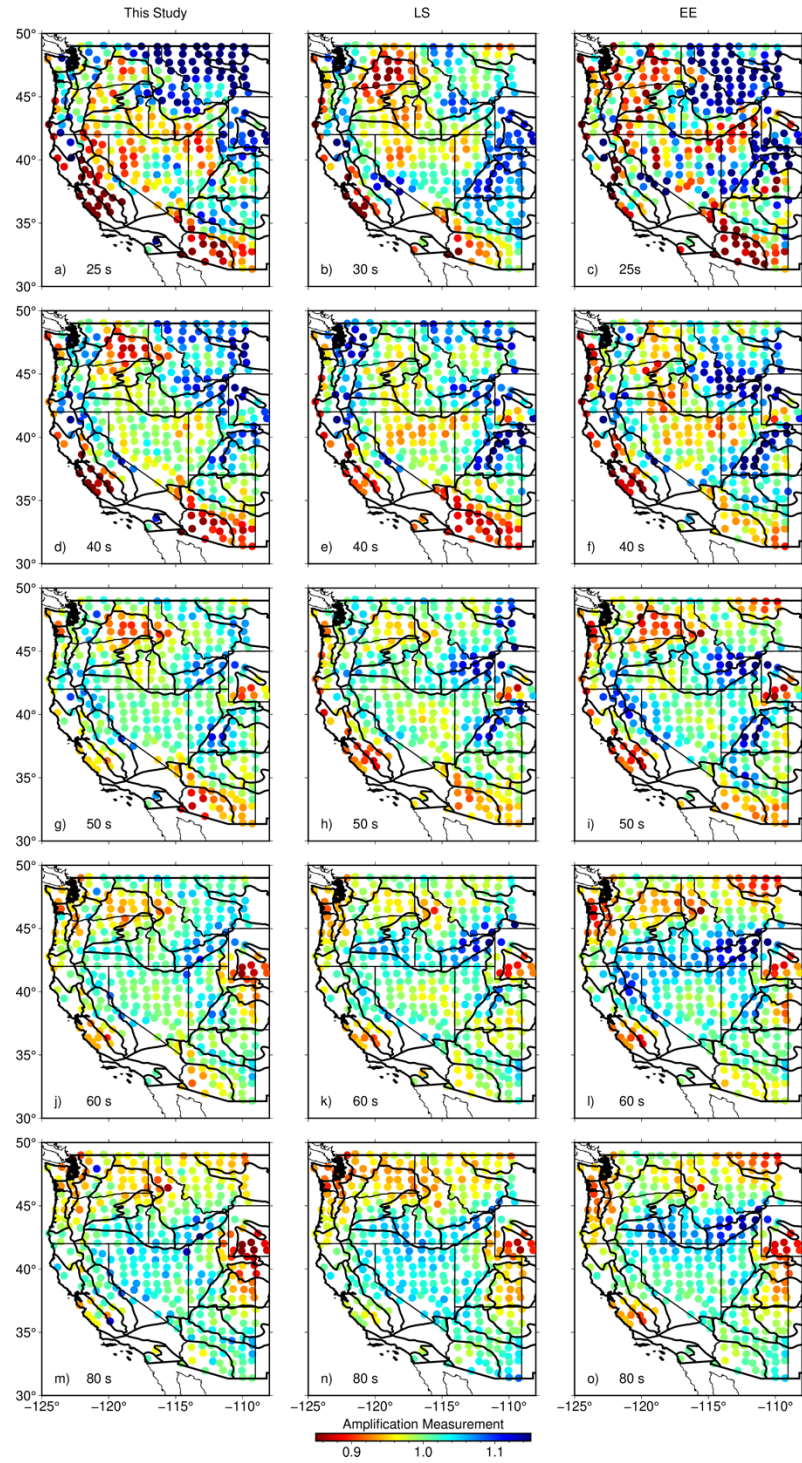




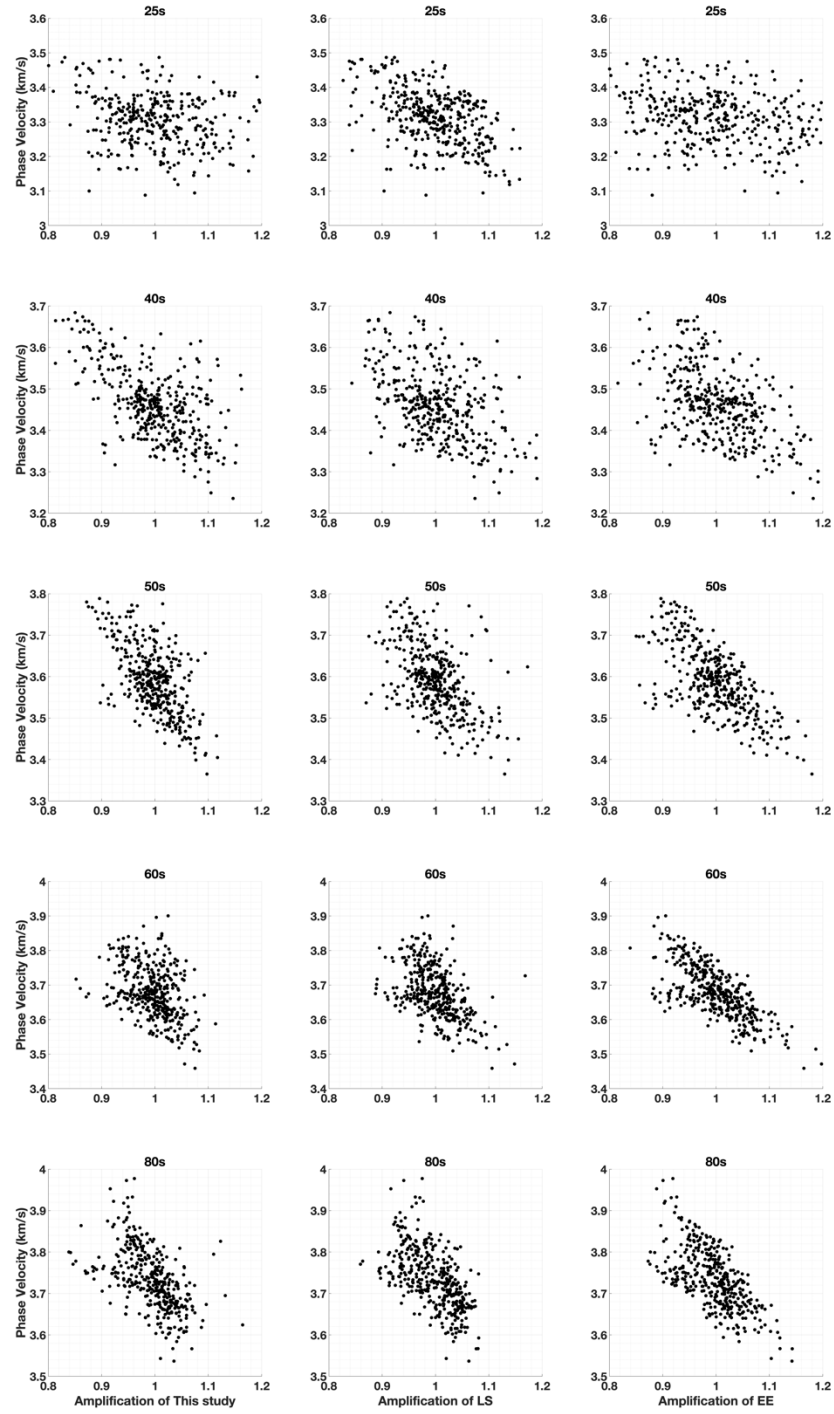
*Figure 2.9: Figure S5. Measured Rayleigh wave amplification uncertainty maps at (a) 25 s, (b) 40 s, (c) 50 s, (d) 60 s, (e) 80 s based on equation (3).*



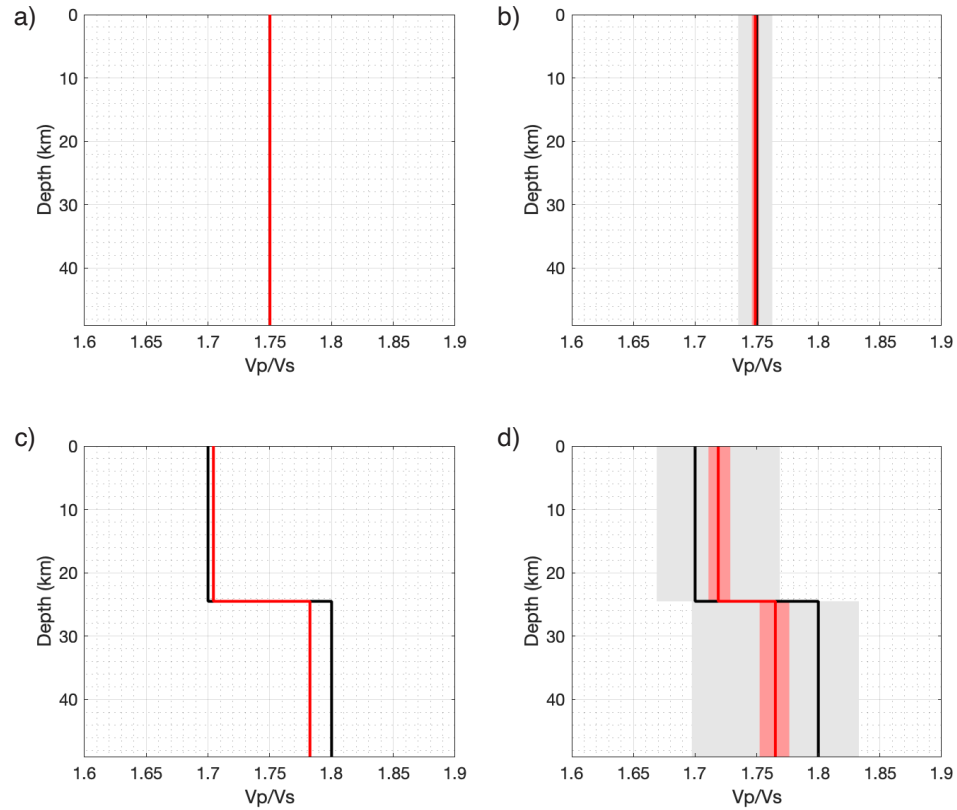
*Figure 2.10: **Figure S6.** Comparisons of amplification measurements: this study vs. LS (a hybrid of Lin et al., 2012, and Schardong et al., 2019) on the left, this study vs. EE (Eddy and Ekström, 2020) in the center, and LS vs. EE on the right. The 1:1 line is also plotted. Periods and correlation coefficients are noted at the top of each figure. For this comparison, the amplification values are shown relative to the median value at each period for each study.*



*Figure 2.11: **Figure S7.** Measured amplification from three studies, in addition to this study on the left: LS (a hybrid of Lin et al., 2012, and Schardong et al., 2019) in the center, and EE (Eddy and Ekström, 2020) on the right. Periods are indicated in the bottom-left corner of each map. Amplification is shown relative to the median value at each period for each study in order to highlight variations.*



*Figure 2.12: **Figure S8.** Comparisons of our phase velocity maps (Fig. S3) and the three amplification data sets: this study (left), LS (middle; a hybrid of Lin et al., 2012, and Schardong et al., 2019), and EE (right; Eddy and Ekström, 2020). Periods are indicated at the top of each figure. All amplification values are shown relative to the median value at each period for each study*



*Figure 2.13: **Figure S9.** Uncertainty tests with synthetic data. (a) Recovery of single-layer crustal  $V_p/V_s$  without adding noise to the synthetic data. Black line shows the target (input) value and red line is the recovered value from the Step 2 joint inversion. (b) Same as (a), but the synthetic data have random noise with uncertainty of 0.04, the same statistical characteristics as the real data from the western U.S. We conduct 100 inversions and show the mean crustal  $V_p/V_s$  with the red line, the 95% confidence interval with the red area, and the standard deviation from the mean value with the gray area. (c,d) As in (a,b) but for a two-layer crustal  $V_p/V_s$  model.*

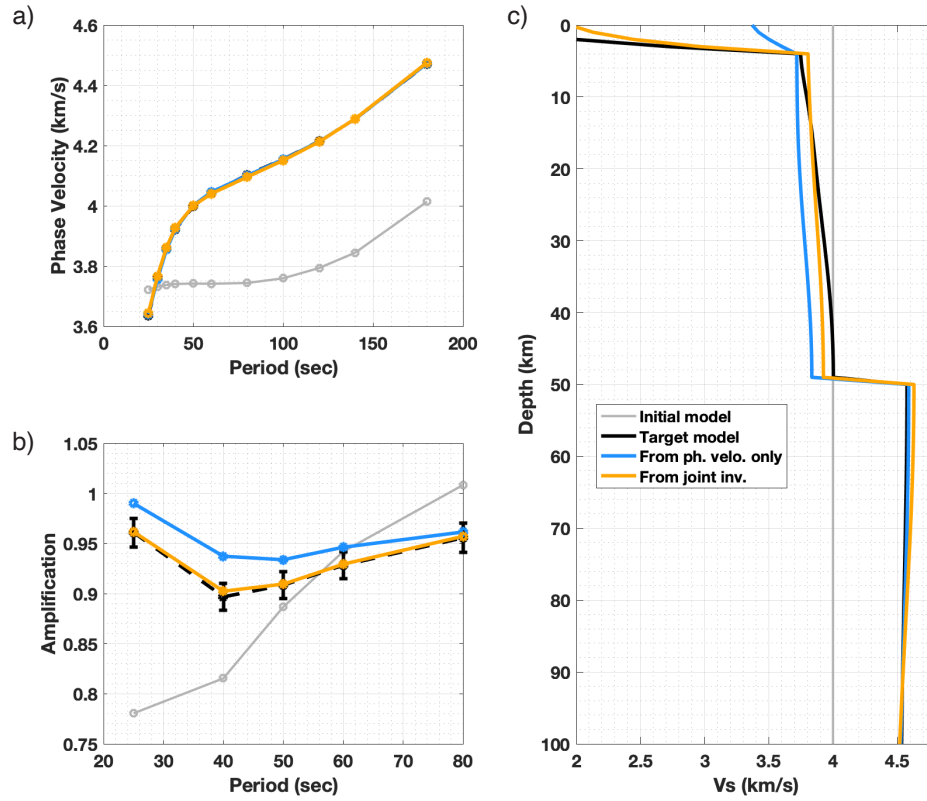
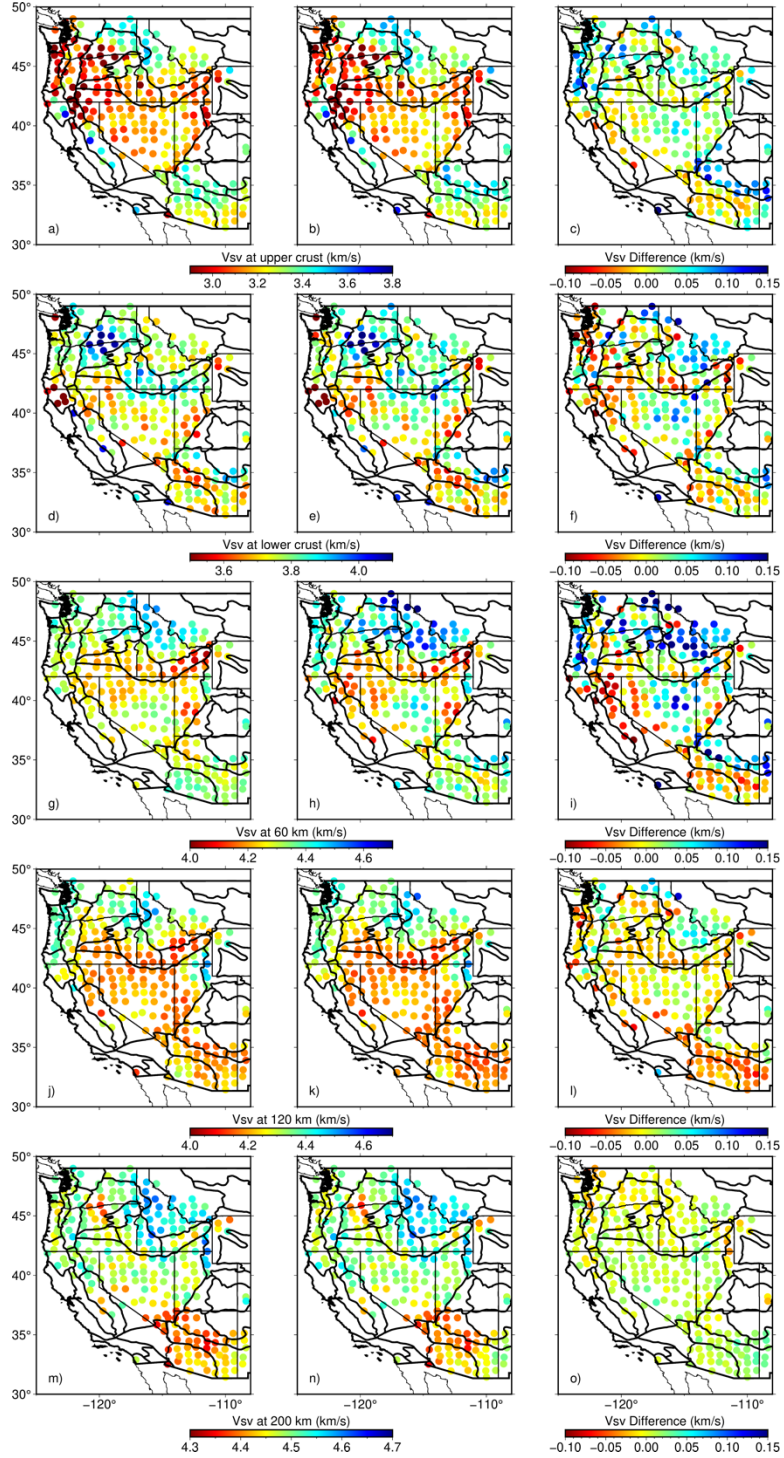
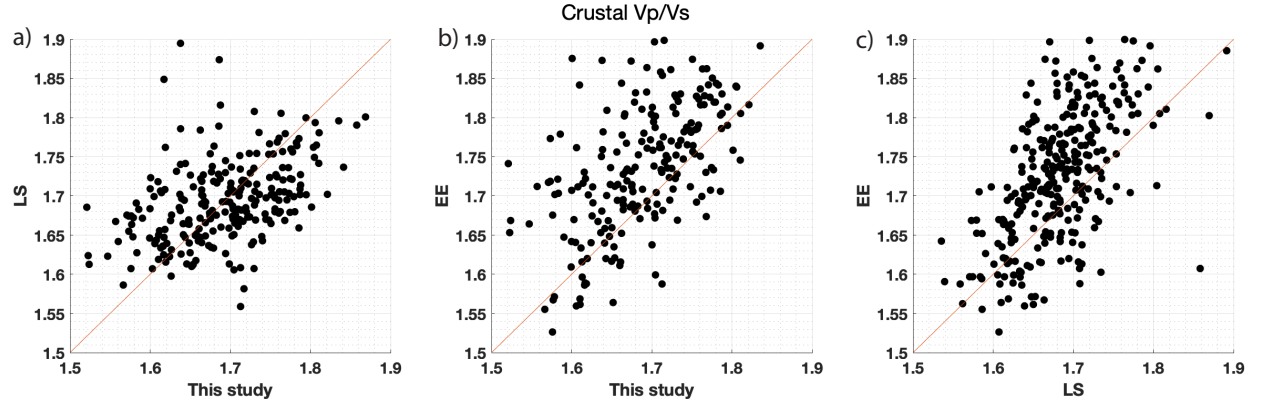


Figure 2.14: **Figure S10.** Results of synthetic experiment to investigate recovery of  $V_{sv}$  when a sedimentary layer is present, holding  $V_p$  fixed. Black=target (input) model, gray=initial model, blue=inverting only phase velocity, yellow=inverting phase velocity and amplification. (a) Rayleigh wave phase velocity. (b) Rayleigh wave amplification. (c) Depth-dependent  $V_{sv}$ .





*Figure 2.15: **Figure S11.** Left column: Vsv from Step 1 inversion of phase velocity data only. Middle column: Vsv from Step 2 joint inversion of phase velocity and amplification data from this study. Right column: the difference between the middle and left columns (i.e., Step 2 – Step 1). From top to bottom, rows are results for upper crust, lower crust, 60 km, 120 km, and 200 km, respectively.*



*Figure 2.16: **Figure S12.** Comparisons of inverted crustal  $V_p/V_s$  based on amplification measurements from: (a) this study vs. LS (a hybrid of Lin et al., 2012, and Schardong et al., 2019), (b) this study vs. EE (Eddy and Ekström, 2020), and (c) LS vs. EE.*



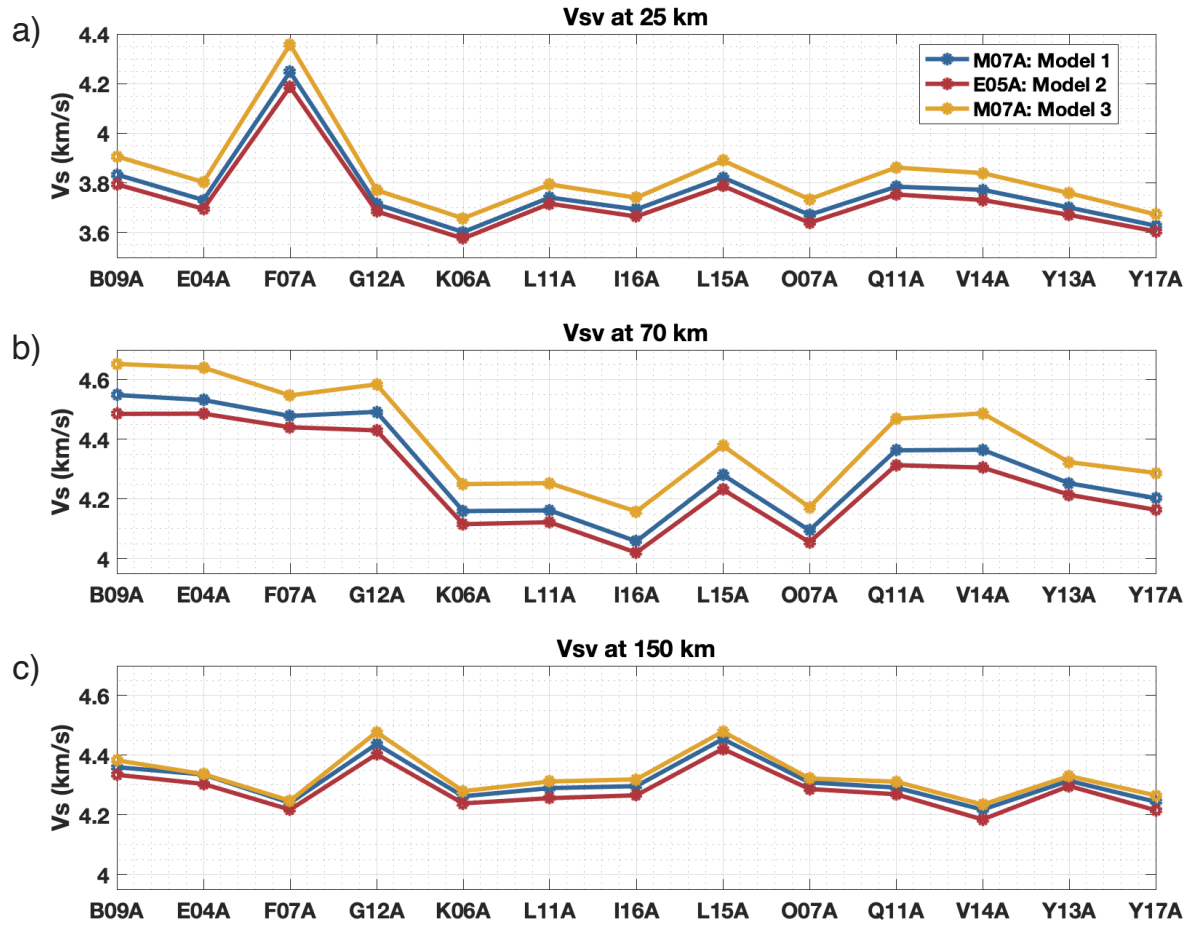


Figure 2.17: **Figure S13.** Similar to Fig. 4b, demonstration of how Vsv at 13 stations after the final Step 2 inversion depends on the normalization station and velocity model used to convert the relative amplification observations into absolute amplification. The tests use the normalization station TA.M07A in Nevada with two different elastic models and normalization station TA.E05A in southern Washington with a third model. (a) Vsv at 25 km, (b) Vsv at 70 km, and (c) Vsv at 150 km.

## **CHAPTER 3| Investigating Layered Cratonic Lithosphere in the Fennoscandian Shield**

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## Abstract

Cold, thick, and chemically depleted lithosphere with high seismic velocity and low attenuation characterizes continental cratons. However, discrepancies between xenolith-based thermal constraints and those inferred from surface-wave seismic models, together with the widespread observation of mid-lithosphere discontinuities (MLDs), challenge the traditional view of a simple cratonic lithospheric structure. Improving constraints on lithospheric layering is therefore essential for understanding the thermal and compositional structure of the cratonic upper mantle. In this study, we develop a three-dimensional model of vertically polarized shear-wave velocity ( $V_{sv}$ ) in Fennoscandia through the joint inversion of Rayleigh wave phase velocities and S-to-P receiver functions, enabling resolution of both absolute velocities and strong vertical gradients. Compared to models derived from phase velocities alone, our results show that the highest velocities ( $>4.7$  km/s) occur predominantly at depths of  $\sim 140$ – $210$  km. We compare the resulting seismic model in the Fennoscandia craton interior with  $V_{sv}$  predictions based on depleted peridotite compositions along xenolith-derived geotherms. After accounting for anelastic effects and radial anisotropy, we find that only a moderate diamond fraction ( $\sim 0.5$  vol.%), combined with a colder, corrected geotherm, can explain the high velocities observed in the lower lithosphere below  $\sim 150$  km. Reduced observed shear-wave velocities in the shallow lithosphere are best explained by the combined effects of radial anisotropy and the presence of carbonate and/or hydrous minerals introduced by metasomatic processes. While the cratonic interior of southern Finland exhibits a clear two-layered structure, northern Finland shows a similar pattern but includes an additional low-velocity zone below  $\sim 200$  km, which may reflect refertilized

lithosphere. The layered seismic velocity structure revealed in this study provides new insights into the thermal and compositional architecture of cratonic lithosphere and has important implications for its formation and modification.

# 1 Introduction

Cratons, the tectonically stable cores of continents since their formation in the Precambrian, are characterized by several well established geophysical and geochemical properties. These includes low surface heat flow (e.g., [Hamza and Vieira 2012](#)), which suggests low temperature contributing to thick lithosphere with high seismic velocity and low attenuation (e.g., [Gung et al. 2003](#); [Dalton et al. 2009](#); [Steinberger and Becker 2018](#); [Fischer et al. 2020](#)); high viscosity due to low temperature and low water content ([Hirth et al. 2000](#)); and high Mg# (magnesium number) in mantle peridotite, indicative of extensive melt depletion and adding buoyancy to the cratonic mantle (e.g., [Lee et al. 2011](#)). These features have long supported a relatively simple picture of cratonic mantle lithosphere as cold, dry, compositionally depleted, and mechanically strong.

However, recent observations from multiple disciplines reveal several puzzles that challenge this canonical view. First, reconciling xenolith-based constraints on thermal structure with surface-wave seismic models of cratonic lithosphere remains difficult. Surface-wave inversions commonly yield very high seismic velocities and positive velocity gradients with depth immediately below the Moho. Depth profiles of seismic velocity predicted from xenolith- and heat-flow-constrained geotherms tend to contain lower velocities and negative velocity gradients with depth ([Garber et al. 2018](#); [Bruneton et al. 2004](#); [Hirsch et al. 2015](#)). These discrepancies have motivated hypotheses invoking non-peridotitic components in the cratonic mantle lithosphere, such as eclogite or diamond ([Garber et al. 2018](#)), in addition to thermal effects. Second, although surface waves have limited resolution for detecting fine-scale lithospheric layering, body-wave

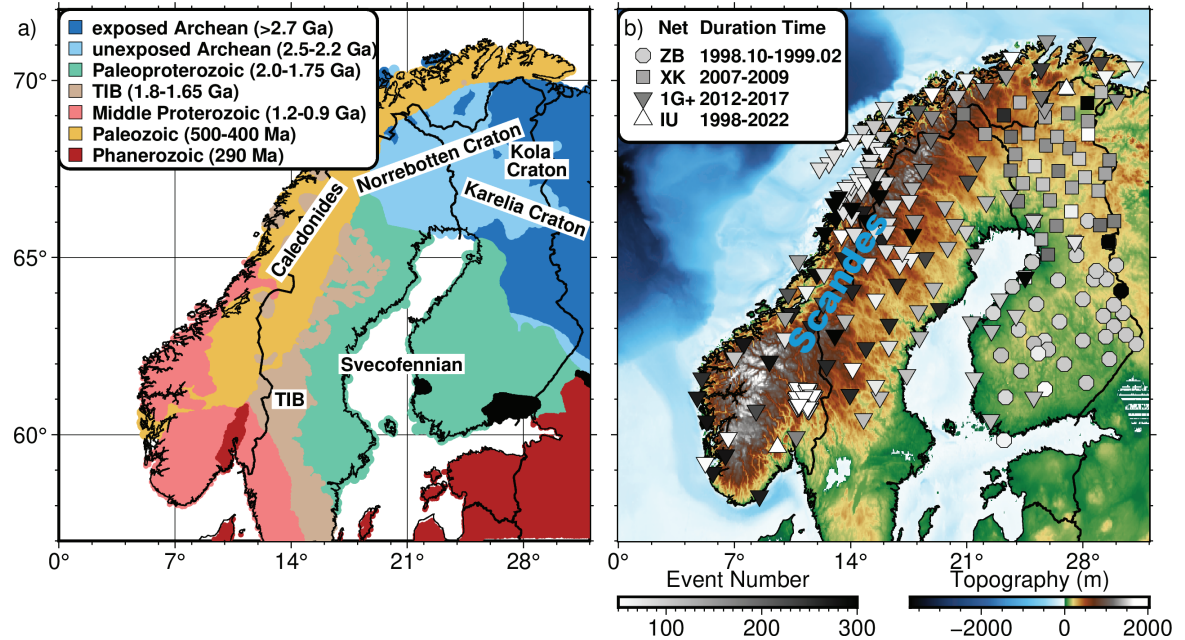
studies suggest that cratonic mantle lithosphere is not a simple, contiguous high-velocity body with uniform chemical composition. Many cratons worldwide exhibit negative velocity gradients (NVGs) at depths of ~60–120 km, with velocity reductions of ~1–12% (Abt et al. 2010; Ford et al. 2010; T. Liu and Shearer 2021; Carr and Olugboji 2024; Savage and Silver 2008; Miller and Eaton 2010; Wölbern et al. 2012; Sun and Kennett 2017; Rychert and Shearer 2009; Chen 2017). These features are commonly referred to as mid-lithosphere discontinuities (MLDs). Because the lithosphere–asthenosphere boundary beneath cratons is generally inferred to lie much deeper (e.g., ~200–250 km), the MLD is often interpreted as the top of a low-velocity zone embedded within the mantle lithosphere, complicating the traditional picture of cratonic structure.

The global prevalence, origin, and relationship between MLDs and craton stability is still under debate. Globally, Sp and Ps receiver-function studies suggest that NVGs are common but intermittent, with occasional positive velocity gradients also observed (Krueger et al. 2021; Carr and Olugboji 2024; Legre and Olugboji 2025), although some studies argue that certain NVGs may arise from data-processing artifacts (Kind et al. 2020). Proposed mechanisms for the velocity reductions associated with MLDs include partial melting (e.g., Perchuc and Thybo 1996; Liu et al. 2023), anomalous compositions involving carbonate or hydrous minerals (e.g., Rader et al. 2015; Selway et al. 2015; Eeken et al. 2018; Boyce et al., 2024), seismic anisotropy (e.g., Wirth and Long 2014; Wittlinger and Farra 2007), and enhanced attenuation due to elastically accommodated grain boundary sliding (e.g., Karato 2012). Moreover, numerical simulations indicate that the thickness, composition, and viscosity of a low-velocity layer with an MLD as its upper boundary can significantly influence craton longevity and

stability (e.g., [Fu and Li 2024](#); [Liu et al. 2018](#)). These findings highlight the importance of better constraining cratonic mantle lithosphere structure to understand the formation and long-term evolution of continents.

Fennoscandia is an ideal place to investigate cratonic mantle lithosphere. It contains extensive Archean and Paleoproterozoic terranes and benefits from dense seismic coverage from two-dimensional temporary arrays such as SVEKALAPKO ([Bock et al. 2001](#)), LAPNET ([Kozlovskaya 2007](#)), and ScanArray ([Thybo et al. 2012](#)). Fennoscandia represents the northwestern segment of the East European Craton, with remnants of Caledonian nappes preserved along its western margin (Fig. 1). The Fennoscandian (or Baltic) Shield formed through multiple episodes of crustal accretion beginning around 3.5 Ga, involving the Archean Karelian Province, the late Archean–Paleoproterozoic Lapland–Kola mobile belt to the northeast, and the Proterozoic Svecofennian domain to the southwest ([Gorbatshev and Bogdanova 1993](#)). These domains were shaped by the Lapland–Kola orogeny (1.94–1.86 Ga; [Daly et al. 2006](#)) and the Svecofennian orogeny (1.92–1.89 Ga; [Lahtinen et al. 2008](#)), respectively. The Svecofennian belt was subsequently truncated by emplacement of the Transscandinavian Igneous Belt (TIB; ~1.8–1.65 Ga), and its southwestern margin was reworked during the Sveconorwegian–Grenvillian orogenic events (~1.2–0.9 Ga; [Gorbatshev and Bogdanova 1993](#)). Later tectonic evolution includes closure of the Iapetus Ocean (between 430 and 410 Ma) due to the collision between Baltic and Laurentia during the Caledonian orogeny (430–410 Ma), forming the Scandinavian Caledonides along the western margin of the shield ([Torsvik and Cocks, 2005](#); [Roberts 2003](#)), following closure of the Tornquist Ocean in the south (440 Ma; [Cocks and Torsvik 2006](#)). Finally, the opening of the North Atlantic

Ocean led to separation of the Caledonides and formation of the present-day passive continental margin (Osmundsen and Ebbing 2008).



**Figure 3.1: Figure 1.** (a) Simplified geological map showing the major tectonic units in Fennoscandia and surrounding regions. Black lines indicate national borders. TIB= Transscandinavian Igneous Belt. (b) Topography map based on SRTM15+ model (Tozer et al. 2019) as well as the locations of all temporary and permanent seismic stations used for Rayleigh wave phase velocity measurements. Different symbol shapes are used to distinguish between the four main networks and are color coded by the number of earthquakes recorded and analyzed for our Rayleigh wave analyses. Network code 1G+ indicates the ScanArray core and affiliated deployments as described in Section 2.1.

Several key questions regarding cratonic lithosphere structure in Fennoscandia remain unresolved. First, as in other cratons, models assuming constant mantle composition and expected cratonic geotherms cannot fully explain observed Rayleigh wave phase-velocity dispersion curves in the southern Finland craton nucleus, implying additional complexity in either thermal or compositional structure (Pedersen et al. 2009). However, previous regional studies have struggled to distinguish between a smoothly varying velocity gradient and a layered lithospheric structure (Bruneton et al. 2004),



highlighting the need for improved seismic resolution. Second, in northern Finland, both the presence and depth extent of a low-velocity zone below  $\sim 150$  km, as well as the magnitude of the velocity reduction, remain controversial, leading to divergent interpretations of lithospheric structure near the craton margin (e.g., [Vinnik et al. 2016](#); [Pedersen et al. 2013](#); [Mauerberger et al. 2022](#)).

In this study, we integrate surface-wave and body-wave observations to better constrain the lithospheric structure of Fennoscandia, resolving shear-wave velocity features that cannot be robustly identified using a single dataset. Our results provide evidence for lithospheric structure with at least two layers in the upper mantle. We then compare the observed shear-velocity structure with predictions from thermodynamic forward modeling based on xenolith-inferred compositions and geotherms. By systematically evaluating potential causes of discrepancies between the seismic observations and petrologic predictions, we explore the implications of the Fennoscandian case for understanding the formation, modification, and long-term stability of layered cratonic lithosphere globally.

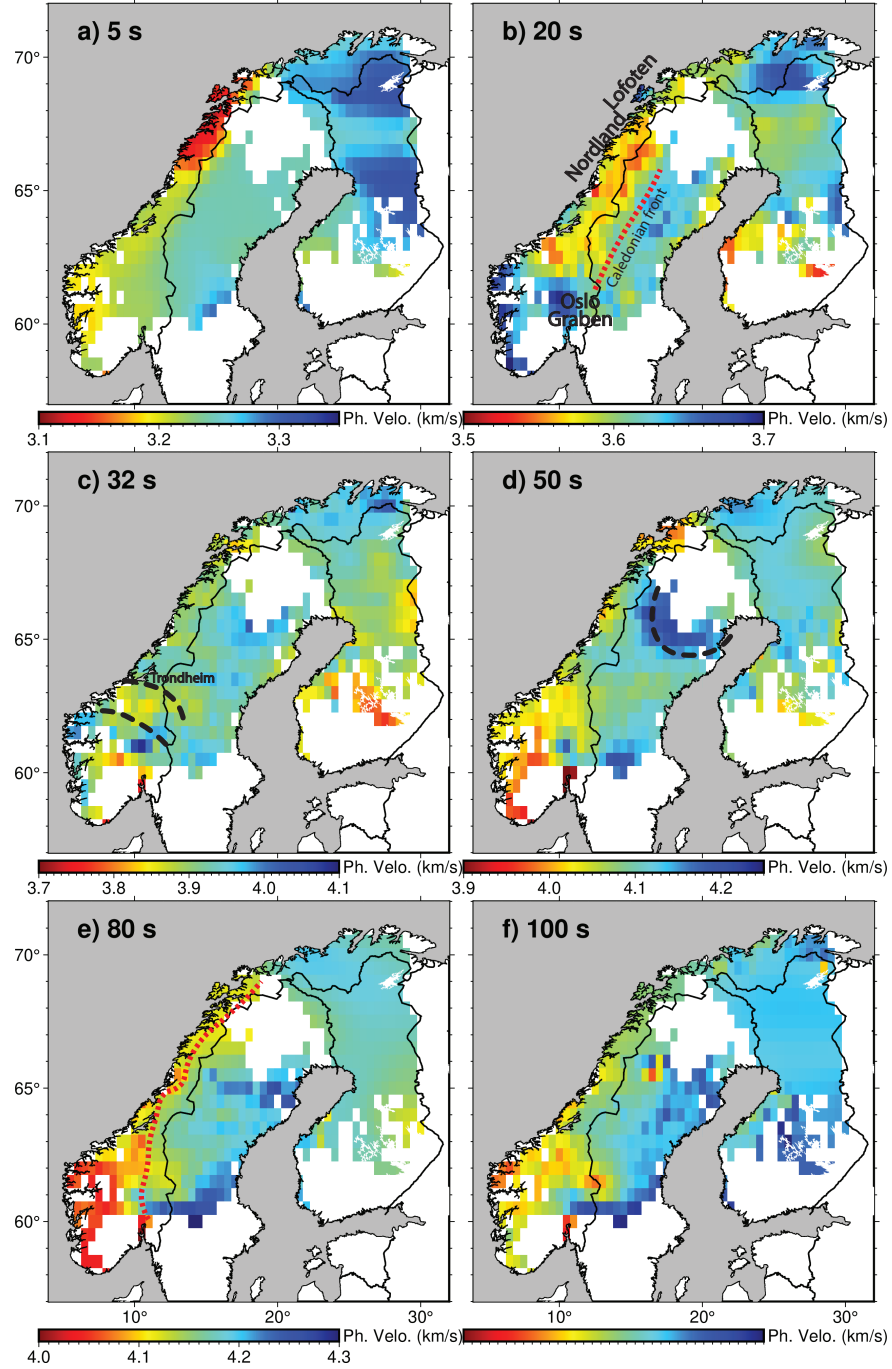
## **2 Data and Method**

### **2.1 Rayleigh Wave Phase Velocities**

To constrain the absolute shear wave velocity structure, we incorporate fundamental -mode Rayleigh wave phase velocities derived from ambient noise and teleseismic earthquakes in the period range 5-100 s. We develop maps of phase velocity from earthquakes by utilizing waveforms recorded by three seismic networks that operated during different time periods and collectively cover the study region. A total of 203

stations from the ScanArray international collaborative program (Hans Thybo et al. 2021) provide coverage from 2012 to 2017 across Norway, Sweden, and Finland. These include the ScanArray Core deployment (1G; Thybo et al. 2012) and affiliated deployments NRPNOR2 (2D; [Ottemoller and Lindholm, 2013](#)); SCANLIPS3D (ZR; [England, 2013](#)); Norwegian permanent (NS; [University of Bergen 1982](#)); and NORSAR (NO; [Norsar 1971](#)), where the two-character network code is given before each reference. In addition, 38 stations from the earlier SVEKALAPKO network (ZB; [Bock et al. 2001](#)) and 40 stations from the POLENET/LAPNET experiment (XK; [Kozlovskaya 2007](#)) supplement coverage in Finland, spanning 1998-1999 and 2007-2009, respectively. The distribution of stations used for measuring Rayleigh wave phase velocities is shown in Fig. 1b.

We include teleseismic earthquakes with magnitudes  $M_w > 6$  and epicentral distances up to  $120^\circ$ , recorded by the respective networks during their operation periods. For the short-lived ZB network, we relax the magnitude threshold to  $M_w > 5.5$  to incorporate more events. A total of 669 earthquakes are included in our analysis. We measure Rayleigh wave phase delays using the Automated Surface Wave Measuring System (ASWMS) software (Jin and Gaherty 2015) and calculate phase velocity maps using Eikonal tomography (Lin et al. 2009) at eight periods from 20 to 100 s. We first generate a phase velocity map for each event and then create a single composite map at each period by taking the mean value over all events in each grid cell. We follow [Lin et al. \(2009\)](#) to estimate the uncertainty in each cell by first calculating the standard deviation of slowness for each cell and then multiplying it by the composite phase velocity. We use a regular grid cell size of  $0.5^\circ \times 0.5^\circ$ , and only grid cells with more than 10 events are retained. No smoothing is applied to the composite phase velocity maps, except in



**Figure 3.2: Figure 2.** (a) Rayleigh wave phase velocity map at 5 s derived from ambient noise (Mauerberger et al., 2022; Köhler et al., 2012, 2015). (b) Weighted average phase velocity map at 20 s combining ambient noise and teleseismic earthquakes measurements as described in the text. (c-f) Eikonal phase velocity maps at 32-100 s derived from teleseismic earthquakes. The two red dashed lines in (b) and (e) mark the transitions described in the main text. Black dashed lines in (c) and (d) outline the velocity anomalies described in the main text.

subregions of the map where at least half of the grid cells have a large standard deviation of phase velocity, where a subregion is defined as a cap of  $2^\circ$  radius and a large standard deviation is defined as  $>0.12$  km/s.

To better constrain shallow structure, we include higher-frequency Rayleigh wave phase velocities derived from ambient noise tomography from 5 to 25 s (Mauerberger et al. 2022; Köhler et al. 2012, 2015). For the periods of 20 s and 25 s where both earthquake and ambient-noise maps are available, we calculate the weighted average of the teleseismic and ambient noise phase velocities, applying weights of 0.5 and 0.8 to the teleseismic measurements at 20 s and 25 s, respectively. The final Rayleigh wave phase velocity maps are displayed in Fig. 2 and uncertainty maps are shown in Fig. S1. The large uncertainty at the edges of study region, particularly at short periods, is likely associated with the limited number of measurements available for extracting ambient noise dispersion curves.

## **2.2 Body Wave Receiver Functions**

To enhance the vertical resolution of layering structure of the cratonic lithosphere, we incorporate 1-D Sp receiver functions, which we derived from 3-D common conversion point (CCP) stacks. In addition to the networks used for Rayleigh wave phase velocity measurements, we include data from 187 additional permanent and temporary seismic stations, with network codes: 7E (Wilde-Piórko et al. 2009), DK (GEUS Geological Survey of Denmark and Greenland 1976), EE (Geological Survey Of Estonia (EGT) 1996), FN (Sodankylä Geophysical Observatory / University of Oulu (Finland) 1980), HE (Institute Of Seismology, 1980), ZA (GEOFON Program), ZC (GEOFON

Program), ZS (GEOFON Program), ZW(Balling et al. 2011), GE(GEOFON Data Centre 1993) , Y1 (SCANLIPS2), YF (SCANLIPS), YV (North East Atlantic Tomography (NEAT)), YY (Norway 05), ZR (Richard England 2013b), and ZU (PASSEQ). We select earthquakes with magnitude  $M_w > 6$ , occurring between 1987 and 2022 and with epicentral distances from  $30^\circ$  to  $90^\circ$ . The seismograms are filtered using a 4-100 s band-pass filter and rotated to P-SV components using a free surface transformation (Kennett 1991) with free surface velocities determined following [Hua et al. \(2020\)](#). Sp receiver functions are obtained for data from a more limited distance range ( $55^\circ$  to  $85^\circ$ ) via a time-domain deconvolution method (Ligorria and Ammon 1999), by deconvolving the direct S arrival on the SV component from the P component to minimize side-lobes (Krueger et al. 2021). We use the global velocity model GLAD-M25 (Lei et al. 2020) to migrate Sp receiver functions from the time domain to the depth domain.

Compared with Ps receiver functions, Sp receiver functions are more effective for resolving vertical shear wave velocity gradients in the upper mantle, as they avoid the crustal reverberations (e.g., [Eilon et al. 2018](#)). However, Sp receiver functions typically exhibit lower signal-to-noise-ratios than Ps receiver functions. To ensure data quality of individual Sp receiver functions while preserving as much usable data as possible, we implement quality control with two quality criteria similar to those of [Hua et al. \(2020\)](#), and [Brown and Fischer \(2025\)](#). First, we discard receiver functions with positive phases (using the natural polarity of Sp receiver functions), in the 15-60 km depth range whose amplitudes exceed three times the median amplitude across all receiver functions in the same depth range, as these are inconsistent with a clear Moho phase. Then, we exclude the individual Sp receiver functions with abnormally large amplitudes at mantle depths.

These outliers are identified using the squared L2 norm of the difference between the observed receiver functions amplitude (RF) and 0.05,  $\|RF - 0.05\|_2^2$ . Receiver functions with values exceeding three times the median of this quantity across all receiver functions are removed. This procedure ensures retention of receiver functions with coherent Moho conversions and avoids stations with spurious amplitudes.

A total of 23,804 receiver functions pass the quality control and are used to calculate a CCP stack using Fresnel-zone-based weighting functions (J. Hua et al. 2020). The CCP stack is generated with a horizontal grid spacing of  $0.1^\circ$ . For the joint inversion, the Sp receiver functions are converted back to the time domain using the GLAD-M25 velocity model (Lei et al., 2020).

## 2.3 Bayesian joint inversion

We use a transdimensional hierarchical Bayesian approach developed by [Eilon et al. \(2018\)](#) to constrain a 1-D profile of vertically polarized shear-wave speed ( $V_{sv}$ ) in the crust and upper mantle at each pixel by jointly inverting the period-dependent Rayleigh wave phase velocity curve and the Sp receiver function extracted from CCP stack. This method has been adapted in prior studies to investigate the lithospheric structure in Alaska (Gama et al. 2022) and the southwestern United States (Golos et al. 2024). A brief summary of the method is provided here.

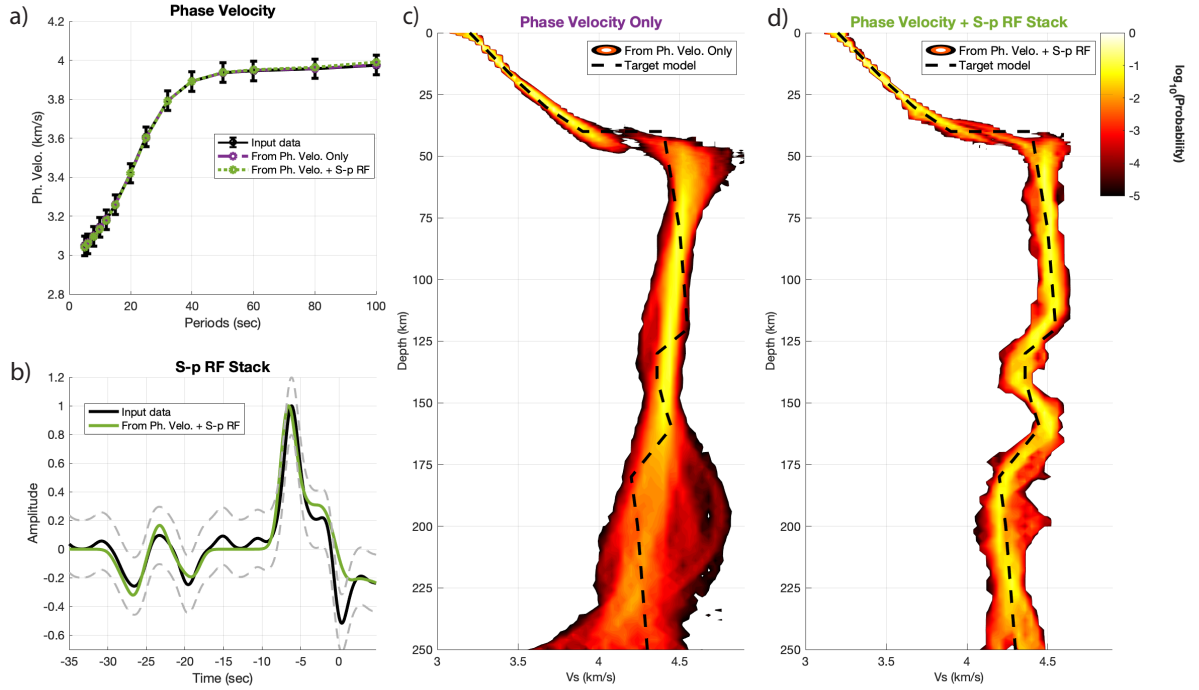
In each iteration, a new velocity model is generated by randomly perturbing the last accepted model; the velocity model is represented using discontinuous cubic B-spline functions. The posterior probability of the new model is evaluated based on the misfit between the observed data and prediction. Rayleigh wave phase velocity predictions are

obtained by integrating the velocity and density perturbations with sensitivity kernels over depth. These kernels are computed using Mineos (Masters et al., 2011) and the equations for the Fréchet derivatives (e.g., Dahlen and Tromp 1998), and they are updated at least every 350 iterations. Density and compressional-wave speed ( $V_p$ ) perturbations are assumed to scale with  $V_{sv}$  perturbations. To calculate predicted  $S_p$  receiver functions, we compute synthetic seismograms using the propagator matrix algorithm (Keith and Crampin 1977) and process them identically to observed data. To assess the misfit of predicted and observed  $S_p$  receiver functions, we follow the approach of Gama et al. (2022): the difference is quantified by comparing (1) the convolution of the observed receiver functions with the synthetic parent component, and (2) the convolution of synthetic receiver functions with the “observed” parent component, which is an assumed Gaussian function centered at 0 s time. This comparison accounts for any differences in frequency content between the observation and prediction. A Metropolis-Hastings acceptance criterion (Hastings 1970) is used to decide whether the proposed model is accepted, based on its posterior probability derived from the calculated misfit and likelihood functions.

During inversion, the crustal  $V_p/V_s$  ratio is fixed at 1.81, while other parameters are allowed to vary from the surface to 250 km following Eilon et al. (2018). These include the shear velocity coefficients, the number and depth of spline knots, the depth and velocity contrast at the Moho, the hyperparameters that describe the relative uncertainties, and the weighting of the two data types. We correct for anelasticity using a reference frequency of 1 Hz. Only physically reasonable models are considered, specifically, those whose seismograms can be computed using the propagator matrix

algorithm. In other words, if a proposed model results in a critical angle violation, where ray paths cannot propagate from the base to the surface of model space, it is rejected, and a new perturbation is generated. The model space is explored using a Monte Carlo Markov Chain (MCMC) approach by running six independent parallel chains for 15,000 iterations. We retain one model every 25 iterations from each chain after the first 8,000 iterations (burn-in period). The final  $V_{sv}$  value at each depth is obtained from the median  $V_{sv}$  across all retained models. The  $V_{sv}$  uncertainties are quantified as the 68th percentile range of  $V_{sv}$  at each depth. Fig. 3 shows results from synthetic tests comparing the Bayesian joint inversion and Bayesian inversion using only Rayleigh wave phase velocity alone. While the phase-velocity-only inversion achieves a good fit to the data, it produces a broad and smooth posterior distribution of  $V_{sv}$  that does not resolve the fine layering of the input model. In contrast, the joint inversion, by integrating two complementary datasets, results in a more tightly constrained posterior distribution that better captures the vertical velocity gradients in the input model.





**Figure 3.3: Figure 3.** A synthetic test comparing the recovery of a vertically layered Vs input structure using Bayesian joint inversion versus inversion from phase velocity only. Black= target (input) model, purple= inverting only phase velocity, green= inverting phase velocity and Sp receiver functions. (a) Rayleigh wave phase velocity. (b) Sp receiver functions. Error bars represent the data uncertainty used for the synthetic test. The sign of the amplitude is flipped to the Ps convention, so that a positive phase corresponds to a velocity increase with depth. (c, d) The posterior probability distributions from inverting only phase velocity and inverting phase velocity and Sp receiver functions jointly respectively.

### 3 Results

#### 3.1 Rayleigh wave phase velocity

The Rayleigh wave phase velocity maps derived from teleseismic earthquakes reveal complex tectonic structures across Fennoscandia (Fig. 2). Two major NNE-WWS-oriented transition zones are evident, particularly at the shortest and longest periods. One transition approximately follows the Caledonian front, with velocities increasing

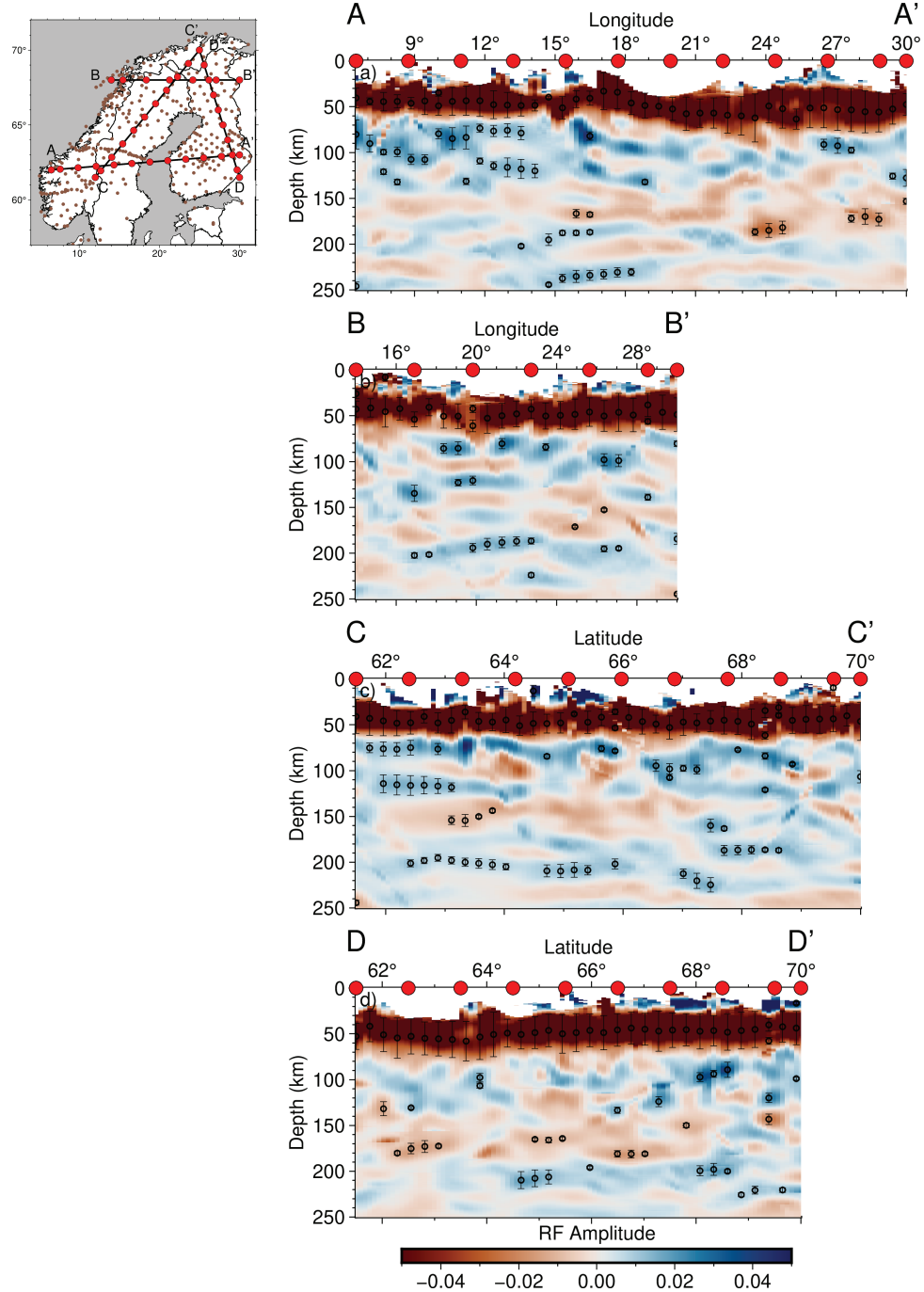
eastward toward the Swedish Svecofennian domain, shown as the red dashed line in Fig. 2b. The other extends along westernmost Norway, as drawn in Fig. 2e, with a velocity decrease toward the east at shorter periods (20-25 s) and a velocity increase toward the east at longer periods (50 s-100s).

The Lofoten Peninsula in northwestern Norway, labeled in Fig. 2b, exhibits the lowest phase velocities at the shortest periods, with a marked velocity increase at longer periods, suggesting a relatively shallow Moho. In contrast, the persistent low velocities observed in southern Finland at short and intermediate periods ( $< 50$  s) are consistent with a deeper Moho observed here (e.g., [Artemieva 2007](#); [Janik et al. 2007](#)). A pronounced low-velocity anomaly between Trondheim and the Norway-Sweden border (bounded by black dashed lines and labeled in Fig. 2c) at 20-32 s agrees with observations by Mauerberger et al. (2022) and Köhler et al. (2012). Fast phase velocity from  $\sim 4.1$  km/s at 40 s to  $\sim 4.25$  km/s at 60 s is imaged in central Sweden and corresponds to the Knaften Arc (black dashed line in Fig. 2d). From 40 s to the longest periods, a low-velocity region appears in southern Norway, bounded sharply across the Oslo Graben by the faster Svecofennian domain. At long periods (80-100 s), high velocities dominate the eastern Svecofennian Paleoproterozoic and Archean cratonic areas, with the highest velocity ( $\sim 4.25$  km/s) observed in the Svecofennian Paleoproterozoic craton at 100 s.

The measured Rayleigh wave phase velocities show good agreement with previous studies that also used the regional temporary seismic arrays. In Fig. S2, we compare the velocities obtained in this study with average Rayleigh wave phase velocity curves from southern Finland using SVEKALAPKO by Pedersen et al. (2013), north Finland using LAPNET by Bruneton et al. (2004), and southern Norway using MAGNUS, the

seismological broadband experiment deployed in southern Norway (Weidle et al. 2010), by Maupin (2011). In all three locations we also compare to regionally averaged phase velocities calculated from the maps of Mauerberger et al. (2022). Overall, the results are consistent. In north Finland, both our measurements and those of Mauerberger et al. (2022) show slightly lower velocities ( $\sim 0.05$  km/s) than Pedersen et al. (2013) at longer periods, yet all remain within one standard deviation. In the part of southern Finland covered by SVEKALAPKO, our median phase velocity at 80 s is marginally slower ( $\sim 0.05$  km/s) than that of Bruneton et al. (2004) and Mauerberger et al. (2022), likely because of the larger uncertainty along northeastern margin of our study region, due to path coverage. Nevertheless, at 100 s the three studies agree well. In southern Norway, all studies consistently reveal a low-velocity structure.

Strong azimuthal anisotropy of Rayleigh wave phase velocity at longer periods ( $> 35$  s) has been reported in the study area (Mauerberger et al. 2020). We consider our Rayleigh wave phase velocity maps to represent isotropic velocity because they are an average of values determined from earthquakes covering a range of back-azimuths. To assess any potential bias introduced by azimuthal anisotropy in our measurements, we calculate azimuth-weighted Rayleigh wave phase velocities by binning earthquake back-azimuths in  $15^\circ$  intervals and computing a weighted median that downweights overrepresented directions. The resulting azimuth-weighted averages (Fig. S3) closely resemble the results from the standard average phase velocities (Fig. 2), with mean difference less than 0.01 km/s. We conclude that any azimuthal anisotropy effects are effectively averaged out by the azimuthal coverage of events. Therefore, we use the direct average phase velocity maps as input for joint inversion.



**Figure 3.4: Figure 4.** Sp receiver function CCP stack amplitudes along four profiles. Sp receiver functions are given with their true polarities, where red colors (negative amplitudes) indicate positive velocity gradients and blue values (positive amplitudes) indicate negative velocity gradients. Black circles with error bars represent the most robust phases with amplitudes exceeding twice their standard deviations and a phase depth extent of more than 5 km. The top-left inset map shows the profile locations, with dots representing all stations used to generate receiver functions.

## 3.2 Sp receiver function stacks

Four CCP stacking profiles across Fennoscandia are shown in Fig 4. A prominent and continuous negative phase in true Sp receiver functions, indicating a velocity increase with depth, from 15 to 69 km depth corresponds to the Moho. In the upper mantle, positive velocity gradients (PVGs) between 140-190 km are observed sporadically across the study region. Negative velocity gradients (NVGs) occur at ~100 km beneath the southeastern study area, while deeper NVGs (~200 km) are most prominent beneath northern Finland and central Sweden.

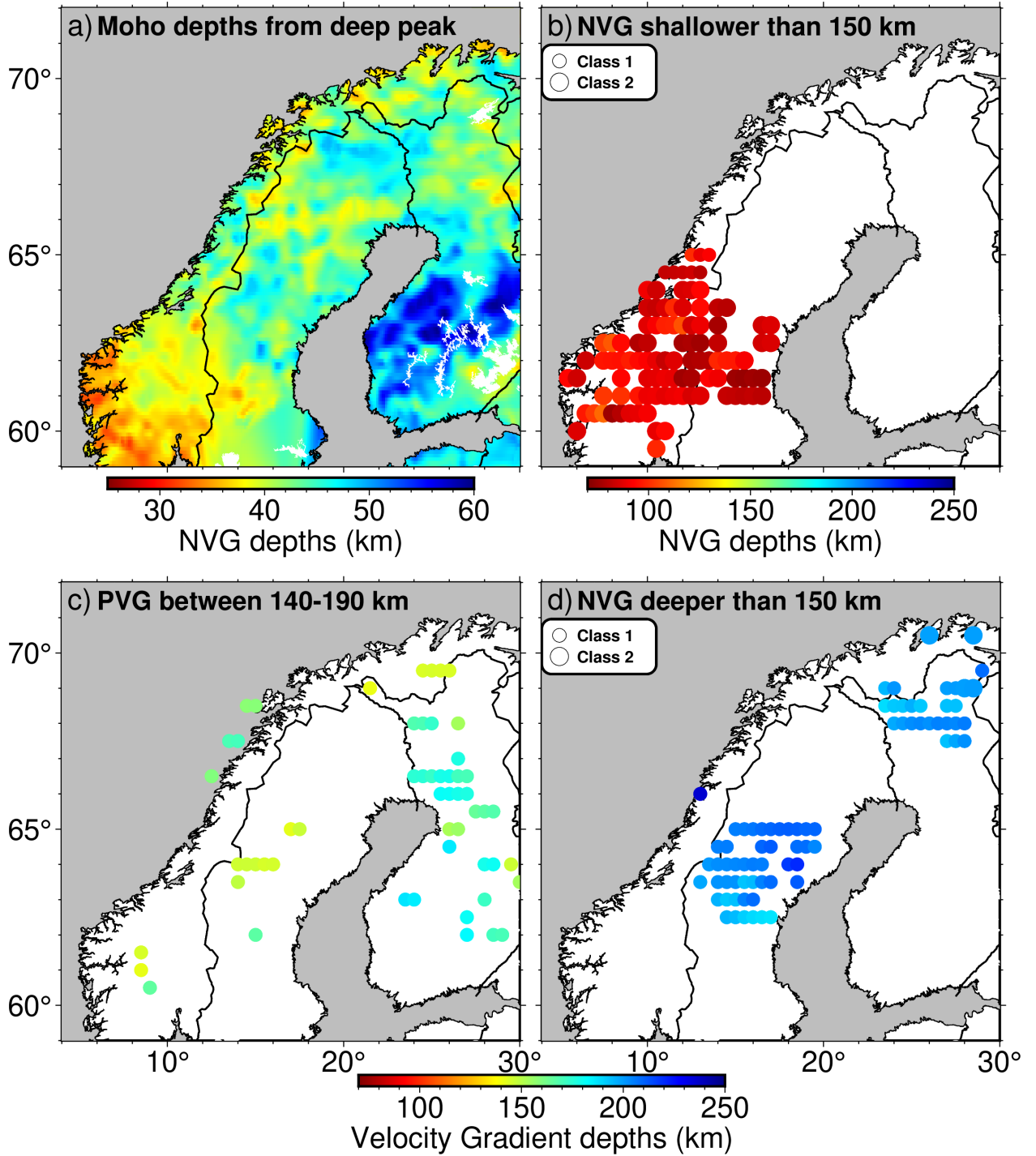
### 3.2.1 Moho

Profiles through the CCP stacking reveal crustal thickening from west to east (e.g. Fig. 4a), with the deepest Moho beneath southern Finland. This pattern is consistent with previous controlled-source and passive seismic experiments (e.g., [Tiira et al. 2020](#); [Alinaghi et al. 2003](#)), and reflects complex tectonic evolution through the accretion of island arcs and microcontinental blocks during Archean and Proterozoic (Elena Kozlovskaya et al. 2008).

The continuous PVG shallower than 70 km enables Moho picking across the entire Fennoscandian region. We pick Moho depth from the maximum amplitude in the depth range 10-70 km in each CCP stack column. However, we found some columns contain more than one peak in this depth range (e.g., Fig. S4a). One possible cause is strong azimuthal anisotropy near the Moho, which produces periodic variations with back-azimuth in the arrival time and amplitude of converted phases. Consequently, rays from different back azimuths can generate multiple peaks in CCP stacks if anisotropy is

significant (Schulte-Pelkum and Mahan 2014). In order to determine if the CCP stack can represent the real seismic velocity gradient rather than azimuthal anisotropy, we perform forward modeling using Telewavesim (Audet et al. 2019), introducing a 15-km-thick anisotropic layer located either above or beneath the Moho with 5% peak-to-peak anisotropy and a fast axis orientation of  $10^\circ$  East of North, consistent with or stronger than the estimates at 50 km depth from Grund and Ritter (2020) and Zhu et al. (2015) in northern Fennoscandia. The modeled S-p receiver functions show a clear oscillation with back-azimuth, but their stacks do not reproduce the double Moho peaks in either case (Fig. S4c,d). Furthermore, our single-station receiver functions sorted by back-azimuth bins show no clear periodic variation in Moho depth and amplitude (Fig. S4b) as would be expected if azimuthal anisotropy were the cause. We therefore conclude that the observed double peaks likely reflect real vertical seismic velocity gradients rather than azimuthal anisotropy.

Double Moho phases have also been reported in earlier receiver function studies of Fennoscandia. For example, Silvennoinen et al. (2014) identified two or three distinct P-s peaks in the northern Fennoscandian shield. Using S-p receiver functions, Makushkina et al. (2025) detected two converters associated with the Moho in Fennoscandia and interpreted them as crustal stacking and/or alteration. Compared to Makushkina et al. (2025), our double peaks are less coherent and more spatially variable. For example, adjacent CCP columns separated by  $0.1^\circ$  often show different Moho shapes in Fig. S4a. This discrepancy may result from our use of a lower-frequency filter band (4-100 s) compared with the 2-100 s band used by Makushkina et al. (2025).



**Figure 3.5: Figure 5.** Discontinuity depth maps from the CCP stack. (a) Moho depth extracted from the deeper positive velocity gradient at depths shallower than 70 km. (b) Negative velocity gradients (NVGs) deeper than the Moho but shallower than 150 km. Symbol size indicates coherence class. (c) Positive velocity gradients (PVGs) between 140-190 km. (d) Same as (b) but for NVGs deeper than 150 km.

In Fig. S5 we compare Moho depths extracted from the strongest positive Sp receiver function peaks and, when two peaks are present, from the deeper peak. In general the differences between them are minor. However, in southern Finland using the deeper peak instead of the stronger peak to estimate Moho depth yields slightly greater Moho depths. Given that the thick crust ( $>50$  km) in southern Finland is well constrained by active source studies (e.g., Luosto et al. 1984), we prefer the deeper peak as representative of the seismic crustal-mantle boundary, a choice we apply to all locations with a double Moho phases. Our final Moho depth map (Fig. 5a) agrees well with previous Moho depth compilations based on seismic reflection, refraction and receiver functions and other available individual studies (EuCRUST-07, [Tesauro et al. 2008](#); Fig. S5c), but it provides a greater level of detail, which is expected since EuCRUST-07 was interpolated from sparser data sets. For instance, we identify a deep Moho ( $\sim 48$  km) in the northeastern-most part of our study region beneath Kola Craton (bounded by red dashed circle in Fig. S5b), consistent with [Silvennoinen et al. \(2014\)](#), who combined controlled-source and receiver function data, but this feature is absent in EuCRUST-07.

Along the northern Sweden-Finland border, the average Moho depth is  $\sim 47$  km, thicker than surrounding crust, consistent with strong Proterozoic reworking ([Hanna Silvennoinen et al. 2014](#)). In the southern Scandes mountains, the relatively thin crust ( $< 40$  km) implies the absence of a thick crustal root to isostatically support the elevated topography of the mountain range; support from the deeper mantle is required ([England and Ebbing 2012](#); [Stratford et al. 2009](#)). We find the shallowest continental Moho in the southwestern region of our study area (i.e., southern Norway), with a progressive



deepening toward the northeast inland, consistent with the findings of Kolstrup and Maupin (2013).

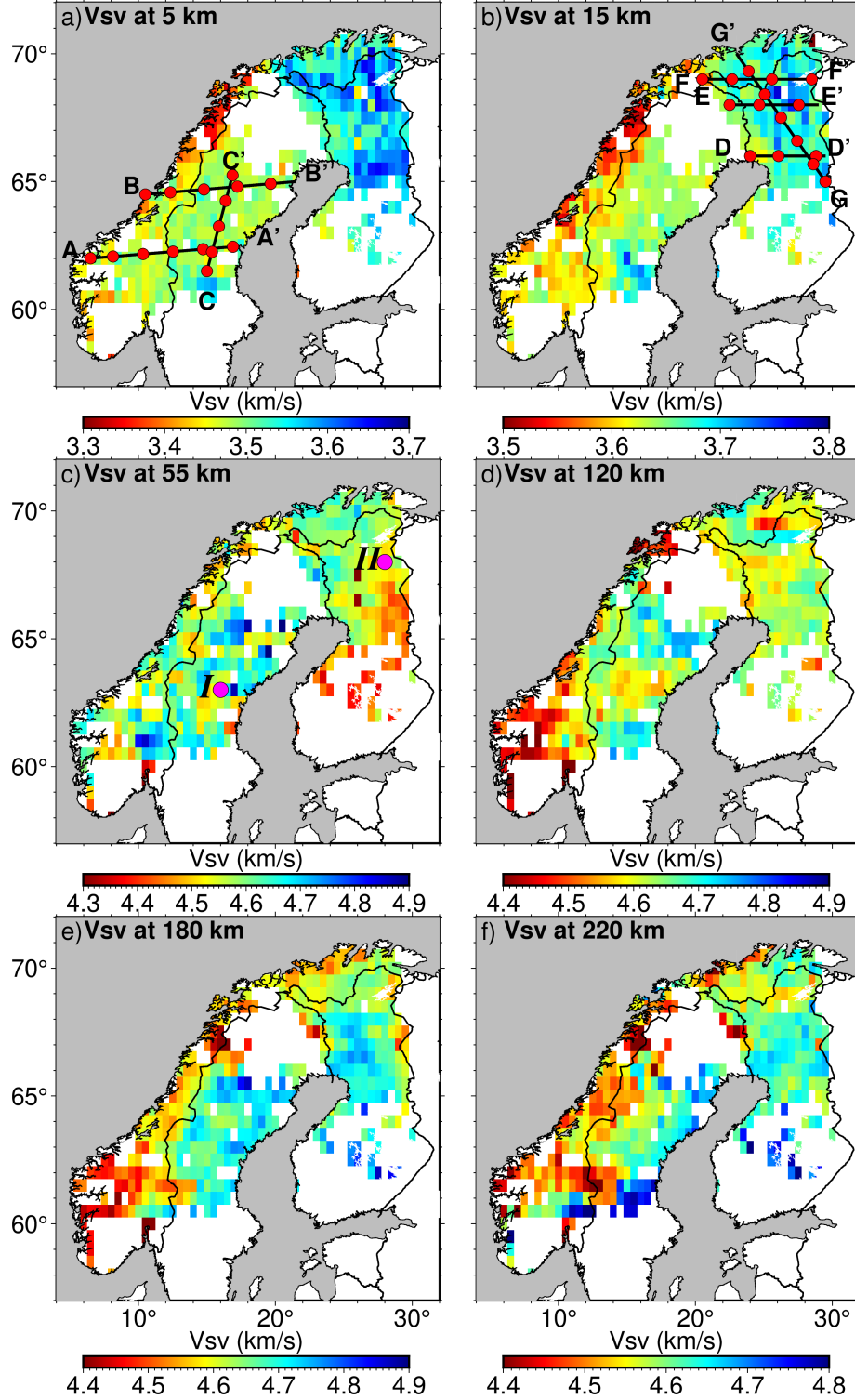
### **3.2.2 Uppermost Mantle**

The CCP stacks reveal multiple PVGs and NVGs within the uppermost mantle of Fennoscandia, indicating complex and layered lithospheric structures. When interpreting the stacks, the most robust PVGs and NVGs (phase amplitude exceeds twice its standard deviation and the phase spans  $> 5$  km vertically) were used as a guide. In the southwestern part of the study area, from west to east (profile AA' in Fig. 4a), a NVG dips eastward from  $\sim 70$  km to  $\sim 120$  km, extending to about  $15^\circ\text{E}$  in longitude. This pattern may correspond to a progressive deepening of the lithosphere-asthenosphere boundary (LAB) from west to east, consistent with previous studies (V. Maupin et al. 2013; Wawerzinek et al. 2013). Beneath the Gulf of Bothnia, in the longitude range  $15^\circ$ - $18^\circ\text{E}$ , two deeper NVGs are imaged at approximately 190 km and 240 km, respectively. Deep NVGs beneath southern Finland are less continuous and less well defined. However, laterally extensive NVGs at  $\sim 200$ -km depth are prominent along profiles BB', CC', and DD'.

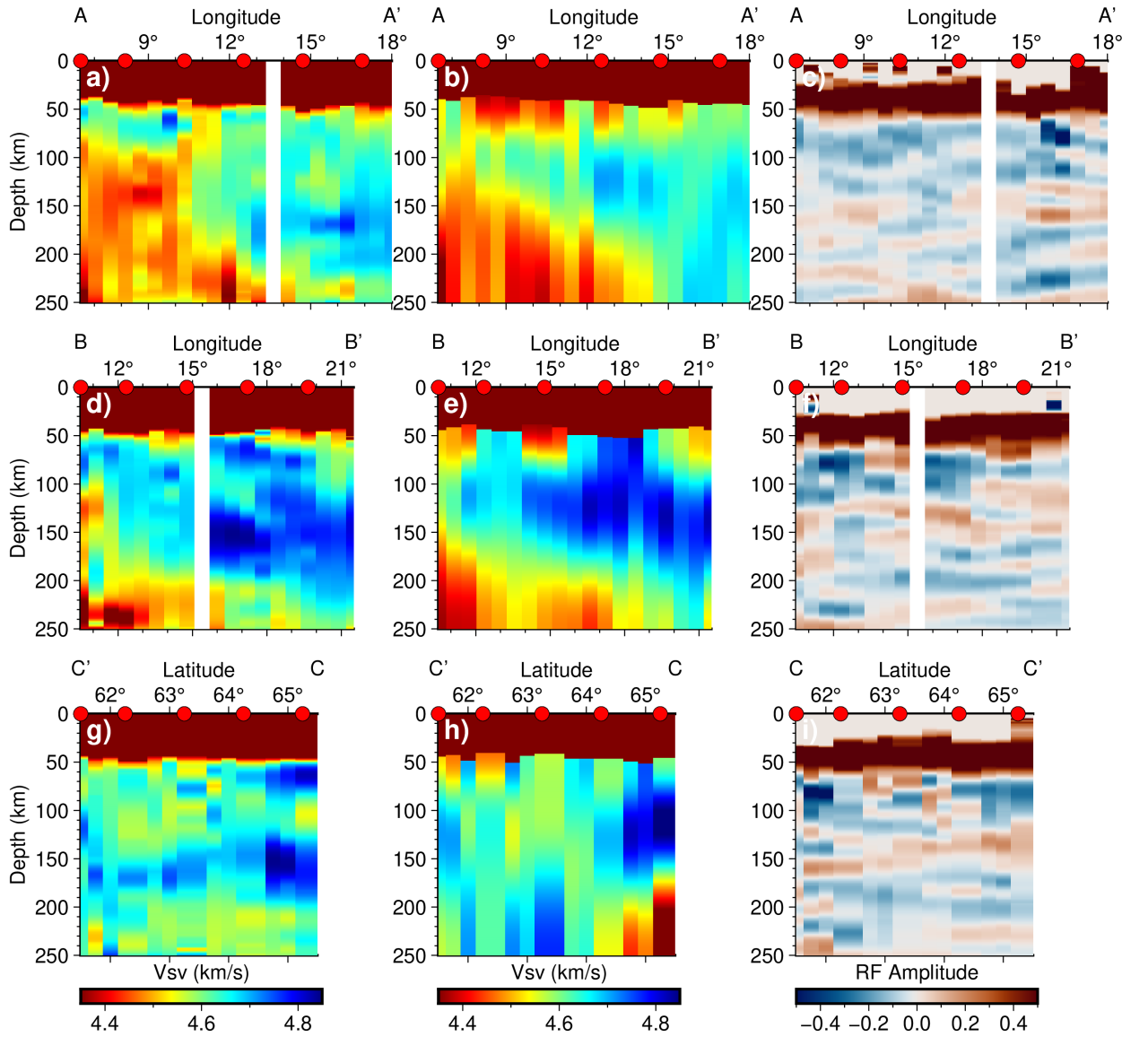
Evidence of PVGs in the depth range 150-200 km is clear beneath southern Finland in the AA' profile (Fig. 4a) and across much of Finland in the DD' profile (Fig. 4d). Similar PVGs at  $\sim 160$ -km depth have been reported in prior studies. Receiver function analyses identified PVGs at the base of a low-velocity zone at 140 km in northern Finland (Goev et al. 2021) and at 110-130 km in southern Finland (Lev Vinnik et al. 2016). Yliniemi et al. (2004) used local events and found a P wave velocity increase at

~130 km in southern Finland. In the Sweden Svecofennian domain, both [Kind et al. \(2013\)](#) and [Olsson et al. \(2007\)](#) observed PVGs at ~150 km to ~160 km using S-p receiver functions. Active-source experiments in the same region also detected a PVG at 160 km, interpreted as the base of a low-velocity zone associated with partial melting ([Perchuć and Thybo 1996](#); [H. Thybo and Perchuć 1997](#)).

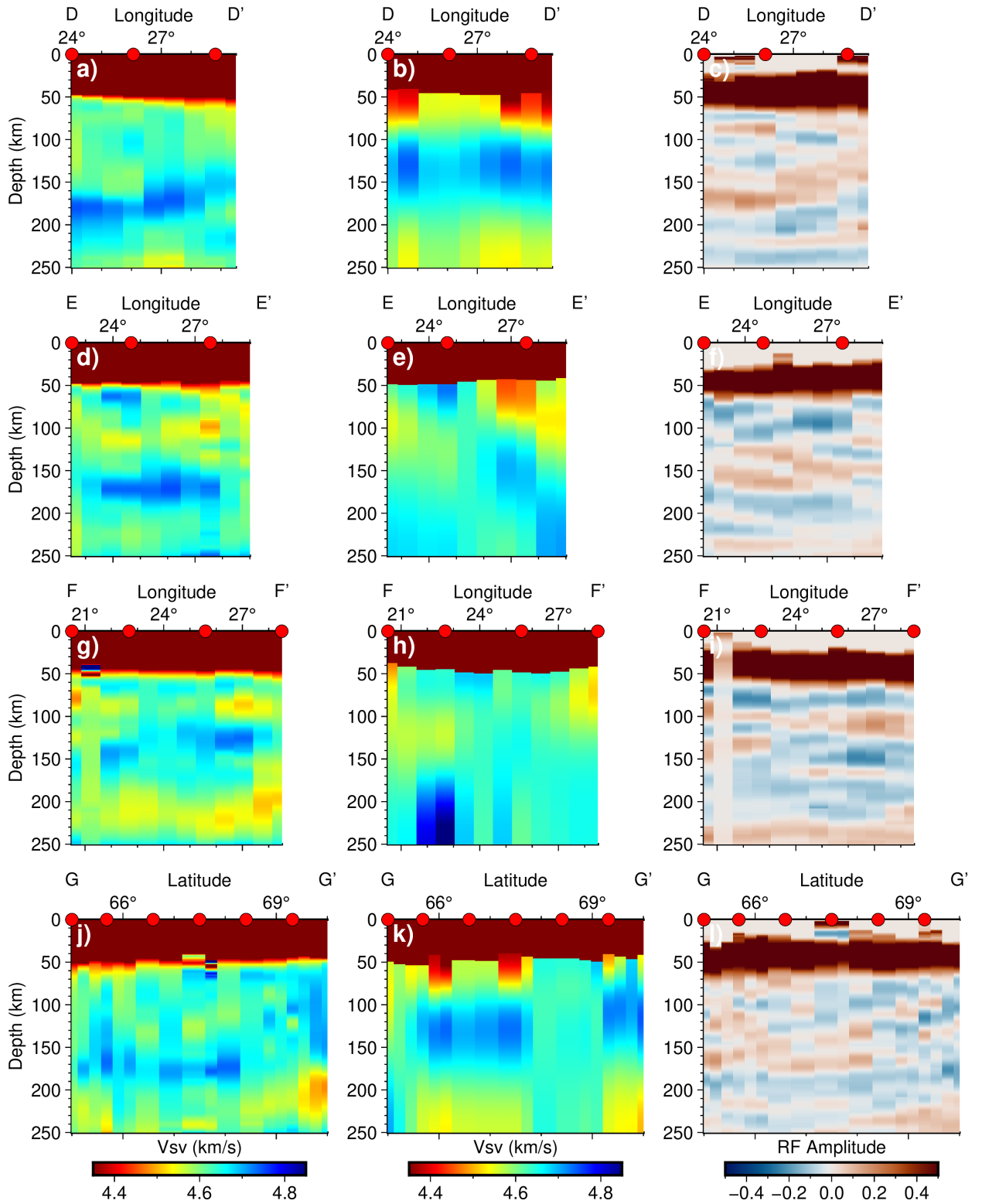
In order to further evaluate the significance of the mantle seismic velocity gradients observed in CCP stacks, we classified them based on their spatial consistency. For all significant NVG and PVG phases (amplitude > than 2 standard deviations and vertical extent > 5 km) between 70 km and 250 km depth, we evaluated each  $0.5^\circ \times 0.5^\circ$  grid cell and defined two confidence classes. Class 1 means that within a  $1.5^\circ$  radius cap of the grid cell, more than half of the CCP stacks columns display the same phase. Class 2 enlarges the searching radius to  $3^\circ$ , indicating higher regional coherence. The resulting classification (Fig. 5) shows that shallower (< 150 km) NVGs are concentrated beneath southern Norway and Sweden, whereas deeper (~200 km) NVGs occur mainly beneath central Sweden and northernmost Finland. Laterally coherent upper-mantle NVGs are largely absent from northern Norway and Sweden and from southern Finland. In contrast, PVGs near 160-km depth are weaker and less continuous. To visualize these weaker features, we relax the criteria used to identify a PVG near 160 km to a search radius of  $0.25^\circ$  and require that only 1/3 of CCP stacking columns show a PVG between 140 and 190 km. The ~160-km-depth PVGs identified in this way are scattered but present across much of Fennoscandia (Fig. 5c), suggesting a regionally persistent but laterally variable mid-lithospheric velocity increase.



**Figure 3.6: Figure 6.** Map views of horizontal  $V_{sv}$  slices extracted from the joint inversion models at depths of (a) 5 km, (b) 15 km, (c) 55 km, (d) 120 km, (e) 180 km, and (f) 220 km. Thick lines in (a) and (b) indicate the locations of cross-section profiles shown in Figs. 6 and 7, respectively. Magenta dots in (c) labeled with I and II indicate the pixel locations of 1D joint inversion examples shown in Figs. 10 and 11 respectively.



*Figure 3.7: **Figure 7.** Vsv cross-sections. Left column: joint-inversion models; Middle column: Vsv model from inversion using phase velocity only; Right column: Sp receiver function CCP stack amplitudes. The sign of the amplitude is flipped to the Ps convention, so that a positive phase corresponds to a velocity increase with depth. Top: profile AA'; middle: profile BB', bottom: profile CC', where profile locations are shown in Fig. 6a.*



*Figure 3.8: Figure 8.* Same as Fig. 7 but for profiles DD', EE', FF', and GG', where profile locations are shown in Fig. 6b.

### 3.3 Regional shear-wave velocity patterns in Fennoscandia Shield

Because the CCP stacks and Rayleigh wave phase velocity maps have different spatial resolutions, we resample the receiver function data at each grid cell of the phase velocity map. For each grid cell, we use the median of all bootstrapped individual receiver functions, with their uncertainties, located within  $0.2^\circ$  of the pixel center and quantified the 95% confidence interval from bootstrapping as the receiver function data uncertainty for joint inversion. The joint inversion is performed for 503 points on a uniform grid of  $0.5^\circ \times 0.5^\circ$ . For each pixel, we consider the median  $V_{sv}$  from the posterior distribution to be the representative model, with the standard deviation as its uncertainty. To exclude models exhibiting extreme mantle structure, we compute the depth integral of synthetic receiver-function amplitudes exceeding 0.3 between 60 and 250 km depth. Pixels with anomalously large integrals ( $>10$  km) are removed, resulting in 495 valid grid points. Six horizontal  $V_{sv}$  slices are shown in Fig. 6, with their uncertainty maps at Fig. S6, together with seven vertical  $V_{sv}$  cross-sections (Figs. 7 and 8).

The  $V_{sv}$  pattern at shallow depths closely matches that of Mauerberger et al. (2022), as expected since both studies use the same ambient noise Rayleigh wave phase velocity data sets. Very low velocities ( $<3.4$  km/s) are observed in Lofoten peninsula and Nordland province on the northwestern coast of Norway within the top 15 km. A pronounced low velocity at 55 km beneath southern Finland corresponds to thick crust (Fig. 6c). A pronounced high-velocity zone ( $>4.7$  km/s) extends to the base of the model space (250 km) beneath the central southern Finland (Fig. 6f), in agreement with Sandobal et al. (2004). In central and southern Finland, the seismic boundary between

Archean Karelian craton and the Svecofennian domain ( $\sim 65^\circ\text{N}$ ) is not clear in the uppermost mantle (Fig. 6), similar to previous findings (e.g., Sandobal et al., 2004; Kozlovskaya et al., 2008; Bruneton et al., 2004b). The persistently fast velocities down to 250 km and absence of a clear velocity drop suggest that the LAB lies deeper than 250 km beneath much of Fennoscandian shield (Sandobal et al., 2004).

Profiles through the Vsv model further illuminate upper mantle structure. In southern Norway, a high velocity layer ( $\sim 4.6$  km/s) extends down to  $\sim 80$ -100 km, and deepens sharply eastward across the Oslo Graben at  $12^\circ\text{E}$  longitude (Profile AA' in Fig. 7a), consistent with previous studies (e.g., [Köhler et al. 2012](#); [Kolstrup et al. 2015](#); [Wawerzinek et al. 2013](#)). This transition is associated with the southwestern edge of the thick lithosphere of the Fennoscandian/Baltic Shield (Medhus et al. 2012), and the high velocity layer was interpreted as the lithospheric mantle. The faster and thicker lithosphere in the Shield reflects the temperature (Medhus et al. 2012) and/or composition (Gradmann et al. 2013) contrasts between Proterozoic and Phanerozoic terranes. Such a sharp lateral transition in the depth of the LAB is less clear further north (profile BB' in Fig. 7d), where the lithosphere remains relatively thick ( $\sim 200$  km) until close to the coast beneath the Central Scandes. Within the Knaften Arc, the eastern end of BB', the lithosphere is consistently fast ( $\sim 4.8$  km/s) to 200 km depth—coinciding with the Knaften anomaly of Mauerberger et al. (2022) and corresponding to the higher phase velocities measured there. In the eastern part of profile AA', the lithosphere in Svecofennian terrain shows a layering structure, with a positive velocity gradient at  $\sim 150$  km, and a velocity decrease below  $\sim 200$  km. The fastest velocity within lithosphere is  $\sim 4.8$  km/s. This layered pattern persists throughout the Swedish Svecofennian domain

(profile CC' in Fig. 7g) and is required by the receiver function data (Figs. 7c and 7i).

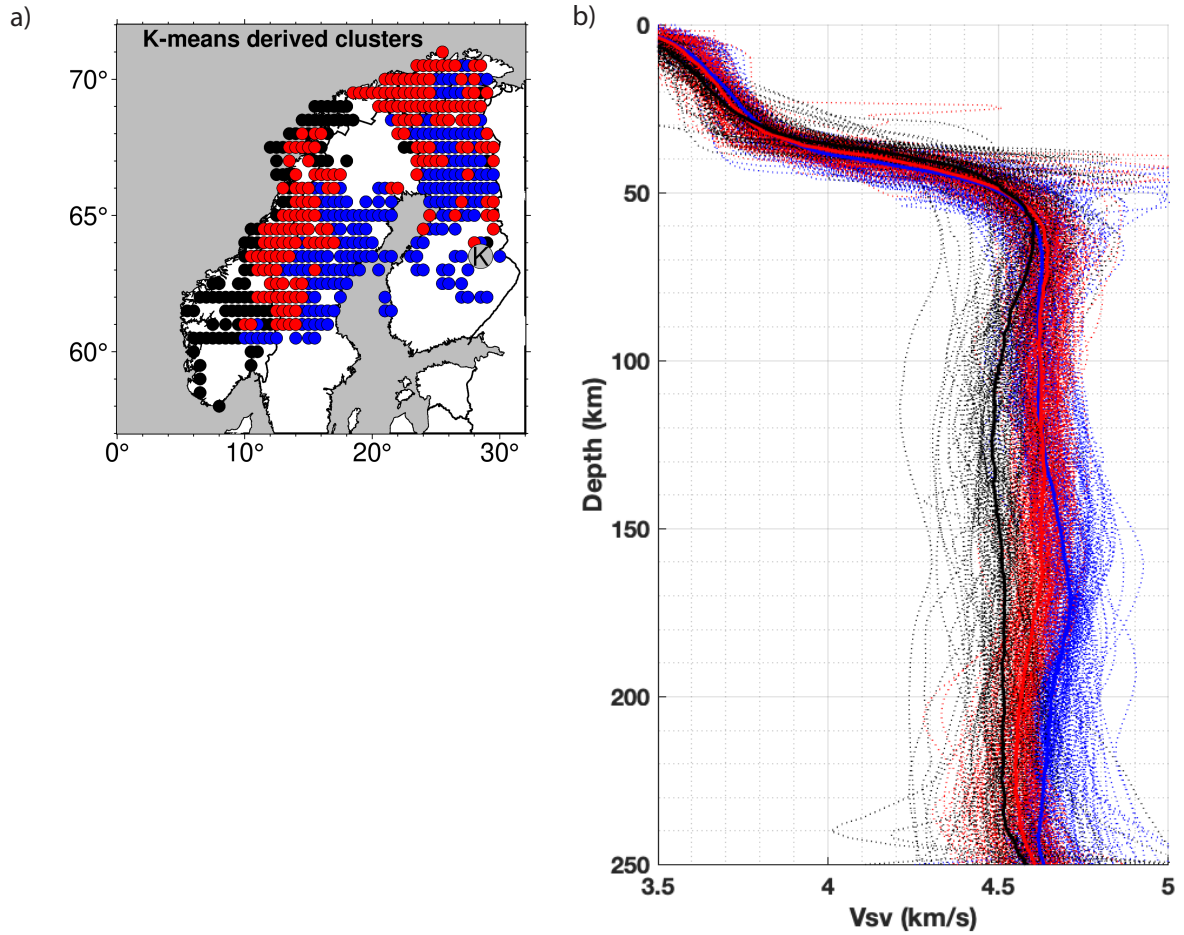
The layering structure disappears when only phase velocities are inverted (Figs. 7b and 7h), as we will discuss in section 4.1.

A similar layering in the lithosphere is observed across most of Karelia Archean craton in northern Finland (Profiles DD', EE', FF' in Figs. 8a, 8d, and 8g). These profiles exhibit a positive velocity gradient between ~140 km and 190 km, a velocity peak at ~4.7 km/s and decreasing velocity at deeper depth. Such a velocity gradient, dividing the lithosphere into a shallow low-velocity-zone above 150 km and deeper, faster layers, also appears in P-wave tomography over the large region covered by LAPNET in northern Finland (H. Silvennoinen et al. 2015). From south to north, a low velocity zone is observed deeper than 200 km in the Karelia craton (north of 68°N in GG' profile in Fig. 8j). Comparable low-velocity zones in the northern Karelian and Kola Cratons are also found in Pedersen et al. (2013), Vinnik et al. (2016), and Mauerberger et al. (2022), as discussed further in section 4.2.

To investigate large-scale trends in the seismic properties of the lithospheric mantle, we apply K-means clustering to the 1-D  $V_{sv}$  models between 70 km and 250 km, grouping them into three clusters based on absolute velocity (Fig. 9). The black cluster is characterized by low median velocities (~4.5 km/s) throughout most of mantle depth range and is distributed mainly across southern Norway and along the Norwegian coast. It is broadly consistent with one of the transitions identified in the phase velocity maps (Fig. 2e), with lower velocity on the western side. The slow velocities in southern Norway were attributed by Hejrani et al. (2017) to a thinner lithosphere, higher temperature, and reduced depletion and/or fluid content in the shallow asthenosphere,



while the coastal low velocities mark the eastern limit of repeated rifting and basin formation since the Devonian.



**Figure 3.9: Figure 9.** (a) Geographical distribution of the three clusters obtained by K-means classification of 1-D Vsv models between 70 km and 250 km. The gray circle labeled by 'K' indicate the Kaavi-Kuopio Kimberlite Province described in Section 4.3.1. (b) Solid lines: median velocities of the three clusters (color coded); dotted lines: all individual velocity models within each cluster.

The red and blue clusters both exhibit high velocity ( $\sim 4.6$  km/s) shallower than 150 km, with the blue cluster characterized by faster velocity deeper than 150 km ( $\sim 0.1$  km/s higher than the red cluster). Geographically, the blue cluster includes the Paleoproterozoic Svecofennian Shield in Sweden and Finland as well as the southern

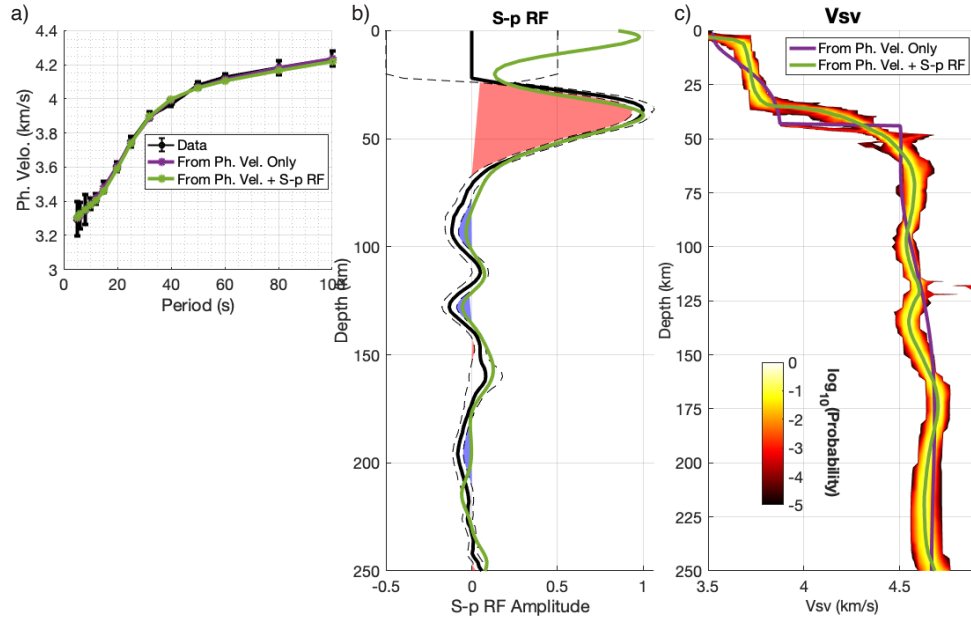
areas of the Archean Karelian craton. For these reasons, we refer to the blue cluster as the “seismic craton interior”. This regime roughly overlaps with the exceptionally high-velocity anomalies within the central Svecofennian domain identified in a P-wave tomography study (Bulut et al. 2022), which are interpreted as the modern-day Fennoscandian cratonic core. The red cluster contains much of the Caledonides as well as the northernmost Archean areas of Sweden and Finland. We refer to the red cluster as the “seismic craton edge”.

## 4 Discussion

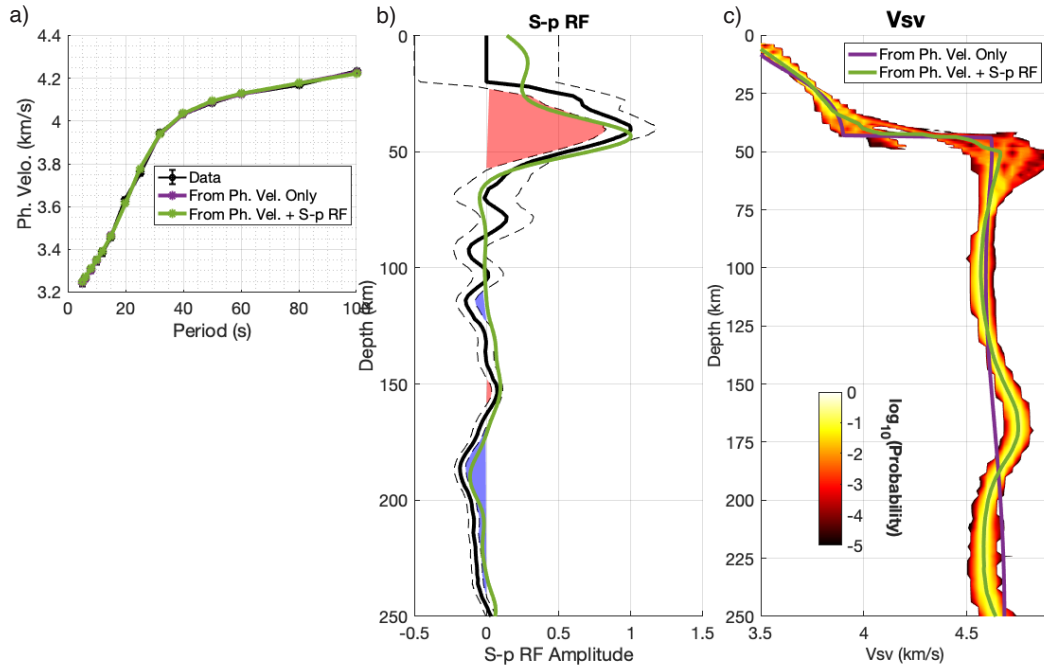
### 4.1 Comparison with inversion from phase velocity only

The layered lithosphere structure revealed in this study is constrained jointly by Rayleigh wave phase velocities, which provide broad depth sensitivity to absolute  $V_{sv}$ , and  $S_p$  receiver functions, which resolve abrupt  $V_{sv}$  changes with depth. To highlight the additional constraints provided by the receiver functions, we compare the joint inversion results with models obtained from inverting only the phase velocities.

To facilitate a meaningful comparison, we perform a 15-iteration linearized inversion using the same Rayleigh wave phase velocity data used in the joint inversion to solve for profiles of  $V_{sv}$  from the surface to 250-km depth. Following step 1 in [Huang et al. \(2025\)](#), we solve for perturbations to a reference  $V_{sv}$  model that is updated at each iteration, and we impose a smoothness constraint that is relaxed across the Moho. The Moho depths are fixed to the values obtained from the  $S_p$  receiver function CCP stacking results (Fig. 5a). Instead of the transdimensional Bayesian framework used in the joint inversion, we use an iterative linearized inversion, which we found was less sensitive to measurement noise when only a single data set is being inverted. All other parameter settings, including the factors that enforce scaling of  $V_p$  and density with  $V_s$  as well as the attenuation factors  $Q_k$  and  $Q_\mu$ , are kept identical to those used in joint inversion. Slices of the resulting  $V_{sv}$  model obtained from phase velocity only are presented in Fig. S7. Several  $V_{sv}$  cross-sections are compared to the joint inversion results in Figs. 7 and 8, and comparisons of 1-D profiles are shown in Figs. 10 and 11.



*Figure 3.10: **Figure 10.** Example of 1D joint inversion at the pixel location indicated by the as the magenta dot labeled as I in Fig. 6c. (a) Rayleigh wave phase velocity. Black = data; Purple = prediction from model obtained using only phase velocity; Green = prediction from model obtained with joint inversion. (b) S-p receiver functions, with polarity flipped to match P-s convention. Error bars show data uncertainty used for inversion, (c) Comparison of Vsv profiles. Posterior probabilities for the joint inversion are shown by the colormap.*



**Figure 11.** As in Fig.10, but for the pixel location indicated by the magenta dot labeled as II in Fig. 6c.

At first order, both approaches recover similar large-scale patterns of heterogeneity, including relatively slow velocities in northern Norway in the crust and faster mantle velocities beneath the Svecofennian domain and the Karelian craton (Figs. 6 and S7). However, important differences emerge in the depth distribution of the highest upper-mantle velocities. In the joint inversion models, the highest velocities ( $>4.7$  km/s) occur predominantly between  $\sim 140$  km and  $\sim 210$  km depth, whereas in the phase-velocity-only models, the highest velocities are concentrated in the uppermost mantle shallower than  $\sim 150$  km. This contrast is evident in profile CC' in Fig. 7 and the profiles in Fig. 8. The depth of the high velocity layer shown in the joint inversion models corresponds closely with positive velocity gradients observed in the Sp receiver function CCP stacks, which are shown in the right columns in Figs. 7 and 8.

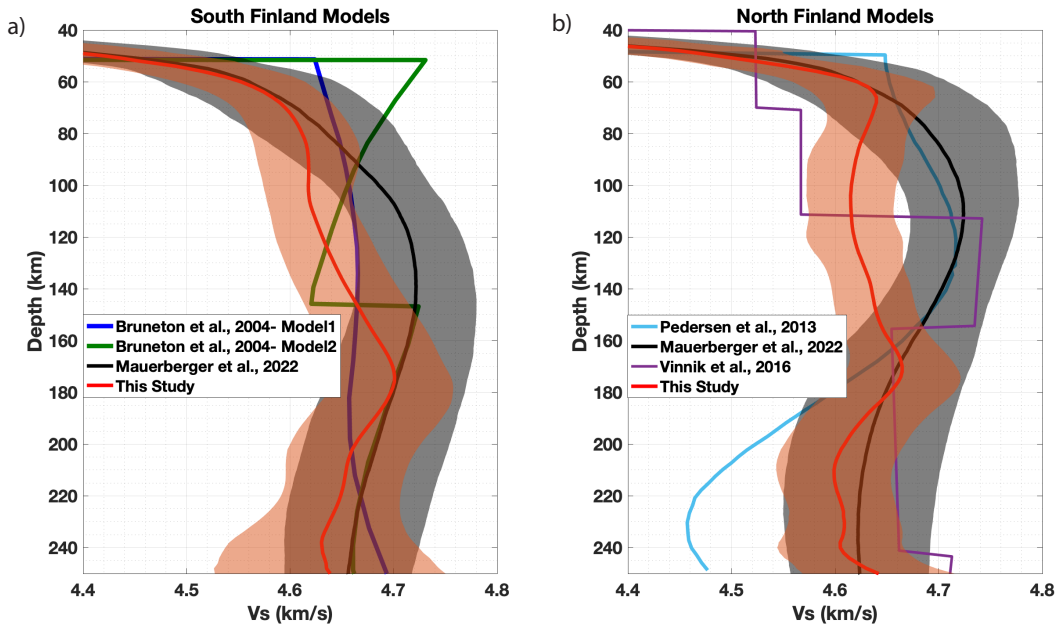
The 1-D velocity examples further clarify the origin of these differences. As shown in Figs. 10 and 11, both inversions fit the Rayleigh wave phase velocity data well. However, the phase-velocity-only inversions produce a relatively smooth and simple velocity structure, with velocity gradually increasing with depth. In contrast, at many locations in the Svecofennian shield and Karelian craton, the joint inversion requires the presence of a lower velocity layer above ~150 km in order to fit the PVG at this depth range. As a result, in the Archean and Proterozoic domains of the study area, the highest-velocity layer is shifted to greater depths. We also find some differences in the mantle immediately below the Moho, but we suspect they may primarily reflect that the joint inversion and phase-velocity-only inversion treat the Moho somewhat differently (Fig. 10). The joint inversion allows a more flexible and gradual Moho structure by incorporating the width of the Moho phase in  $S_p$  receiver functions, whereas the phase-velocity-only inversion prescribes the Moho as a sharp discontinuity at a fixed depth.

## **4.2 Comparison with previous layered seismic lithosphere models**

Previous studies in Fennoscandia have suggested the possibility of layered lithosphere, although they often cannot robustly distinguish between multi-layered and single-layered models. For example, [Bruneton et al. \(2004\)](#) showed that surface-wave-derived velocity models beneath southern Finland (Model 1 in Fig. 12a), which were characterized by nearly constant velocity with depth in the upper mantle, were inconsistent with xenolith-based predictions for harzburgitic or lherzolitic compositions, which required significantly higher velocities in the uppermost mantle that decreased with increasing depth. To reconcile this discrepancy, [Bruneton et al. \(2004\)](#) proposed a stratified model with an interface at ~150 km depth (Model 2 in Fig. 12a), interpreting

the shallower slow-Vs layer as the product of mantle metasomatism or the accumulation of dense ultramafic material. However, they noted that Rayleigh wave phase velocities alone could not discriminate between these two models.

Our models for southern Finland support the notion of layered lithosphere with a boundary near 150 km depth. The deep layer in our model closely resembles the deep layer of the stratified model proposed by Bruneton et al. (2004), both in the absolute value and in gradient, and is also consistent with the model of Mauerberger et al. (2022), shown as the black line with gray shading in Fig. 12a. Both our model and the stratified model of Bruneton et al. (2004) require a relatively slow lithospheric mantle above 150 km. However, the velocity gradients differ: our model exhibits a gradual velocity increase from the Moho to 150 km, whereas Bruneton et al. (2004) required a velocity decrease, with higher velocities than ours between ~50 and 100 km depth. In contrast, models of Mauerberger et al. (2022) without explicit layering show faster velocities in the shallow mantle above 150 km, particularly near 120 km. Importantly, the layered structure in our study is uniquely required by the positive velocity gradients observed in Sp receiver functions, ruling out the models with only weak or continuous velocity variations.



*Figure 3.12: **Figure 12.** Comparisons of velocity models for (a) southern Finland region, covered by the SVEKALAPKO network, and (b) northern Finland region, covered by the LAPNET network. Black and red lines show the mean  $V_{sv}$  models from Mauerberger et al. (2022) and this study, respectively. Shaded regions indicate the corresponding minimum and maximum velocity bounds.*

In northern Finland (Fig. 12b), our models broadly follow the same layered structure as in southern Finland, with minor variations in gradients just below the Moho. The most pronounced difference between northern and southern Finland lies below 150 km depth, where we find velocities in northern Finland that are approximately 0.2 km/s lower than those in southern Finland, consistent with previous studies. In our model, the  $V_{sv}$  of the deeper lithospheric layer in northern Finland lies between the low velocities of Pedersen et al. (2013) and the higher wave speeds of Vinnik et al. (2016), and it overlaps with the values from Mauerberger et al. (2022).

Several hypotheses have been proposed to explain the origin of the deep low-velocity layer in northern Finland. Pedersen et al. (2013) interpreted this layer as



refertilized lower lithosphere with a more fertile composition due to asthenosphere upwelling or small-scale erosion at the craton edge. Incorporating observations of changes in the fast direction of anisotropy across this boundary (Lev Vinnik et al. 2014), Vinnik et al. (2016) favored that this layer belongs to the asthenosphere, given that the top boundary of the low-velocity layer (160 km) coincided with the intersection of the geotherm and the wet peridotite solidus. However, xenolith samples from these depths do not show evidence for volatile-bearing primary minerals or partial melting.

### **4.3 Implications for the temperature and composition of the lithospheric mantle**

In this section we compare our seismic models with thermal and compositional constraints. We compute seismic velocity predictions based on realistic thermal and compositional models of the lithosphere. We first constrain the average lithospheric peridotite composition in Fennoscandia using mantle xenolith data, and assess the possible presence of eclogite and diamond-bearing components. We then estimate a range of steady-state conductive geotherms for Fennoscandia using xenolith pressure–temperature (P–T) constraints and observed surface heat flow. Finally, we predict  $V_{sv}$  using forward modeling that incorporates the inferred thermal and compositional structures, accounting for anelastic effects and radial anisotropy, and compare these predictions with seismic observations.

#### **4.3.1 Mantle Xenolith Constraints on Composition**

Mantle xenoliths transported to the Earth’s surface by kimberlite magmas provide direct constraints on the composition and thermal state of the cratonic mantle lithosphere.

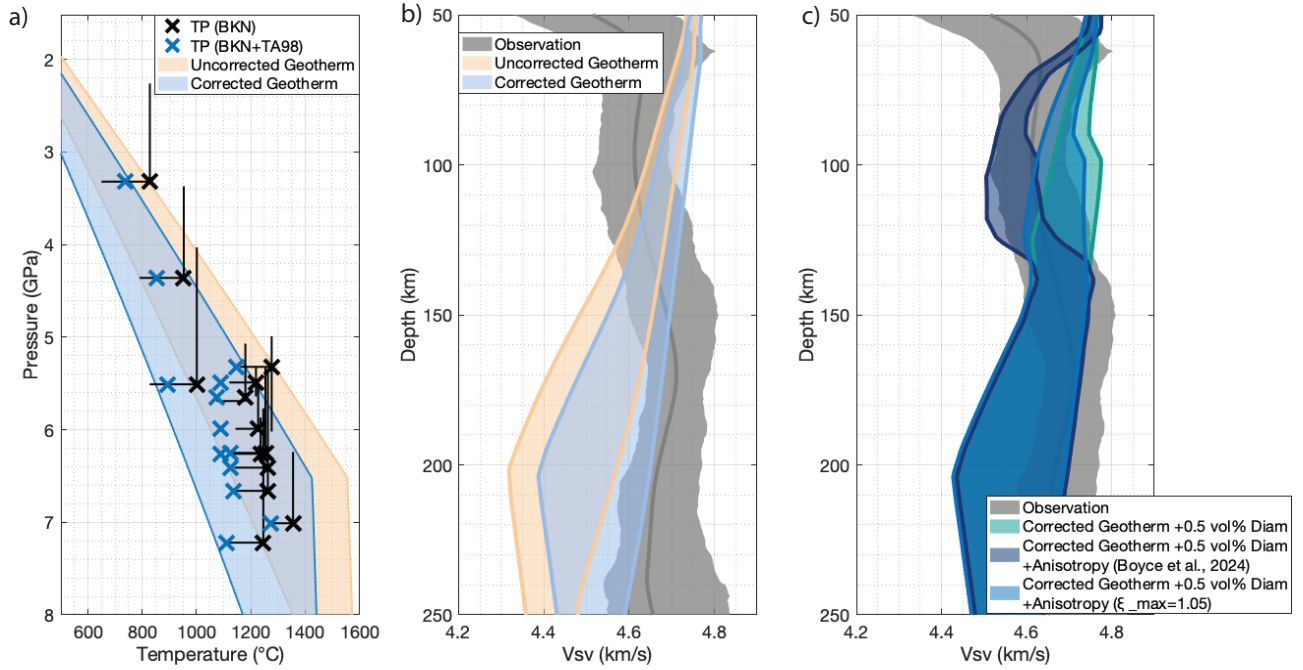
The Kaavi–Kuopio Kimberlite Province in eastern Finland (gray circle in Fig. 9a) contains at least 19 kimberlite intrusions emplaced into Archean basement gneisses and allochthonous Proterozoic metasediments, near an ancient suture between the Karelian craton and the Svecofennian mobile belt (Lehtonen et al. 2004). Kimberlite emplacement ages are estimated at 589–626 Ma based on U-Pb ion probe dating of perovskites (O’Brien et al. 2005) with a minimum age of 530 Ma from an internal Sm-Nd isochron (P Peltonen et al. 1998; Kukkonen and Peltonen 1999). Based on garnet peridotite xenoliths, Peltonen et al. (1998) divided the cratonic lithosphere into two main layers: (1) a shallower (90–130 km), strongly depleted garnet-spinel facies zone, interpreted as remnants of the Archean subcontinental lithospheric mantle, containing trace amounts of phlogopite and amphibole; and (2) a deeper (170–220 km), more compositionally diverse and fertile garnet facies mantle, interpreted as Proterozoic additions of possible cumulate origin. Subsequent studies divided the lithospheric mantle into three layers based on pressure-temperature (P-T) estimates from mantle xenocrysts and xenoliths (Lehtonen et al. 2004) as well as Re-Os isotopic composition and geochemical constraints (Peltonen and Brüggmann, 2006). In the three-layer model, the middle layer (~110–180 km) was interpreted as unmodified Archean continental mantle.

In addition to garnet peridotite xenoliths, eclogite xenoliths are also present in the Kaavi–Kuopio Kimberlite Province, although they occur only in minor amounts, representing less than ~1% of the mantle-derived xenolith suite (P Peltonen et al. 2002). Some of these eclogites are diamondiferous, with the diamond content of ~0.88–1 vol.% (P Peltonen et al. 1998; B. J. Griffin et al. 1995), and are inferred to originate from depths of ~150–220 km.

For the thermodynamic modeling described below, we adopt the average Baltic Shield peridotite xenolith composition compiled by Garber et al. (2018), which they obtained from the GEOROC database [<http://georoc.mpch-mainz.gwdg.de>]. This composition is derived from the Kaavi–Kuopio Kimberlite Province xenoliths. For an eclogite composition, we use the average eclogite xenolith composition from eastern Finland reported by Peltonen et al. (2002). The compositions are shown in Table S1.

#### **4.3.2 Mantle Xenolith Constraints on Temperature**

Peridotite xenoliths provide constraints not only on lithospheric composition but also on lithospheric P–T conditions, which we use to define an acceptable subset of geotherms (Fig. 13a). We consider two sets of P–T estimates, which are based on 13 peridotite mantle xenoliths from the Kaavi–Kuopio Kimberlite Province. The first set is from Kukkonen and Peltonen (1999), who used the two-pyroxene thermometer and Al-in-orthopyroxene barometer of Brey and Kohler (1990). These estimates, shown as black crosses in Fig. 13a with the “BKN” label, were assumed to be the most reliable among the available thermobarometric combinations. To illustrate thermobarometric uncertainty, black error bars represent the range of P–T estimates obtained from additional calibrations reported by Kukkonen and Peltonen (1999), such as the two-pyroxene thermometer of Finnerty and Boyd (1987) and the garnet–orthopyroxene barometer of MacGregor (1974). We refer to this first data set as the uncorrected P–T data.



**Figure 13.** (a) Xenolith thermobarometry and calculated conductive geotherms. Black crosses show T(BKN) and P(BKN) from Kukkonen and Peltonen (1999); error bars indicate the range from the three thermobarometric combinations reported therein. The yellow shaded region denotes the acceptable geotherm range constrained by the uncorrected thermobarometry. Blue crosses show temperatures corrected following Nimis and Grütter (2010) from BKN (Brey and Köhler 1990) to TA98 (Taylor 1998b), and the blue shaded region denotes the corresponding acceptable (colder) geotherm range. (b) Comparison of the observed Vsv model in the Fennoscandian craton interior from this study (gray shading; gray line indicates the mean) with isotropic Vsv predicted from thermodynamic modeling for peridotite along the hotter (yellow) and colder (blue) geotherm ranges in (a). (c) Comparison of observed Vsv (gray shading) with predictions under more complex assumptions. Light green shading: peridotite + 0.5 vol% graphite/diamond along the colder geotherm range (blue in a), assuming isotropy. Dark blue: same model but including strong radial anisotropy from Boyce et al. (2024), with peak  $\xi \approx 1.2$ . Light blue: same as dark blue, but with weak radial anisotropy (peak  $\xi \approx 1.05$ ).

The second data set accounts for two sources of possible bias in the BKN P-T estimates. One, Nimis and Grütter (2010) show that the Brey and Köhler (1990) thermometer exhibits a systematic positive temperature bias that increases with sodium (Na) content in clinopyroxene. It can be corrected using the empirical relationship:  $\Delta T$

$(^{\circ}\text{C}) = 740.3 \cdot \text{Na}_{\text{Cpx}} (\text{apfu}) - 35.7$ , where  $\text{Na}_{\text{Cpx}}$  is the sodium content in clinopyroxene expressed in atoms per formula unit (apfu). The correction term,  $\Delta T$  is subtracted from  $T_{\text{BKN}}$  to remove the Na-content bias and obtain temperatures consistent with the Taylor (1998) two-pyroxene thermometer, which is considered more reliable. We apply this correction using the  $\text{Na}_{\text{Cpx}}$  values reported Kukkonen and Peltonen (1999). Two, we attempt to account for the additional cooling of the lithosphere that has occurred since kimberlite emplacement at  $\sim 530$  Ma. It produces  $\sim 82.24$   $^{\circ}\text{C}$  of cooling using a half space cooling model at 200 km and similar values at other depths. Together, these two corrections reduce the estimated temperature by an average of  $119.89$   $^{\circ}\text{C}$ . The corrected P–T estimates, which we will refer to as the corrected P-T data, are shown as blue crosses in Fig. 13a.

We next calculate a set of steady state conductive geotherms that simultaneously satisfy the xenolith P–T constraints described above and observed surface heat flow observations through a parameter search. We follow the methodology of Hirsch et al. (2015), which is summarized here. We solve the 1-D steady state heat conduction equation using a surface temperature of  $0$   $^{\circ}\text{C}$  and measured surface heat flow as boundary conditions. Temperature and heat flow are calculated from the surface to 400 km depth using 2000 layers of 200 m thickness. We adopt a two-layer continental crustal structure, with the base of the upper crust fixed at 10 km in all scenarios, and a single layer for the part of the uppermost mantle in which heat transfer occurs dominantly by conduction (i.e., the extension of the thermal boundary layer into the mantle). Values of 45, 50, and 55 km are tested for the base of the crust based on Moho depths obtained from the CCP stacks (Fig. 5a). The base of the mantle thermal boundary layer is defined as the depth at

which the conductive geotherm intersects the mantle adiabat. In our parameter search, we consider adiabats corresponding to a range of mantle potential temperatures from 1275 to 1500 °C in 25 °C increments. We search over four additional parameters: radiogenic heat production in the upper crust, in the lower crust, and in the mantle as well as thermal conductivity, following the values considered by Hirsch et al. (2015). Surface heat flow is restricted to 25-35 mW/m<sup>2</sup>, based on the low heat flow observed in Fennoscandia (Veikkolainen et al. 2017).

In total, 2930 geotherms are generated and compared to the thirteen xenolith P-T estimates in order to identify the subset of geotherms that provide a reasonable fit to the observations. We calculate the misfit for each geotherm using the maximum temperature difference between the geotherm and each xenolith. From this misfit, we calculate a likelihood for each geotherm. We select the minimum likelihood threshold that approximately brackets the xenolith P-T data and define the corresponding maximum and minimum temperatures at 220 km depth as the upper and lower bounds of the acceptable geotherm range. In Fig. 13a, the light-yellow shaded zone denotes the acceptable geotherm range defined by the uncorrected P-T estimates, with the thick yellow lines representing the cold and hot bounds on temperature. The blue shaded region with the thick blue lines represents the acceptable geotherm range that fits the corrected P-T estimates, with corresponding cold and hot bounds. Below, we refer to these as the uncorrected and corrected geotherm ranges.

### 4.3.3 Thermodynamic modeling

We use HeFESTo (Stixrude and Lithgow-Bertelloni 2005) following the workflow of Garber et al. (2018) to convert the compositional and thermal structures described above to  $V_{sv}$  in five steps. The forward modeling proceeds as follows: (1) Phase equilibria and mineral modes are calculated along the bounds of the acceptable geotherm ranges using HeFESTo for peridotite and eclogite compositions (Table S1). (2) Shear moduli for each solid-solution phase are calculated using the Reuss average of end-member elastic moduli. (3) Bulk rock shear moduli are computed using the Voigt–Reuss–Hill average over all solid-solution phases. (4) Where multiple lithologies are present, we assume purely mechanical mixing and compute the bulk cratonic shear modulus using a Voigt–Reuss–Hill average weighted by the volume fractions of peridotite, eclogite, and/or diamond. Bulk density is calculated using a Voigt average of volume-weighted densities. For diamond, steps 1-3 are performed using `Perple_X` (Connolly 2005) with the diamond thermodynamic dataset of Valdez et al. (2012) and graphite data from Holland and Powell (1998), similar to what Garber did for diamond. For peridotite and eclogite, we use HeFESTo with the thermodynamic dataset STX24 (Stixrude and Lithgow-Bertelloni 2005, 2024), which is the primary difference between our modeling approach and that of Garber et al. (2018), who used `Perplex_X` with dataset STX11 (Stixrude and Lithgow-Bertelloni 2011). (5) These calculations produce predictions of elastic and isotropic  $V_s$ , but the Bayesian joint inversion described in Section 2.3 produces models of  $V_{sv}$  that include anelastic effects. All predicted  $V_s$  profiles are adjusted for anelasticity in order to allow a meaningful comparison with the inverted velocity profiles. Following Garber et al. (2018), we use a Burgers model (Jackson and Faul 2010) applied to the geotherms to

calculate the anelastic moduli and the corrected Vs at 1 Hz with the grain size of 1 cm. In some cases we also attempt to account for radial anisotropy, which is described below.

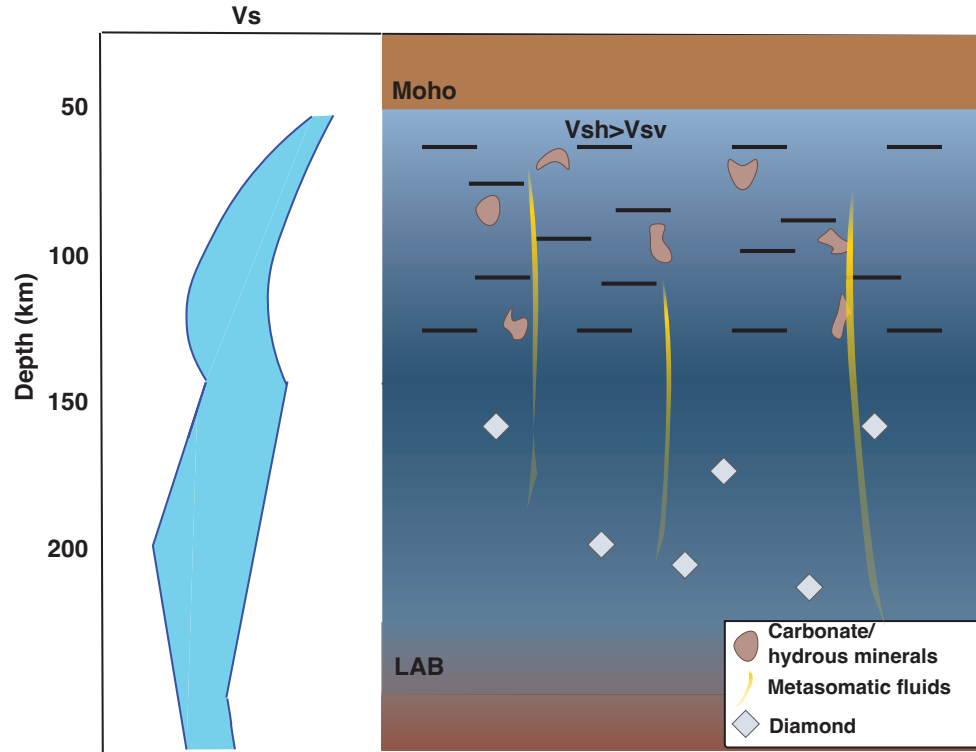
Fig. 13b compares the inverted Vsv models for the Fennoscandia craton interior with the Vs predictions based on a constant peridotite composition (Table S1) and the uncorrected and corrected geotherm ranges. For the coldest geotherms, for which the thermal boundary layer intersects the mantle adiabat at depths  $> 250$  km, the predicted shear-wave velocities decrease throughout the depth range, whereas Vs for the warmest geotherms reaches a minimum near the intersection of the conductive geotherms and the adiabat and increases with depth below this point. With the uncorrected geotherm range, we reproduce the conclusion of Garber et al. (2018) that peridotite alone cannot explain the fast seismic velocity observed in the lithosphere at depths  $> \sim 150$  km. Predicted velocities are slower than observed deeper than  $\sim 150$  km and slightly faster than observed shallower than  $\sim 100$  km. Moreover, the predicted velocity profiles are relatively smooth, particularly for the coldest geotherms, whereas the inverted profiles exhibit at least two distinct layers. Vs predictions made with the corrected geotherm range are faster than those with the uncorrected geotherm range and provide a better fit to the inverted profiles at depths  $> 130$  km, although most inverted models remain faster than the predictions.

In Fig. 13c we explore if we can improve the agreement between the inverted models and the predictions made with the corrected geotherm range by allowing 0.5 vol.% diamond, an amount that is consistent with rough estimates from global diamond-bearing peridotite xenoliths (Viljoen et al. 2004; Stachel and Luth 2015), in addition to peridotite. The inclusion of diamond increases predicted Vs by  $\sim 0.04$  km/s in the depth range where



diamond is stable, producing an improved match at depths  $> \sim 130$  km. As noted by Garber et al. (2018), the graphite-diamond transition depth depends on the geotherm, occurring at shallower depths for colder geotherms and greater depths for warmer geotherms.

We also try to account for radial anisotropy by implementing two different estimates of how the speed of horizontally polarized shear waves ( $V_{sh}$ ) is faster than  $V_{sv}$  in the lithospheric mantle. With these tests, the predicted isotropic  $V_s$  is converted into  $V_{sv}$  by assuming  $V_s = \frac{1}{3} V_{sh} + \frac{2}{3} V_{sv}$ . The Baltic Shield model of Boyce et al. (2024), which is estimated from Rayleigh and Love wave phase velocities, contains positive radial anisotropy ( $\xi = (\frac{v_{SH}}{v_{SV}})^2 > 1$ ) between  $\sim 55$ -140 km depth, with a peak  $\xi > 1.2$ . A second estimate of  $\xi$  comes from [Zhu et al. \(2015\)](#) using full waveform inversion, suggesting a weaker radial anisotropy in the Fennoscandian craton interior with  $\xi \approx 1.1$  in the uppermost mantle. In both cases, the radial anisotropy correction reduces predicted  $V_{sv}$  in the shallow lithosphere relative to isotropic  $V_s$  (Fig. 13c), improving agreement with the observed low velocities at depths  $< \sim 130$  km. A model combining the corrected geotherm range,  $\sim 0.5$  vol.% diamond, and strong radial anisotropy yields the best agreement with observations in both absolute velocity and the shape of the vertical structure.



*Figure 3.14: Figure 14.* Schematic diagram of the layered structure of the cratonic lithosphere inferred from this study. Fennoscandia exhibits a colder geotherm than that suggested by uncorrected xenolith P–T constraints. Only a moderate abundance of diamond is present in the lower lithosphere below ~150 km depth. Above this depth, carbonate and/or hydrous minerals formed through the upward migration of metasomatic fluids likely contribute to the observed low  $V_{sv}$  structure, accompanied by radial anisotropy within this layer.

#### 4.3.4 Implications of Comparison between Forward Modeling and Seismic Observations in Fennoscandia

Comparison between observed seismic velocities in the Fennoscandian craton interior and predictions from isotropic peridotite models along uncorrected geotherms inferred from Brey and Kohler (1990) thermobarometry confirms the conclusion of Garber et al. (2018): such models fail to reproduce both the absolute velocities and

vertical gradients of the seismic observations (Fig. 13b). Specifically, the observed lithosphere exhibits two layers: an upper layer shallower than ~150 km that is slower than predicted, and a deeper layer that is significantly faster than predicted. We therefore favor an Earth model (Fig. 13c) characterized by a colder lithospheric geotherm, peridotite composition containing a moderate diamond fraction (~0.5 vol.%), and strong radial anisotropy ( $\xi \approx 1.2$ ), potentially accompanied by carbonate or hydrous minerals with intrinsically low shear-wave velocities in the shallow lithosphere (<150 km) (Fig. 14). This combination provides the closest match to both the magnitude and shape of the observed seismic velocity profile.

We identify five lines of evidence supporting the interpretation that the deep lithosphere beneath Fennoscandia is colder than implied by uncorrected xenolith thermobarometry. (1) The uncorrected geotherm inferred from xenolith thermobarometers from the widely used Brey and Kohler two-pyroxene thermometer (1990) exhibits a systematic positive temperature bias that increases with  $\text{Na}_{\text{cpx}}$  (+150°C at  $\text{Na}_{\text{cpx}} = 0.25$ ) (Nimis and Grütter 2010) and does not account for the conductive cooling since kimberlite emplacement. These two effects yield an average temperature reduction of ~120 °C for the Fennoscandian xenolith suite (Fig. 13a). (2) Different thermobarometric calibrations applied to different sample types introduce substantial uncertainty. Compilations of P-T estimates for xenoliths in southern Finland, cpx-garnet pairs and cpx xenocrysts using multiple thermobarometers span ~180 °C at 4.5 GPa and ~300 °C at 6GPa (Lehtonen et al. 2004). For the 13 xenoliths examined here, Kukkonen and Peltonen (1999) reported large uncertainties(>100 °C and >0.6GPa) where most corrected temperatures from (1) fall within these uncertainty bounds (Fig. 13a). (3) As in other

cratons (e.g., Kaapvaal and Slave), xenolith thermobarometry may underestimate the long-term depth extent of the petrologic lithosphere due to transient thermal and compositional perturbations associated with kimberlite magmatism, biasing inferred geotherms toward warmer conditions (Eaton et al. 2009). (4) Kimberlites tend to cluster near the margins of high-velocity SCLM domains rather than their interiors (W. L. Griffin et al. 2009), implying that xenolith suites may preferentially sample lower-velocity, more disturbed lithosphere. The Kaavi–Kuopio kimberlites are located near a major suture, and it remains unclear whether they sample true cratonic interior mantle, since our robust measurements in the southern Finland are limited. (5) Xenoliths record transient thermal states associated with magmatic events rather than the long-term stable geotherm of the lithosphere, an effect that may cannot be ignored for kimberlite magmatism (Deen et al. 2006). Together, these considerations suggest that the corrected cold geotherms represent an upper bound on lithospheric temperatures, and that the true cratonic geotherm may be even colder, yielding faster deep lithospheric velocities.

To further evaluate the need for cold temperatures in the deep mantle lithosphere, we consider a wider range of compositional variations that may contribute to the high deep-lithospheric velocities. We evaluate three possibilities: diamond, eclogite, highly depleted peridotite. Garber et al. (2018) proposes 1-3 vol.% diamond could reconcile observed velocities with hot geotherm predictions. In Fennoscandia, 1.5–2 vol.% diamond combined with uncorrected geotherms reproduces observed velocities (Fig. S8a), but such abundances far exceed constraints from xenolith data in both Fennoscandia and global cratons (Viljoen et al. 2004; Stachel and Luth 2015; P Peltonen et al. 2002), although Garber et al. (2018) argued there could be a sampling bias of the particular

kimberlite, or preferential preservation of certain lithologies. We therefore adopt a more conservative diamond fraction of 0.5 vol.%. Adding up to 20 vol% eclogite, the maximum allowed by buoyancy constraints (Garber et al., 2018), produces only a 0.0185 km/s velocity increase at 200 km depth (Fig. S8b), insufficient to resolve the discrepancy. Similarly, using peridotite compositions representative of the inferred unmodified Archean Layer B in Peltonen and Brügmann (2006), the predicted velocity is very similar to predicted velocity using average peridotite composition from all xenoliths (Fig. S8c), even slightly lower. Highly depleted ARC-9 peridotite ( $Mg\# = 93.1$ ; Griffin et al. 2009) yields only modest velocity increases  $\sim 0.03367$  km/s; Fig. S8c), still below observed values. These results indicate that colder geotherms remain essential, even when compositional anomalies are considered.

Uncertainties associated with thermodynamic databases, anelastic corrections, and radial anisotropy also influence predicted velocities. Using corrected cold-limit geotherms and peridotite compositions, we test several thermodynamic datasets (Fig. S9). The STX24 (Stixrude and Lithgow-Bertelloni 2024) yields very similar  $V_{sv}$  prediction as STX11(Stixrude and Lithgow-Bertelloni 2011) using HeFESTo, with differences less than 0.01 km/s. Using STX11, HeFESTo(Stixrude and Lithgow-Bertelloni 2005) and Perple\_X (Connolly 2005) produce consistent predictions with negligible differences. In contrast, Perple\_X calculations using dataset HP02, the 2002 version of the thermodynamic data base from Holland and Powell (1998) yields higher velocities with different gradients. With HP02 combined with the elastic parameter compilation of Abers and Hacker (2016)(AH16) for the compositions in terms of solid-solution end members as obtained from HP02, the predicted  $V_s$  (HP02+AH16) produces velocities  $\sim 0.056$  km/s

faster than STX24. These differences reflect variations in both solid-solution models and elastic parameters and are consistent with findings of Garber et al. (2018) and Eeken et al. (2018). We choose to use the most up-to-dated STX24 in the forward modeling calculations.

Anelastic corrections significantly reduce shear velocities relative to purely elastic predictions and depend on temperature, frequency, and grain size. Following Garber et al. (2018), we set grain size and frequency at 1 cm and 1 Hz, respectively; smaller grain sizes or lower frequencies would further reduce velocities. However, velocities along the cold geotherm limit are only weakly affected by anelastic corrections.

Applying radial anisotropy model in Baltic Shield from Boyce et al. (2024) reduces predicted  $V_{sv}$  in the shallow lithosphere ( $<150$  km), bringing predictions into closer agreement with observations. Nevertheless, uncertainties remain: the anisotropy model from Boyce et al. (2024) was derived from a single pixel in southern Finland, different inversion approaches (Bayesian inversion vs. least square inversion) yield different  $\xi$  amplitudes, and alternative models (Zhu et al., 2015) suggested weaker anisotropy ( $\xi \approx 1.1$ ). If weaker anisotropy is adopted, predicted  $V_{sv}$  from peridotite + diamond models becomes faster than observed in the shallow lithosphere (Fig. 13c).

In such cases, low-velocity hydrous or carbonate phases produced by metasomatism provide a plausible explanation for shallow lithospheric velocity reductions. While pargasitic amphibole is commonly invoked (Selway et al. 2015; Saha et al. 2021; Sudholz et al. 2024), it is stable only above  $\sim 3$  GPa in depleted peridotite. Phlogopite, stable over much of the cratonic lithospheric depth range, is therefore a more likely candidate (S. M. Hansen et al. 2015; Eeken et al. 2018; Rader et al. 2015; Saha et al., 2021). In addition,

carbonate phases such as dolomite (Zhao et al. 2024) and  $\text{CaCO}_3$  (Zhang et al. 2025) are also capable of producing isotropically low Vs. Eeken et al. (2018) proposed that such phases are restricted to depths shallower than  $\sim 150$  km because metasomatic fluids freeze and react where temperatures intersect hydrous or carbonated solids, consistent with our results. The precise abundance of these low-velocity phases, however, remains dependent on the magnitude of radial anisotropy.

#### **4.4 Indication of the layered lithosphere and global cratons**

Layered cratonic lithosphere, often with a negative mid-lithospheric discontinuity (MLD) at  $\sim 60$ – $110$  km, has been widely reported beneath many cratons (Abt et al. 2010; Chen 2017; Fischer et al. 2020; Rychert and Shearer 2009; Boyce et al., 2021; Carr and Olugboji, 2024; Legre and Olugboji, 2025; Liu et al., 2023). In contrast, our preferred model for the Fennoscandian craton interior is characterized by a positive velocity gradient (PVG) at greater depth ( $\sim 160$  km). Although less commonly emphasized, layered cratonic structures featuring PVGs at similar depths have also been observed globally. For example, global Sp receiver-function analyses identified clear PVGs near  $\sim 150$  km beneath stations where  $V_{sv}$  exceeds  $4.75$  km/s at  $110$ – $130$  km depth (Hua et al. 2023). In addition, global Sp studies found pronounced PVGs in southern Africa and other cratons at depths of  $\sim 150$ – $250$  km (Krueger et al. 2021), and Ps receiver functions have found PVGs over a wide range of depths in the US (Carr and Olugboji, 2024) and Africa (Legre and Olugboji, 2025). Global craton models satisfying attenuation data additionally required a low velocity zone in the shallow lithosphere and faster velocity in deep lithosphere to fit phase velocity (Dalton et al. 2017). Similarly, active-source experiments had identified PVGs at  $\sim 150$ – $250$  km that mark the base of low-velocity

zones beneath cold continental regions (Hans Thybo 2006). Comparable PVGs at ~150–250 km have been reported across the North American craton (Calò et al. 2016) and Kalahari craton (Wittlinger and Farra 2007). This style of layered structure is also consistent with cratonic mantle models that include anisotropy: global radial anisotropy models (Boyce et al. 2024), radial and azimuthal anisotropy beneath the North American craton (Yuan and Romanowicz 2010), and azimuthal anisotropy in the Southern Africa craton (L. Vinnik et al. 2012) all suggested two lithospheric layers characterized by changes in anisotropy magnitude and/or fast-axis orientation. Taken together, these observations suggested that the velocity structure inferred beneath Fennoscandia may reflect a broader pattern of global cratonic layering.

A key implication of our study is that the cratonic interior may be colder than inferred from xenolith thermobarometry. In Fig. 13a, the corrected cold geotherm range (blue) should be viewed as an upper bound on the true interior geotherm, which may be even colder. Under corrected geotherms, a “normal” peridotitic composition (with a moderate diamond fraction) can reproduce the high velocities observed in the deep lithosphere. This possibility shifts the primary explanation for lithospheric layering away from requiring an anomalously fast deep layer, and toward an abnormally slow shallow layer. The possibility that seismic data require colder lithosphere than xenolith thermometry has been noted not only in Fennoscandia (Ramirez et al. 2024), but also in global synthesis craton studies (Eeken et al. 2018).

Several mechanisms have been proposed to explain layered cratonic upper mantle structure, including (1) depletion variations created by melt extraction (Yuan and Romanowicz 2010), (2) partial melt (H. Thybo and Perchuc 1997; Perchuc and Thybo



1996), (3) elastically accommodated grain boundary sliding (Shun-ichiro Karato 2012), (4) radial anisotropy (Wirth and Long 2014), and (5) metasomatic minerals (Rader et al. 2015; Selway et al. 2015; Aulbach et al. 2017; Saha et al. 2021). In our preferred interpretation, the shallow low-velocity layer is most plausibly explained by a combination of radial anisotropy and metasomatic minerals.

Depletion-related compositional layering alone is unlikely to explain the observed shallow low velocities. Depletion is expected to produce only modest shear-velocity changes (1%–1.7%; Afonso and Schutt 2012), and geochemical evidence suggests that the upper lithosphere tends to be more depleted than the lower lithosphere (W. L. Griffin et al. 2009), which would predict faster, not slower velocities in the shallow layer. Partial-melt explanations are also disfavored in our preferred thermal regime. The “8° discontinuity” has been interpreted as marking the onset of partial melting where geotherms intersect volatile-bearing solidi (H. Thybo and Perchuc 1997), and Liu et al. (2023) proposed carbonated melt and melt due to decomposition of hydrous minerals. Our results do not support either scenario. In regard to H. Thybo and Perchuc (1997), the top boundary of the low velocity zone (referred to as the 8° discontinuity) is interpreted as the initiation of partial melting because of the intersection of the geotherm and solidus curves for peridotite with small amounts of C-H-O. For the latter study, Liu et al. (2023) interpreted low velocities initiating near ~110 km in North America as carbonated melt where geotherms intersect the magnesite–carbonated melt boundary. They further suggested that shallower low-velocity zones may reflect melts produced by decomposition of hydrous minerals, because CO<sub>2</sub>-rich fluids promote melt generation during metasomatism. However, the corrected geotherms favored here are too cold to

intersect either volatile-bearing solids or key boundaries such as magnesite-out and phlogopite-out (Fig. S10). In regard to elastically-accommodated grain boundary sliding, an analysis of recent experimental results concluded that water (>100 ppm) would need to be present to produce a significant velocity reduction (Shun-ichiro Karato 2025) and this requirement likely exceeds water contents observed for most cratons (Peslier et al. 2017; Chin and Palin 2022).

As discussed in session 4.3.4, radial anisotropy and metasomatic minerals provide two plausible explanations for the shallow low-velocity layer, but we cannot uniquely separate them because Rayleigh-wave phase velocities primarily constrain  $V_{sv}$ . Positive radial anisotropy ( $V_{sh} > V_{sv}$ ) in the shallow lithosphere beneath Fennoscandia has been inferred in several studies, although with differing amplitudes:  $\xi \approx 1.05$  (Valérie Maupin et al. 2022),  $\xi \approx 1.1$  (Zhu et al. 2015), and  $\xi \approx 1.2$  (Boyce et al. 2024). We therefore treat the Boyce et al. (2024) estimate as an upper bound on  $\xi$ . If instead we impose a weaker anisotropy (e.g., peak  $\xi \approx 1.05$ ; Fig. 13c), representing a plausible lower bound, then the observed  $V_{sv}$  requires an additional low-velocity compositional component in the shallow lithosphere.

Radial anisotropy is commonly attributed to the lattice-preferred orientation (LPO) of olivine in peridotite (e.g., Yuan and Romanowicz 2010), and Boyce et al. (2024) interpreted the radial anisotropy in their models in terms of a plume/underplating scenario in which upwelling material spreads laterally beneath a rigid plate, cools, and develops predominantly horizontal aligned mineral fabrics, producing  $V_{sh} > V_{sv}$  in the shallow lithosphere. Notably, metasomatic models can also contribute to anisotropy. In the “melt-percolation barrier” framework of Liu et al. (2023), predominantly horizontal

emplacement of veins and sills during metasomatism can both enhance radial anisotropy and reduce seismic velocities.

Except for the upper bound of radial anisotropy estimates, metasomatic minerals with intrinsically low Vs are likely required at least above ~100 km to explain the shallow low velocities in our preferred model. Two lines of evidence support metasomatism in Fennoscandia: (1) hydrous phases such as phlogopite and amphibole are observed in xenoliths from the Kaavi–Kuopio kimberlites; and (2) magnetotelluric results indicate that peridotite-only models consistent with seismic velocities predict conductivities that are too low above ~150 km, implying hydrous phases (Ramirez et al. 2024). Assuming weak anisotropy ( $\xi = 1.05$ ) and the hot-limit geotherm, the observed shallow lithosphere requires an additional ~2.7% reduction in Vsv, consistent with global estimates of shallow Archean velocity reductions (<3%; Krueger et al. 2021). At the cold temperatures inferred for the Fennoscandian cratonic interior, such a reduction could be explained by ~10% pargasitic amphibole, ~2.1% phlogopite, and ~0.2 magnesite, which is consistent with the expected abundances of these minerals in depleted cratonic mantle (Saha et al., 2021)., Note this estimate should be regarded as an upper bound on metasomatic mineral abundance because both a colder geotherm and stronger anisotropy would reduce the required fraction of low-velocity phases. The restriction of these phases to the shallow lithosphere may reflect either (i) the stability field of amphibole, or (ii) depth-dependent “freezing” of metasomatic fluids into hydrous/carbonate minerals as temperatures cross relevant solidi (Rader et al., 2015; Selway et al., 2015; Eeken et al. 2018). More broadly, the presence of metasomatic phases is consistent with modification of cratonic lithosphere by CO<sub>2</sub>-H<sub>2</sub>O-rich siliceous melts associated with subduction (R. T.

J. Hansen et al. 2012; Krueger et al. 2021; Selway et al. 2015; Aulbach et al., 2017) and/or cooling-induced water exsolution from nominally anhydrous mineral during craton cooling (Chin and Palin 2022). Given the similarity between our Fennoscandian results and global observations, a combination of colder cratonic interiors and shallow metasomatic modification could be widespread as the origin of layered cratonic lithosphere.

## 5 Conclusions

We measure Rayleigh wave phase velocities and Sp receiver functions across Fennoscandia and develop a 3-D shear-wave velocity ( $V_{sv}$ ) model through a Bayesian Markov chain Monte Carlo joint inversion of the two datasets. Synthetic tests demonstrate that this approach recovers both absolute velocity magnitudes and strong vertical gradients. Higher phase velocities at longer periods are observed in the eastern Svecofennian Paleoproterozoic and Archean cratonic regions, consistent with fast cratonic lithosphere. In the upper mantle, layered structures are resolved, including sporadic positive velocity gradients (PVGs) between  $\sim 140$ – $190$  km across the study region, shallower continuous negative velocity gradients (NVGs) near  $\sim 100$  km beneath the southeastern area, and deeper NVGs ( $\sim 200$  km) beneath northern Finland and central Sweden.

The resulting 3-D  $V_{sv}$  model can be classified into three clusters. The Paleoproterozoic Svecofennian Shield in Sweden and Finland, together with the southern Archean Karelian craton, defines the craton interior, where the highest velocities ( $>4.7$  km/s) occur predominantly at depths of  $\sim 140$ – $210$  km. In contrast, northern Finland

exhibits a similar shallow lithospheric structure above ~150 km but a lower velocity underlying mantle, representing the craton edge and likely reflecting compositional modification near the craton margin.

To interpret these seismic observations, we perform thermodynamic forward modeling using HeFESTo that incorporates thermal and compositional constraints from xenoliths, including anelastic effects and radial anisotropy. A corrected, colder geotherm, that accounts for xenolith thermobarometry bias and conductive cooling, combined with a moderate diamond fraction (~0.5 vol.%), reconciles the high velocities observed in the lower lithosphere (~>150 km). In contrast, the reduced velocities in the shallow uppermost mantle are best explained by radial anisotropy and/or the presence of carbonate or hydrous minerals introduced through metasomatism.

Overall, our results support a model of cratonic lithosphere characterized by a layered structure, with a low velocity, metasomatized upper layer and a colder, higher velocity lower layer. The similarity between the features observed in Fennoscandia and those reported in other cratons suggests that a cold geotherm and predominantly peridotite composition, modified by metasomatism, may represent a common framework for layered cratonic lithosphere.

## 6 References

- Abers, Geoffrey A., and Bradley R. Hacker. 2016. "A MATLAB Toolbox and E Xcel Workbook for Calculating the Densities, Seismic Wave Speeds, and Major Element Composition of Minerals and Rocks at Pressure and Temperature." *Geochemistry, Geophysics, Geosystems* 17 (2): 616–24.  
<https://doi.org/10.1002/2015GC006171>.
- Abt, David L., Karen M. Fischer, Scott W. French, Heather A. Ford, Huaiyu Yuan, and Barbara Romanowicz. 2010. "North American Lithospheric Discontinuity Structure Imaged by *Ps* and *Sp* Receiver Functions." *Journal of Geophysical Research: Solid Earth* 115 (B9): 2009JB006914.  
<https://doi.org/10.1029/2009JB006914>.
- Afonso, Juan Carlos, and Derek L. Schutt. 2012. "The Effects of Polybaric Partial Melting on Density and Seismic Velocities of Mantle Restites." *Lithos* 134–135 (March): 289–303. <https://doi.org/10.1016/j.lithos.2012.01.009>.
- Alinaghi, A., G. Bock, R. Kind, et al. 2003. "Receiver Function Analysis of the Crust and Upper Mantle from the North German Basin to the Archaean Baltic Shield." *Geophysical Journal International* 155 (2): 641–52.  
<https://doi.org/10.1046/j.1365-246X.2003.02075.x>.
- Artemieva, Irina M. 2007. "Dynamic Topography of the East European Craton: Shedding Light upon Lithospheric Structure, Composition and Mantle Dynamics." *Global*

*and Planetary Change* 58 (1–4): 411–34.

<https://doi.org/10.1016/j.gloplacha.2007.02.013>.

Audet, Pascal, Colin Thomson, Michael Bostock, and Tom Eulenfeld. 2019.

“Telewavesim: Python Software for Teleseismic Body Wave Modeling.” *Journal of Open Source Software* 4 (44): 1818. <https://doi.org/10.21105/joss.01818>.

Aulbach, Sonja, Malcolm Massuyeau, and Fabrice Gaillard. 2017. “Origins of Cratonic

Mantle Discontinuities: A View from Petrology, Geochemistry and

Thermodynamic Models.” *Lithos* 268–271 (January): 364–82.

<https://doi.org/10.1016/j.lithos.2016.11.004>.

Balling, Niels, Frederik Tilmann, and Rainer Kind. 2011. “The Jutland-to-Lower-Saxony

(JULS) Profile.” With Deutsches GeoForschungsZentrum GFZ, GEOFON Data

Centre, [geofon@gfz.de](mailto:geofon@gfz.de), et al. GFZ Data Services. ..Mseed,XML, 833GB.

<https://doi.org/10.14470/MJ7565224230>.

Bock, Günter, U. Achauer, A. Alinaghi, et al. 2001. “Seismic Probing of Fennoscandian

Lithosphere.” *Eos, Transactions American Geophysical Union* 82 (50): 621–29.

<https://doi.org/10.1029/01EO00356>.

Boyce, Alistair, Thomas Bodin, Stéphanie Durand, Dorian Soergel, and Eric Debayle.

2024. “Seismic Evidence for Craton Formation by Underplating and Development of the MLD.” *Geophysical Research Letters* 51 (4): e2023GL106170.

<https://doi.org/10.1029/2023GL106170>.

- Brey, G. P., and T. Köhler. 1990. “Geothermobarometry in Four-Phase Lherzolites II. New Thermobarometers, and Practical Assessment of Existing Thermobarometers.” *Journal of Petrology* 31 (6): 1353–78.  
<https://doi.org/10.1093/petrology/31.6.1353>.
- Brown, Sarah E., and Karen M. Fischer. 2025. “Investigating the Antarctic Lithosphere Through Sp Receiver Function Analysis.” *Geochemistry, Geophysics, Geosystems* 26 (10): e2025GC012268. <https://doi.org/10.1029/2025GC012268>.
- Bruneton, Marianne, Helle A. Pedersen, Pierre Vacher, et al. 2004. “Layered Lithospheric Mantle in the Central Baltic Shield from Surface Waves and Xenolith Analysis.” *Earth and Planetary Science Letters* 226 (1–2): 41–52.  
<https://doi.org/10.1016/j.epsl.2004.07.034>.
- Bulut, N., H. Thybo, and V. Maupin. 2022. “Highly Heterogeneous Upper-Mantle Structure in Fennoscandia from Finite-Frequency *P* -Body-Wave Tomography.” *Geophysical Journal International* 230 (2): 1197–214.  
<https://doi.org/10.1093/gji/ggac107>.
- Calò, M., T. Bodin, and B. Romanowicz. 2016. “Layered Structure in the Upper Mantle across North America from Joint Inversion of Long and Short Period Seismic Data.” *Earth and Planetary Science Letters* 449 (September): 164–75.  
<https://doi.org/10.1016/j.epsl.2016.05.054>.



- Carr, Steve A. B., and Tolulope Olugboji. 2024. "A Taxonomy of Upper-Mantle Stratification in the US." *Journal of Geophysical Research: Solid Earth* 129 (5): e2024JB028781. <https://doi.org/10.1029/2024JB028781>.
- Chen, Ling. 2017. "Layering of Subcontinental Lithospheric Mantle." *Science Bulletin* 62 (14): 1030–34. <https://doi.org/10.1016/j.scib.2017.06.003>.
- Chin, Emily J., and Richard M. Palin. 2022. "Water Storage in Cratonic Mantle." *Terra Nova* 34 (5): 369–80. <https://doi.org/10.1111/ter.12599>.
- Cocks, L. R. M., and T. H. Torsvik. 2006. "European Geography in a Global Context from the Vendian to the End of the Palaeozoic." *Geological Society, London, Memoirs* 32 (1): 83–95. <https://doi.org/10.1144/GSL.MEM.2006.032.01.05>.
- Connolly, J. A. D. 2005. "Computation of Phase Equilibria by Linear Programming: A Tool for Geodynamic Modeling and Its Application to Subduction Zone Decarbonation." *Earth and Planetary Science Letters* 236 (1–2): 524–41. <https://doi.org/10.1016/j.epsl.2005.04.033>.
- Dahlen, F. A., and Jeroen Tromp. 1998. *Theoretical Global Seismology*. Princeton University Press.
- Dalton, Colleen A., Xueyang Bao, and Zhitu Ma. 2017. "The Thermal Structure of Cratonic Lithosphere from Global Rayleigh Wave Attenuation." *Earth and Planetary Science Letters* 457 (January): 250–62. <https://doi.org/10.1016/j.epsl.2016.10.014>.

- Daly, J. Stephen, Victor V. Balagansky, Martin J. Timmerman, and Martin J. Whitehouse. 2006. "The Lapland-Kola Orogen: Palaeoproterozoic Collision and Accretion of the Northern Fennoscandian Lithosphere." *Geological Society, London, Memoirs* 32 (1): 579–98. <https://doi.org/10.1144/GSL.MEM.2006.032.01.35>.
- Deen, Tara J., W. L. Griffin, G. Begg, Suzanne Y. O'Reilly, L. M. Natapov, and J. Hronsky. 2006. "Thermal and Compositional Structure of the Subcontinental Lithospheric Mantle: Derivation from Shear Wave Seismic Tomography." *Geochemistry, Geophysics, Geosystems* 7 (7): 2005GC001120. <https://doi.org/10.1029/2005GC001120>.
- Eaton, David W., Fiona Darbyshire, Rob L. Evans, Herman Grütter, Alan G. Jones, and Xiaohui Yuan. 2009. "The Elusive Lithosphere–Asthenosphere Boundary (LAB) beneath Cratons." *Lithos* 109 (1–2): 1–22. <https://doi.org/10.1016/j.lithos.2008.05.009>.
- Eeken, Thomas, Saskia Goes, Helle A. Pedersen, Nicholas T. Arndt, and Pierre Bouilhol. 2018. "Seismic Evidence for Depth-Dependent Metasomatism in Cratons." *Earth and Planetary Science Letters* 491 (June): 148–59. <https://doi.org/10.1016/j.epsl.2018.03.018>.
- Eilon, Zachary, Karen M. Fischer, and Colleen A. Dalton. 2018. "An Adaptive Bayesian Inversion for Upper-Mantle Structure Using Surface Waves and Scattered Body Waves." *Geophysical Journal International* 214 (1): 232–53. <https://doi.org/10.1093/gji/ggy137>.

- England, Richard W., and Jörg Ebbing. 2012. “Crustal Structure of Central Norway and Sweden from Integrated Modelling of Teleseismic Receiver Functions and the Gravity Anomaly: Crustal Structure of Central Norway and Sweden.” *Geophysical Journal International* 191 (1): 1–11. <https://doi.org/10.1111/j.1365-246X.2012.05607.x>.
- Finnerty, A. A., and F. R. Boyd. 1987. *Thermobarometry for Garnet Peridotites: Basis for the Determination of Thermal and Compositional Structure of the Upper Mantle. Mantle Xenoliths*.
- Fischer, Karen M., Catherine A. Rychert, Colleen A. Dalton, Meghan S. Miller, Caroline Beghein, and Derek L. Schutt. 2020. “A Comparison of Oceanic and Continental Mantle Lithosphere.” *Physics of the Earth and Planetary Interiors* 309 (December): 106600. <https://doi.org/10.1016/j.pepi.2020.106600>.
- Ford, Heather A., Karen M. Fischer, David L. Abt, Catherine A. Rychert, and Linda T. Elkins-Tanton. 2010. “The Lithosphere–Asthenosphere Boundary and Cratonic Lithospheric Layering beneath Australia from Sp Wave Imaging.” *Earth and Planetary Science Letters* 300 (3–4): 299–310. <https://doi.org/10.1016/j.epsl.2010.10.007>.
- Fu, Hui-Ying, and Zhong-Hai Li. 2024. “Roles of Continental Mid-Lithosphere Discontinuity in the Craton Instability Under Variable Tectonic Regimes.” *Journal of Geophysical Research: Solid Earth* 129 (1): e2023JB028022. <https://doi.org/10.1029/2023JB028022>.

Gama, Isabella, Karen M. Fischer, Colleen A. Dalton, and Zachary Eilon. 2022.

“Variations in Lithospheric Thickness Across the Denali Fault and in Northern Alaska.” *Geophysical Research Letters* 49 (24): e2022GL101256.

<https://doi.org/10.1029/2022GL101256>.

Garber, Joshua M., Satish Maurya, Jean-Alexis Hernandez, et al. 2018.

“Multidisciplinary Constraints on the Abundance of Diamond and Eclogite in the Cratonic Lithosphere.” *Geochemistry, Geophysics, Geosystems* 19 (7): 2062–86.

<https://doi.org/10.1029/2018GC007534>.

GEOFON Data Centre. 1993. “GEOFON Seismic Network.” With GEOFON Data

Centre, [geofon@gfz.de](mailto:geofon@gfz.de), and Deutsches GeoForschungsZentrum GFZ. GFZ Data Services. ...Mseed,XML, Approximately 80 active stations; greater than 440MB/day. <https://doi.org/10.14470/TR560404>.

Geological Survey Of Estonia (EGT). 1996. “Estonian National Seismic Network

(EESN).” With Geological Survey Of Estonia, GFZ Helmholtz-Zentrum Für Geoforschung, GEOFON Data Centre, and [Geofon@Gfz.De](mailto:Geofon@Gfz.De). GFZ Data Services. ...Mseed,XML, on avg. 300GB per Year.

<https://doi.org/10.14470/RR425351>.

GEUS Geological Survey of Denmark and Greenland. 1976. “Danish Seismological

Network.” International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/NW3X-DF02>.

- Goev, A. G., I. A. Sanina, S. I. Oreshin, R. A. Reznichenko, S. A. Tarasov, and A. V. Fedorov. 2021. “Receiver-Function Velocity Structure of the Lithosphere beneath the Khibiny and Lovozero Massifs, Northeastern Baltic Shield.” *Izvestiya, Physics of the Solid Earth* 57 (5): 605–13. <https://doi.org/10.1134/S1069351321050062>.
- Golos, E. M., B. Brunsvik, Z. Eilon, K. M. Fischer, J. Byrnes, and J. Gaherty. 2024. “A New View of Shear Wavespeed and the Lithosphere-Asthenosphere Boundary in the Southwestern United States.” *Journal of Geophysical Research: Solid Earth* 129 (8): e2024JB029220. <https://doi.org/10.1029/2024JB029220>.
- Gorbatshev, Roland, and Svetlana Bogdanova. 1993. “Frontiers in the Baltic Shield.” *Precambrian Research* 64 (1–4): 3–21. [https://doi.org/10.1016/0301-9268\(93\)90066-B](https://doi.org/10.1016/0301-9268(93)90066-B).
- Gradmann, S., J. Ebbing, and J. Fullea. 2013. “Integrated Geophysical Modelling of a Lateral Transition Zone in the Lithospheric Mantle under Norway and Sweden.” *Geophysical Journal International* 194 (3): 1358–73. <https://doi.org/10.1093/gji/ggt213>.
- Griffin, Brendon J., Juha Rissanen, Gregory D. Pooley, Dearn C. Lee, Ian Macdonald, and Peter D. Kinny. 1995. “A New Diamondiferous Eciogite-Bearing Kimberlitic Occurrence from Finland.” Paper presented at International Kimberlite Conference. *International Kimberlite Conference Extended Abstracts: 1995*, September 19. <https://doi.org/10.29173/ikc1832>.

- Griffin, W. L., S. Y. O'Reilly, J. C. Afonso, and G. C. Begg. 2009. "The Composition and Evolution of Lithospheric Mantle: A Re-Evaluation and Its Tectonic Implications." *Journal of Petrology* 50 (7): 1185–204.  
<https://doi.org/10.1093/petrology/egn033>.
- Grund, Michael, and Joachim R. R. Ritter. 2020. "Shear-Wave Splitting beneath Fennoscandia — Evidence for Dipping Structures and Laterally Varying Multilayer Anisotropy." *Geophysical Journal International* 223 (3): 1525–47.  
<https://doi.org/10.1093/gji/ggaa388>.
- Gung, Yuancheng, Mark Panning, and Barbara Romanowicz. 2003. "Global Anisotropy and the Thickness of Continents." *Nature* 422 (6933): 707–11.  
<https://doi.org/10.1038/nature01559>.
- Hamza, V. M., and F. P. Vieira. 2012. "Global Distribution of the Lithosphere-Asthenosphere Boundary: A New Look." *Solid Earth* 3 (2): 199–212.  
<https://doi.org/10.5194/se-3-199-2012>.
- Hansen, Ralf T. J., Michael G. Bostock, and Nikolas I. Christensen. 2012. "Nature of the Low Velocity Zone in Cascadia from Receiver Function Waveform Inversion." *Earth and Planetary Science Letters* 337–338 (July): 25–38.  
<https://doi.org/10.1016/j.epsl.2012.05.031>.
- Hansen, Steven M., Ken Dueker, and Brandon Schmandt. 2015. "Thermal Classification of Lithospheric Discontinuities beneath USArray." *Earth and Planetary Science Letters* 431 (December): 36–47. <https://doi.org/10.1016/j.epsl.2015.09.009>.

- Hastings, W. K. 1970. “Monte Carlo Sampling Methods Using Markov Chains and Their Applications.” *Biometrika* 57 (1): 97–109. <https://doi.org/10.1093/biomet/57.1.97>.
- Hirsch, Aaron C., Colleen A. Dalton, and Jeroen Ritsema. 2015. “Constraints on Shear Velocity in the Cratonic Upper Mantle from R Ayleigh Wave Phase Velocity.” *Geochemistry, Geophysics, Geosystems* 16 (11): 3982–4005. <https://doi.org/10.1002/2015GC006066>.
- Hirth, Greg, Rob L. Evans, and Alan D. Chave. 2000. “Comparison of Continental and Oceanic Mantle Electrical Conductivity: Is the Archean Lithosphere Dry?” *Geochemistry, Geophysics, Geosystems* 1 (12): 2000GC000048. <https://doi.org/10.1029/2000GC000048>.
- Holland, T. J. B., and R. Powell. 1998. “An Internally Consistent Thermodynamic Data Set for Phases of Petrological Interest.” *Journal of Metamorphic Geology* 16 (3): 309–43. <https://doi.org/10.1111/j.1525-1314.1998.00140.x>.
- Hua, J., K. M. Fischer, M. Wu, and N. A. Blom. 2020. “New Approaches to Multifrequency *Sp* Stacking Tested in the Anatolian Region.” *Journal of Geophysical Research: Solid Earth* 125 (11): e2020JB020313. <https://doi.org/10.1029/2020JB020313>.
- Hua, Junlin, Karen M. Fischer, Thorsten W. Becker, Esteban Gazel, and Greg Hirth. 2023. “Asthenospheric Low-Velocity Zone Consistent with Globally Prevalent Partial Melting.” *Nature Geoscience* 16 (2): 175–81. <https://doi.org/10.1038/s41561-022-01116-9>.

- Huang, Yiran, Colleen A. Dalton, and Anant Hariharan. 2025. “A New Approach to Constrain Crustal Vp/Vs From Rayleigh Wave Phase Velocity and Local Amplification: Application to the Western US.” *Geophysical Research Letters* 52 (6): e2024GL111980. <https://doi.org/10.1029/2024GL111980>.
- Jackson, Ian, and Ulrich H. Faul. 2010. “Grainsize-Sensitive Viscoelastic Relaxation in Olivine: Towards a Robust Laboratory-Based Model for Seismological Application.” *Physics of the Earth and Planetary Interiors* 183 (1–2): 151–63. <https://doi.org/10.1016/j.pepi.2010.09.005>.
- Janik, Tomasz, Elena Kozlovskaya, and Jukka Yliniemi. 2007. “Crust-mantle Boundary in the Central Fennoscandian Shield: Constraints from Wide-angle *P* and *S* Wave Velocity Models and New Results of Reflection Profiling in Finland.” *Journal of Geophysical Research: Solid Earth* 112 (B4): 2006JB004681. <https://doi.org/10.1029/2006JB004681>.
- Jin, Ge, and James B. Gaherty. 2015. “Surface Wave Phase-Velocity Tomography Based on Multichannel Cross-Correlation.” *Geophysical Journal International* 201 (3): 1383–98. <https://doi.org/10.1093/gji/ggv079>.
- Karato, Shun-ichiro. 2012. “On the Origin of the Asthenosphere.” *Earth and Planetary Science Letters* 321–322 (March): 95–103. <https://doi.org/10.1016/j.epsl.2012.01.001>.
- Karato, Shun-ichiro. 2025. “Causality and Its Implications for the Interpretation of Seismological Observations on the Upper Mantle.” *Journal of Geophysical*



*Research: Solid Earth* 130 (9): e2024JB030639.

<https://doi.org/10.1029/2024JB030639>.

Keith, Colum M., and Stuart Crampin. 1977. “Seismic Body Waves in Anisotropic Media: Synthetic Seismograms.” *Geophysical Journal International* 49 (1): 225–43. <https://doi.org/10.1111/j.1365-246X.1977.tb03710.x>.

Kennett, B. L. N. 1991. “The Removal of Free Surface Interactions From Three-Component Seismograms.” *Geophysical Journal International* 104 (1): 153–54. <https://doi.org/10.1111/j.1365-246X.1991.tb02501.x>.

Kind, R., F. Sodoudi, X. Yuan, et al. 2013. “Scandinavia: A Former Tibet?” *Geochemistry, Geophysics, Geosystems* 14 (10): 4479–87. <https://doi.org/10.1002/ggge.20251>.

Kind, Rainer, Walter D. Mooney, and Xiaohui Yuan. 2020. “New Insights into the Structural Elements of the Upper Mantle beneath the Contiguous United States from *S*-to-*P* Converted Seismic Waves.” *Geophysical Journal International* 222 (1): 646–59. <https://doi.org/10.1093/gji/ggaa203>.

Köhler, Andreas, Valérie Maupin, and Niels Balling. 2015. “Surface Wave Tomography across the Sorgenfrei–Tornquist Zone, SW Scandinavia, Using Ambient Noise and Earthquake Data.” *Geophysical Journal International* 203 (1): 284–311. <https://doi.org/10.1093/gji/ggv297>.

Köhler, Andreas, Christian Weidle, and Valérie Maupin. 2012. “Crustal and Uppermost Mantle Structure of Southern Norway: Results from Surface Wave Analysis of

Ambient Seismic Noise and Earthquake Data: *Surface Wave Tomography of Southern Norway*.” *Geophysical Journal International*, October, no-no.

<https://doi.org/10.1111/j.1365-246X.2012.05698.x>.

Kolstrup, Marianne L., Shu-Huei Hung, and Valerie Maupin. 2015. “Multiscale, Finite-Frequency P and S Tomography of the Upper Mantle in the Southwestern Fennoscandian Shield.” *Geophysical Journal International* 202 (1): 190–218.

<https://doi.org/10.1093/gji/ggv130>.

Kolstrup, Marianne L., and Valérie Maupin. 2013. “A Proterozoic Boundary in Southern Norway Revealed by Joint-Inversion of P-Receiver Functions and Surface Waves.” *Precambrian Research* 238 (November): 186–98.

<https://doi.org/10.1016/j.precamres.2013.10.004>.

Kozlovskaya, E. 2007. “Seismic Network XK:LAPNET/POLENET Seismic Temporary Array (RESIF-SISMOB).” With Institut des Sciences de la Terre, sismo-help@resif.fr, Sodankyla Geophysical Observatory (FINLAND), et al. RESIF - Réseau Sismologique et géodésique Français, January 1.

Application/vnd.fdsn.mseed,application/vnd.fdsn.station+xml+xml.

<https://doi.org/10.15778/RESIF.XK2007>.

Kozlovskaya, Elena, Grigoriy Kosarev, Igor Aleshin, Oksana Riznichenko, and Irina Sanina. 2008. “Structure and Composition of the Crust and Upper Mantle of the Archean-Proterozoic Boundary in the Fennoscandian Shield Obtained by Joint Inversion of Receiver Function and Surface Wave Phase Velocity of Recording of

- the SVEKALAPKO Array.” *Geophysical Journal International* 175 (1): 135–52.  
<https://doi.org/10.1111/j.1365-246X.2008.03876.x>.
- Krueger, Hannah E., Isabella Gama, and Karen M. Fischer. 2021. “Global Patterns in Cratonic Mid-Lithospheric Discontinuities From Sp Receiver Functions.” *Geochemistry, Geophysics, Geosystems* 22 (6): e2021GC009819.  
<https://doi.org/10.1029/2021GC009819>.
- Kukkonen, I. T., and P. Peltonen. 1999. “Xenolith-Controlled Geotherm for the Central Fennoscandian Shield: Implications for Lithosphere–Asthenosphere Relations.” *Tectonophysics* 304 (4): 301–15. [https://doi.org/10.1016/S0040-1951\(99\)00031-1](https://doi.org/10.1016/S0040-1951(99)00031-1).
- Lahtinen, Raimo, Adam A. Garde, and Victor A. Melezhik. 2008. “Paleoproterozoic Evolution of Fennoscandia and Greenland.” *Episodes* 31 (1): 20–28.  
<https://doi.org/10.18814/epiiugs/2008/v31i1/004>.
- Lars Ottemoller, and Conrad Lindholm. 2013. “NEONOR2 Temporary Seismic Deployment.” International Federation of Digital Seismograph Networks, December 31. SEED data, 1478250 MB. <https://doi.org/10.7914/CTAX-CA85>.
- Lee, Cin-Ty A., Peter Luffi, and Emily J. Chin. 2011. “Building and Destroying Continental Mantle.” *Annual Review of Earth and Planetary Sciences* 39 (1): 59–90. <https://doi.org/10.1146/annurev-earth-040610-133505>.
- Legre, Jean-Joel, and Tolulope Olugboji. 2025. “The Origins of Africa’s Upper Mantle Discontinuities.” *Geochemistry, Geophysics, Geosystems* 26 (7): e2025GC012315. <https://doi.org/10.1029/2025GC012315>.

- Lehtonen, M. L., H. E. O'Brien, P. Peltonen, B. S. Johanson, and L. K. Pakkanen. 2004. "Layered Mantle at the Karelian Craton Margin: P–T of Mantle Xenocrysts and Xenoliths from the Kaavi–Kuopio Kimberlites, Finland." *Lithos* 77 (1–4): 593–608. <https://doi.org/10.1016/j.lithos.2004.04.026>.
- Lei, Wenjie, Youyi Ruan, Ebru Bozdağ, et al. 2020. "Global Adjoint Tomography—Model GLAD-M25." *Geophysical Journal International* 223 (1): 1–21. <https://doi.org/10.1093/gji/ggaa253>.
- Ligorria, Juan Pablo, and Charles J. Ammon. 1999. "Iterative Deconvolution and Receiver-Function Estimation." *Bulletin of the Seismological Society of America* 89 (5): 1395–400. <https://doi.org/10.1785/BSSA0890051395>.
- Lin, Fan-Chi, Michael H. Ritzwoller, and Roel Snieder. 2009. "Eikonal Tomography: Surface Wave Tomography by Phase Front Tracking across a Regional Broad-Band Seismic Array." *Geophysical Journal International* 177 (3): 1091–110. <https://doi.org/10.1111/j.1365-246X.2009.04105.x>.
- Liu, Liang, Jason P. Morgan, Yigang Xu, and Martin Menzies. 2018. "Craton Destruction 1: Cratonic Keel Delamination Along a Weak Midlithospheric Discontinuity Layer." *Journal of Geophysical Research: Solid Earth* 123 (11). <https://doi.org/10.1029/2017JB015372>.
- Liu, Tianze, Emily J. Chin, and Peter Shearer. 2023. "Strong Physical Contrasts Across Two Mid-Lithosphere Discontinuities Beneath the Northwestern United States:

- Evidence for Cratonic Mantle Metasomatism.” *AGU Advances* 4 (6): e2023AV001014. <https://doi.org/10.1029/2023AV001014>.
- Liu, Tianze, and Peter M. Shearer. 2021. “Complicated Lithospheric Structure Beneath the Contiguous US Revealed by Teleseismic S-Reflections.” *Journal of Geophysical Research: Solid Earth* 126 (5): e2020JB021624. <https://doi.org/10.1029/2020JB021624>.
- MacGregor, I. D. 1974. “The System MgO-SiO<sub>2</sub>-Al<sub>2</sub>O<sub>3</sub>: Solubility of Al<sub>2</sub>O<sub>3</sub> in Enstatite for Spinel and Garnet Peridotite Compositions.” *Am. Mineral.* 59: 110–19.
- Makushkina, Anna, Benoit Tauzin, Meghan S. Miller, Hrvoje Tkalčić, and Hans Thybo. 2025. “Opening of the North Atlantic Ocean and the Rise of Scandinavian Mountains.” *Geology* 53 (1): 8–12. <https://doi.org/10.1130/G52735.1>.
- Mauerberger, Alexandra, Valérie Maupin, Ólafur Gudmundsson, and Frederik Tilmann. 2020. “Anomalous Azimuthal Variations with 360° Periodicity of Rayleigh Phase Velocities Observed in Scandinavia.” *Geophysical Journal International* 224 (3): 1684–704. <https://doi.org/10.1093/gji/ggaa553>.
- Mauerberger, Alexandra, Hamzeh Sadeghisorkhani, Valérie Maupin, Ólafur Gudmundsson, and Frederik Tilmann. 2022. “A Shear-Wave Velocity Model for the Scandinavian Lithosphere from Rayleigh Waves and Ambient Noise - Implications for the Origin of the Topography of the Scandes Mountain Range.” *Tectonophysics* 838 (September): 229507. <https://doi.org/10.1016/j.tecto.2022.229507>.

- Maupin, V., A. Agostini, I. Artemieva, et al. 2013. “The Deep Structure of the Scandes and Its Relation to Tectonic History and Present-Day Topography.” *Tectonophysics* 602 (August): 15–37. <https://doi.org/10.1016/j.tecto.2013.03.010>.
- Maupin, Valérie. 2011. “Upper-Mantle Structure in Southern Norway from Beamforming of Rayleigh Wave Data Presenting Multipathing: Upper Mantle in Southern Norway.” *Geophysical Journal International* 185 (2): 985–1002. <https://doi.org/10.1111/j.1365-246X.2011.04989.x>.
- Maupin, Valérie, Alexandra Mauerberger, and Frederik Tilmann. 2022. “The Radial Anisotropy of the Continental Lithosphere From Analysis of Love and Rayleigh Wave Phase Velocities in Fennoscandia.” *Journal of Geophysical Research: Solid Earth* 127 (10): e2022JB024445. <https://doi.org/10.1029/2022JB024445>.
- Medhus, A. B., N. Balling, B. H. Jacobsen, et al. 2012. “Upper-Mantle Structure beneath the Southern Scandes Mountains and the Northern Tornquist Zone Revealed by P-Wave Traveltime Tomography: Upper Mantle beneath Southern Scandinavia.” *Geophysical Journal International* 189 (3): 1315–34. <https://doi.org/10.1111/j.1365-246X.2012.05449.x>.
- Miller, Meghan S., and David W. Eaton. 2010. “Formation of Cratonic Mantle Keels by Arc Accretion: Evidence from S Receiver Functions.” *Geophysical Research Letters* 37 (18): 2010GL044366. <https://doi.org/10.1029/2010GL044366>.

- Nimis, Paolo, and Herman Grütter. 2010. "Internally Consistent Geothermometers for Garnet Peridotites and Pyroxenites." *Contributions to Mineralogy and Petrology* 159 (3): 411–27. <https://doi.org/10.1007/s00410-009-0455-9>.
- Norsar. 1971. "NORSAR Station Network." NORSAR. <https://doi.org/10.21348/D.NO.0001>.
- O'Brien, H. E., P. Peltonen, and H. Vartiainen. 2005. "Chapter 14 Kimberlites, Carbonatites, and Alkaline Rocks." In *Developments in Precambrian Geology*, vol. 14. Elsevier. [https://doi.org/10.1016/S0166-2635\(05\)80015-5](https://doi.org/10.1016/S0166-2635(05)80015-5).
- Olsson, Sverker, Roland G. Roberts, and Reynir Böðvarsson. 2007. "Analysis of Waves Converted from S to P in the Upper Mantle beneath the Baltic Shield." *Earth and Planetary Science Letters* 257 (1–2): 37–46. <https://doi.org/10.1016/j.epsl.2007.02.017>.
- Osmundsen, P. T., and J. Ebbing. 2008. "Styles of Extension Offshore mid-Norway and Implications for Mechanisms of Crustal Thinning at Passive Margins." *Tectonics* 27 (6): 2007TC002242. <https://doi.org/10.1029/2007TC002242>.
- Pedersen, H. A., E. Debayle, and V. Maupin. 2013. "Strong Lateral Variations of Lithospheric Mantle beneath Cratons – Example from the Baltic Shield." *Earth and Planetary Science Letters* 383 (December): 164–72. <https://doi.org/10.1016/j.epsl.2013.09.024>.

- Pedersen, H. A., S. Fishwick, and D. B. Snyder. 2009. "A Comparison of Cratonic Roots through Consistent Analysis of Seismic Surface Waves." *Lithos* 109 (1–2): 81–95. <https://doi.org/10.1016/j.lithos.2008.09.016>.
- Peltonen, P, H. Huhma, M. Tyni, and N. Shimizu. 1998. "Garnet Peridotite Xenoliths from Kimberlites of Finland: Nature of the Lithospheric Mantle at Archaean Craton - Palaeoproterozoic Mobile Belt Transition." Paper presented at International Kimberlite Conference. *International Kimberlite Conference Extended Abstracts: 1998*, April 17. <https://doi.org/10.29173/ikc2844>.
- Peltonen, P, K. A. Kinnunen, and H. Huhma. 2002. "Petrology of Two Diamondiferous Eclogite Xenoliths from the Lahtojoki Kimberlite Pipe, Eastern Finland." *Lithos* 63 (3–4): 151–64. [https://doi.org/10.1016/S0024-4937\(02\)00119-6](https://doi.org/10.1016/S0024-4937(02)00119-6).
- Peltonen, Petri, and Gerhard Brügmann. 2006. "Origin of Layered Continental Mantle (Karelian Craton, Finland): Geochemical and Re–Os Isotope Constraints." *Lithos* 89 (3–4): 405–23. <https://doi.org/10.1016/j.lithos.2005.12.013>.
- Perchuć, E., and H. Thybo. 1996. "A New Model of Upper Mantle P-Wave Velocity below the Baltic Shield: Indication of Partial Melt in the 95 to 160 Km Depth Range." *Tectonophysics* 253 (3–4): 227–45. [https://doi.org/10.1016/0040-1951\(95\)00057-7](https://doi.org/10.1016/0040-1951(95)00057-7).
- Peslier, Anne H., Maria Schönbächler, Henner Busemann, and Shun-Ichiro Karato. 2017. "Water in the Earth's Interior: Distribution and Origin." *Space Science Reviews* 212 (1–2): 743–810. <https://doi.org/10.1007/s11214-017-0387-z>.



- Rader, Erika, Erica Emry, Nicholas Schmerr, et al. 2015. "Characterization and Petrological Constraints of the Midlithospheric Discontinuity." *Geochemistry, Geophysics, Geosystems* 16 (10): 3484–504.  
<https://doi.org/10.1002/2015GC005943>.
- Ramirez, Florence D. C., Kate Selway, Clinton P. Conrad, Valerie Maupin, and Maxim Smirnov. 2024. "Lateral and Radial Viscosity Structure beneath Fennoscandia Inferred from Seismic and Magnetotelluric Observations." *Physics of the Earth and Planetary Interiors* 351 (June): 107178.  
<https://doi.org/10.1016/j.pepi.2024.107178>.
- Richard England. 2013a. "SCANLIPS3D." International Federation of Digital Seismograph Networks. SEED data, 350000 MB.  
[https://doi.org/10.7914/SN/ZR\\_2013](https://doi.org/10.7914/SN/ZR_2013).
- Richard England. 2013b. "SCANLIPS3D." International Federation of Digital Seismograph Networks. SEED data, 350000 MB.  
[https://doi.org/10.7914/SN/ZR\\_2013](https://doi.org/10.7914/SN/ZR_2013).
- Roberts, David. 2003. "The Scandinavian Caledonides: Event Chronology, Palaeogeographic Settings and Likely Modern Analogues." *Tectonophysics* 365 (1–4): 283–99. [https://doi.org/10.1016/S0040-1951\(03\)00026-X](https://doi.org/10.1016/S0040-1951(03)00026-X).
- Rychert, Catherine A., and Peter M. Shearer. 2009. "A Global View of the Lithosphere-Asthenosphere Boundary." *Science* 324 (5926): 495–98.  
<https://doi.org/10.1126/science.1169754>.

- Saha, Sriparna, and Rajdeep Dasgupta. 2019. "Phase Relations of a Depleted Peridotite Fluxed by a CO<sub>2</sub>-H<sub>2</sub>O Fluid—Implications for the Stability of Partial Melts Versus Volatile-Bearing Mineral Phases in the Cratonic Mantle." *Journal of Geophysical Research: Solid Earth* 124 (10): 10089–106.  
<https://doi.org/10.1029/2019JB017653>.
- Saha, Sriparna, Ye Peng, Rajdeep Dasgupta, Mainak Mookherjee, and Karen M. Fischer. 2021. "Assessing the Presence of Volatile-Bearing Mineral Phases in the Cratonic Mantle as a Possible Cause of Mid-Lithospheric Discontinuities." *Earth and Planetary Science Letters* 553 (January): 116602.  
<https://doi.org/10.1016/j.epsl.2020.116602>.
- Savage, Brian, and Paul G. Silver. 2008. "Evidence for a Compositional Boundary within the Lithospheric Mantle beneath the Kalahari Craton from S Receiver Functions." *Earth and Planetary Science Letters* 272 (3–4): 600–609.  
<https://doi.org/10.1016/j.epsl.2008.05.026>.
- Schulte-Pelkum, Vera, and Kevin H. Mahan. 2014. "A Method for Mapping Crustal Deformation and Anisotropy with Receiver Functions and First Results from USArray." *Earth and Planetary Science Letters* 402 (September): 221–33.  
<https://doi.org/10.1016/j.epsl.2014.01.050>.
- Selway, Kate, Heather Ford, and Peter Kelemen. 2015. "The Seismic Mid-Lithosphere Discontinuity." *Earth and Planetary Science Letters* 414 (March): 45–57.  
<https://doi.org/10.1016/j.epsl.2014.12.029>.

- Silvennoinen, H., E. Kozlovskaya, and E. Kissling. 2015. "POLENET/LAPNET Teleseismic P-Wave Traveltime Tomography Model of the Upper Mantle beneath Northern Fennoscandia." Preprint, *Seismology*, September 10.  
<https://doi.org/10.5194/sed-7-2527-2015>.
- Silvennoinen, Hanna, Elena Kozlovskaya, Eduard Kissling, and Grigoriy Kosarev. 2014. "A New Moho Boundary Map for the Northern Fennoscandian Shield Based on Combined Controlled-Source Seismic and Receiver Function Data." *GeoResJ* 1–2 (March): 19–32. <https://doi.org/10.1016/j.grj.2014.03.001>.
- Sodankylä Geophysical Observatory / University of Oulu (Finland). 1980. "Northern Finland Seismological Network." With University of Oulu, Deutsches GeoForschungsZentrum GFZ, GEOFON Data Centre, and [geofon@gfz.de](mailto:geofon@gfz.de). GFZ Data Services. ..Mseed,XML, ~300G. <https://doi.org/10.14470/SA879454>.
- Stachel, T., and R. W. Luth. 2015. "Diamond Formation — Where, When and How?" *Lithos* 220–223 (April): 200–220. <https://doi.org/10.1016/j.lithos.2015.01.028>.
- Steinberger, Bernhard, and Thorsten W. Becker. 2018. "A Comparison of Lithospheric Thickness Models." *Tectonophysics* 746 (October): 325–38.  
<https://doi.org/10.1016/j.tecto.2016.08.001>.
- Stixrude, Lars, and Carolina Lithgow-Bertelloni. 2005. "Thermodynamics of Mantle Minerals - I. Physical Properties." *Geophysical Journal International* 162 (2): 610–32. <https://doi.org/10.1111/j.1365-246X.2005.02642.x>.

- Stixrude, Lars, and Carolina Lithgow-Bertelloni. 2011. “Thermodynamics of Mantle Minerals - II. Phase Equilibria.” *Geophysical Journal International* 184 (3): 1180–213. <https://doi.org/10.1111/j.1365-246X.2010.04890.x>.
- Stixrude, Lars, and Carolina Lithgow-Bertelloni. 2024. “Thermodynamics of Mantle Minerals – III: The Role of Iron.” *Geophysical Journal International* 237 (3): 1699–733. <https://doi.org/10.1093/gji/ggae126>.
- Stratford, Wanda, Hans Thybo, Jan Inge Faleide, Odleiv Olesen, and Ari Tryggvason. 2009. “New Moho Map for Onshore Southern Norway.” *Geophysical Journal International* 178 (3): 1755–65. <https://doi.org/10.1111/j.1365-246X.2009.04240.x>.
- Sudholz, Z. J., P. Zhang, C. M. Eakin, et al. 2024. “Ancient Craton-Wide Mid-Lithosphere Discontinuity Controlled by Pargasite Channels.” *Geophysical Research Letters* 51 (9): e2024GL108433. <https://doi.org/10.1029/2024GL108433>.
- Sun, Weijia, and B. L. N. Kennett. 2017. “Mid-lithosphere Discontinuities beneath the Western and Central North China Craton.” *Geophysical Research Letters* 44 (3): 1302–10. <https://doi.org/10.1002/2016GL071840>.
- Taylor, W. R. 1998a. “An Experimental Test of Some Geothermometer and Geobarometer Formulations for Upper Mantle Peridotites with Application to the Thermobarometry of Fertile Lherzolite and Garnet Websterite.” *Neues Jahrbuch Für*

*Mineralogie - Abhandlungen* 172 (2–3): 381–408.

<https://doi.org/10.1127/njma/172/1998/381>.

Taylor, W. R. 1998b. “An Experimental Test of Some Geothermometer and Geobarometer Formulations for Upper Mantle Peridotites with Application to the Thermobarometry of Fertile Lherzolite and Garnet Websterite.” *Neues Jahrbuch Für Mineralogie - Abhandlungen* 172 (2–3): 381–408.

<https://doi.org/10.1127/njma/172/1998/381>.

Tesauro, Magdala, Mikhail K. Kaban, and Sierd A. P. L. Cloetingh. 2008. “EuCRUST-07: A New Reference Model for the European Crust.” *Geophysical Research Letters* 35 (5): 2007GL032244. <https://doi.org/10.1029/2007GL032244>.

Thybo, H., and E. Perchuc. 1997. “The Seismic 8° Discontinuity and Partial Melting in Continental Mantle.” *Science* 275 (5306): 1626–29.

<https://doi.org/10.1126/science.275.5306.1626>.

Thybo, Hans. 2006. “The Heterogeneous Upper Mantle Low Velocity Zone.” *Tectonophysics* 416 (1–4): 53–79. <https://doi.org/10.1016/j.tecto.2005.11.021>.

Thybo, Hans, Niels Balling, Valerie Maupin, Joachim Ritter, and Frederik Tilmann. 2012. “ScanArray Core (1G 2012-2017).” With Deutsches GeoForschungsZentrum GFZ, GEOFON Data Centre, and [geofon@gfz.de](mailto:geofon@gfz.de). The ScanArray consortium. ..Mseed,XML, 1200GB. <https://doi.org/10.14470/6T569239>.

- Thybo, Hans, Nevra Bulut, Michael Grund, et al. 2021. “ScanArray—A Broadband Seismological Experiment in the Baltic Shield.” *Seismological Research Letters* 92 (5): 2811–23. <https://doi.org/10.1785/0220210015>.
- Tiira, Timo, Tomasz Janik, Tymon Skrzynik, et al. 2020. “Full-Scale Crustal Interpretation of Kokkola–Kymi (KOKKY) Seismic Profile, Fennoscandian Shield.” *Pure and Applied Geophysics* 177 (8): 3775–95. <https://doi.org/10.1007/s00024-020-02459-3>.
- Till, Christy B., Timothy L. Grove, and Anthony C. Withers. 2012. “The Beginnings of Hydrous Mantle Wedge Melting.” *Contributions to Mineralogy and Petrology* 163 (4): 669–88. <https://doi.org/10.1007/s00410-011-0692-6>.
- Torsvik, Trond H., and L. Robin M. Cocks. 2025. “Norway in Space and Time: A Centennial Cavalcade.” *Norwegian Journal of Geology* 85 (73–86).
- Tozer, B., D. T. Sandwell, W. H. F. Smith, C. Olson, J. R. Beale, and P. Wessel. 2019. “Global Bathymetry and Topography at 15 Arc Sec: SRTM15+.” *Earth and Space Science* 6 (10): 1847–64. <https://doi.org/10.1029/2019EA000658>.
- University of Bergen. 1982. “University of Bergen Seismic Network.” International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/SN/NS>.
- Veikkolainen, Toni, Ilmo T. Kukkonen, and Timo Tiira. 2017. “Heat Flow, Seismic Cut-off Depth and Thermal Modeling of the Fennoscandian Shield.” *Geophysical Journal International* 211 (3): 1414–27. <https://doi.org/10.1093/gji/ggx373>.

- Viljoen, K. S., René Dobbe, Braam Smit, Emilie Thomassot, and Pierre Cartigny. 2004. "Petrology and Geochemistry of a Diamondiferous Lherzolite from the Premier Diamond Mine, South Africa." *Lithos* 77 (1–4): 539–52.  
<https://doi.org/10.1016/j.lithos.2004.03.023>.
- Vinnik, L., S. Kiselev, M. Weber, S. Oreshin, and L. Makeyeva. 2012. "Frozen and Active Seismic Anisotropy beneath Southern Africa." *Geophysical Research Letters* 39 (8): 2012GL051326. <https://doi.org/10.1029/2012GL051326>.
- Vinnik, Lev, Elena Kozlovskaya, Sergey Oreshin, Grigoriy Kosarev, Katerina Piiponen, and Hanna Silvennoinen. 2016. "The Lithosphere, LAB, LVZ and Lehmann Discontinuity under Central Fennoscandia from Receiver Functions." *Tectonophysics* 667 (January): 189–98.  
<https://doi.org/10.1016/j.tecto.2015.11.024>.
- Vinnik, Lev, Sergey Oreshin, Larissa Makeyeva, et al. 2014. "Anisotropic Lithosphere under the Fennoscandian Shield from P Receiver Functions and SKS Waveforms of the POLENET/LAPNET Array." *Tectonophysics* 628 (July): 45–54.  
<https://doi.org/10.1016/j.tecto.2014.04.024>.
- Wawerzinek, B., J. R. R. Ritter, and C. Roy. 2013. "New Constraints on the 3D Shear Wave Velocity Structure of the Upper Mantle underneath Southern Scandinavia Revealed from Non-Linear Tomography." *Tectonophysics* 602 (August): 38–54.  
<https://doi.org/10.1016/j.tecto.2012.12.033>.

- Weidle, C., V. Maupin, J. Ritter, et al. 2010. “MAGNUS--A Seismological Broadband Experiment to Resolve Crustal and Upper Mantle Structure beneath the Southern Scandes Mountains in Norway.” *Seismological Research Letters* 81 (1): 76–84. <https://doi.org/10.1785/gssrl.81.1.76>.
- Wilde-Piórko, Monika, Wolfram H. Geissler, Jaroslava Plomerová, et al. 2009. “PASSEQ 2006-2008: Passive Seismic Experiment in Trans-European Suture Zone.” With Deutsches GeoForschungsZentrum GFZ, GEOFON Data Centre, [Geofon@Gfz.De](mailto:Geofon@Gfz.De), et al. GFZ Data Services, January 1. ..Mseed,XML, 2125GB. <https://doi.org/10.14470/2R383989>.
- Wirth, Erin A., and Maureen D. Long. 2014. “A Contrast in Anisotropy across Mid-Lithospheric Discontinuities beneath the Central United States—A Relic of Craton Formation.” *Geology* 42 (10): 851–54. <https://doi.org/10.1130/G35804.1>.
- Wittlinger, Gérard, and Véronique Farra. 2007. “Converted Waves Reveal a Thick and Layered Tectosphere beneath the Kalahari Super-Craton.” *Earth and Planetary Science Letters* 254 (3–4): 404–15. <https://doi.org/10.1016/j.epsl.2006.11.048>.
- Wölbern, Ingo, Georg Rümpker, Klemens Link, and Forough Sodoudi. 2012. “Melt Infiltration of the Lower Lithosphere beneath the Tanzania Craton and the Albertine Rift Inferred from S Receiver Functions.” *Geochemistry, Geophysics, Geosystems* 13 (8): 2012GC004167. <https://doi.org/10.1029/2012GC004167>.
- Yliniemi, Jukka, Elena Kozlovskaya, Sven-Erik Hjelt, Kari Komminaho, and Anton Ushakov. 2004. “Structure of the Crust and Uppermost Mantle beneath Southern

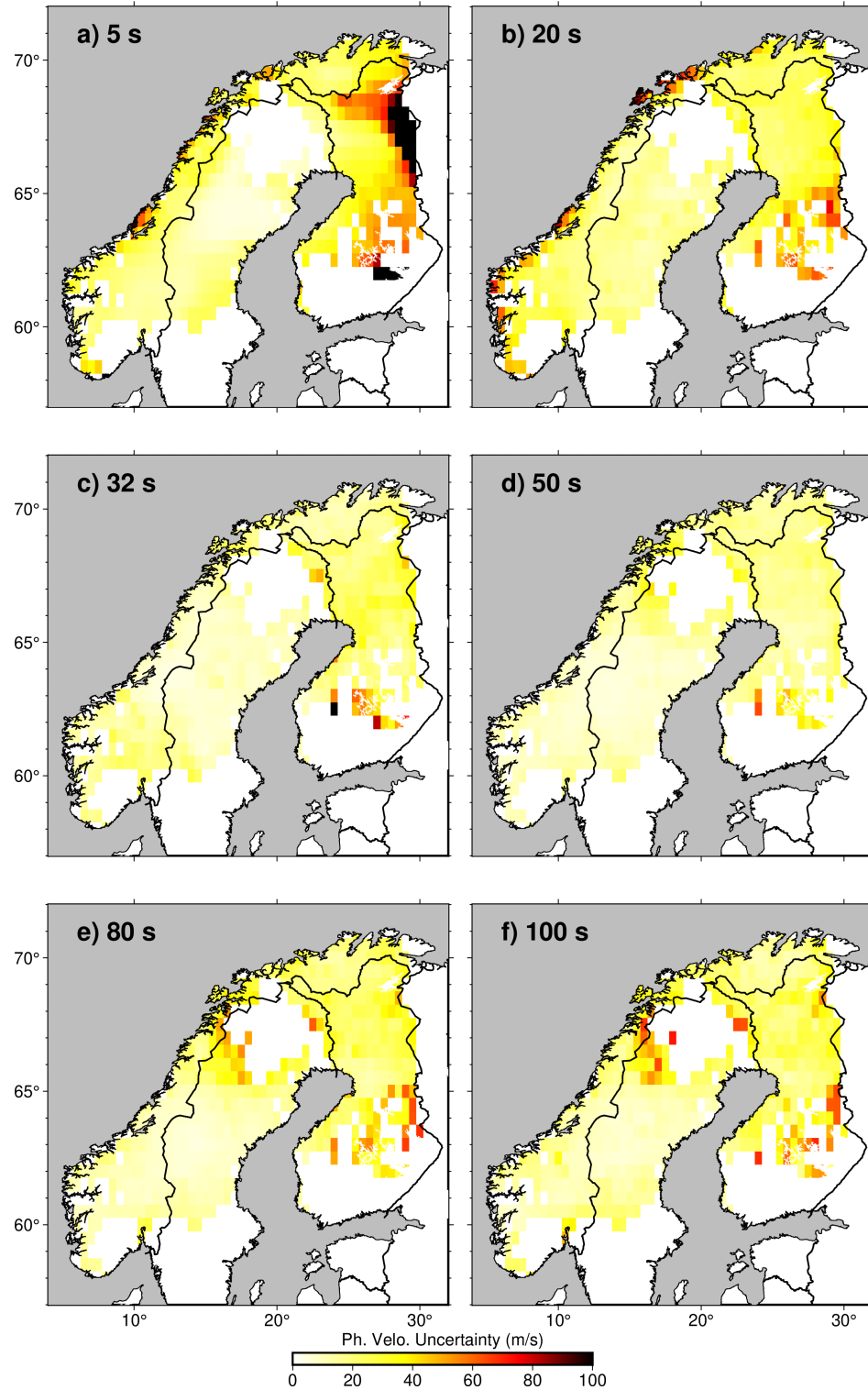


- Finland Revealed by Analysis of Local Events Registered by the SVEKALAPKO Seismic Array.” *Tectonophysics* 394 (1–2): 41–67.  
<https://doi.org/10.1016/j.tecto.2004.07.056>.
- Yuan, Huaiyu, and Barbara Romanowicz. 2010. “Lithospheric Layering in the North American Craton.” *Nature* 466 (7310): 1063–68.  
<https://doi.org/10.1038/nature09332>.
- Zhang, Peiyu, Lianjie Man, Liang Yuan, Xiang Wu, and Junfeng Zhang. 2025. “Ultra-Low-Velocity Disordered CaCO<sub>3</sub> May Explain Mid-Lithospheric Discontinuities.” *Journal of Geophysical Research: Solid Earth* 130 (9): e2025JB031906. <https://doi.org/10.1029/2025JB031906>.
- Zhao, Yajie, Xin Deng, Ling Chen, and Zhongqing Wu. 2024. “Is There a Carbonated Mid-Lithosphere Discontinuity in Cratons?” *Journal of Geophysical Research: Solid Earth* 129 (6): e2024JB028925. <https://doi.org/10.1029/2024JB028925>.
- Zhu, Hejun, Ebru Bozdağ, and Jeroen Tromp. 2015. “Seismic Structure of the European Upper Mantle Based on Adjoint Tomography.” *Geophysical Journal International* 201 (1): 18–52. <https://doi.org/10.1093/gji/ggu492>.

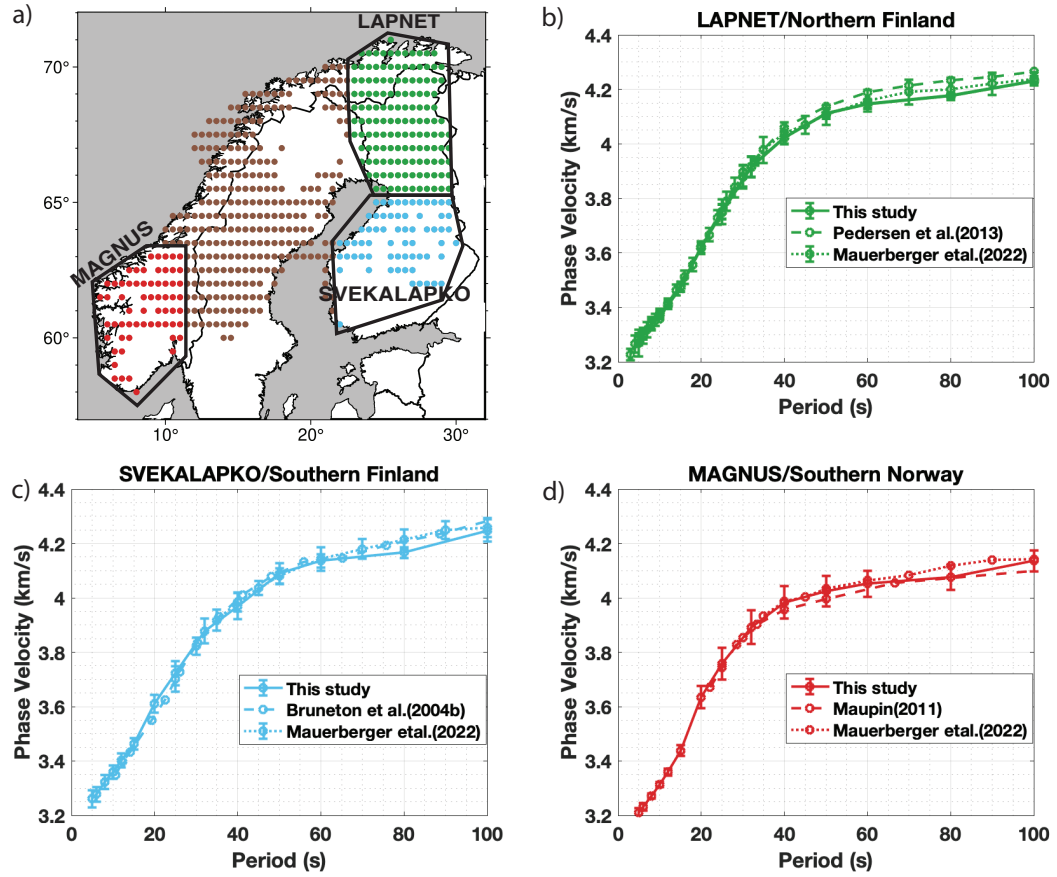
## 7 Supplemental Material

|                                | Peridotite | Eclogite | ARC_9 |
|--------------------------------|------------|----------|-------|
| SiO <sub>2</sub>               | 44.34      | 48.86    | 42.9  |
| MgO                            | 45.13      | 17.14    | 49.2  |
| FeO <sub>t</sub>               | 8.00       | 7.19     | 6.5   |
| CaO                            | 0.97       | 10.58    | 0.1   |
| Al <sub>2</sub> O <sub>3</sub> | 1.09       | 14.16    | 0.3   |
| Na <sub>2</sub> O              | 0.12       | 1.60     | 0.1   |
| Cr <sub>2</sub> O <sub>3</sub> | 0.27       | 0.48     | 0.4   |

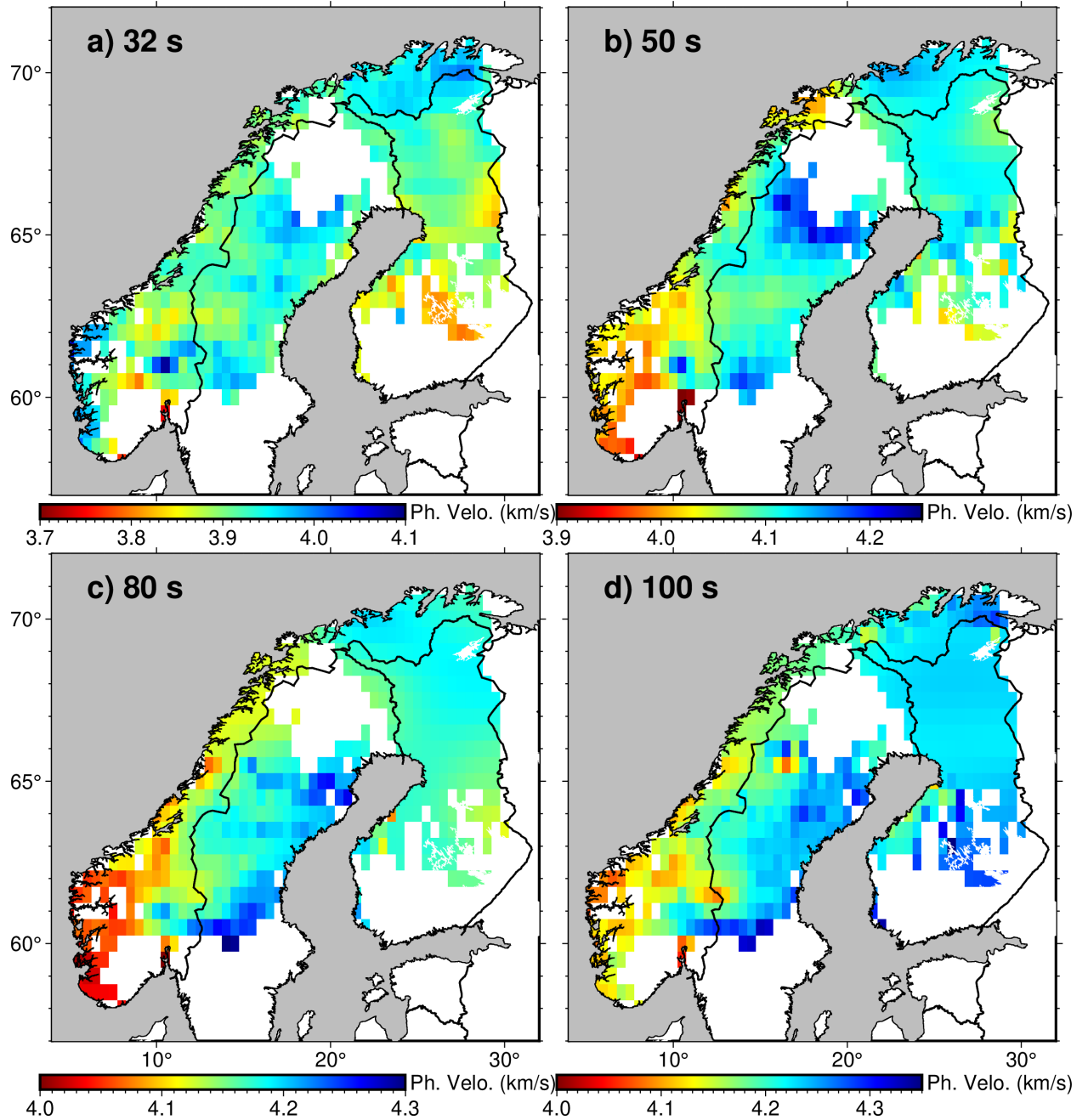
*Table 3.1: Table S1.* Bulk compositions used for forward modeling of shear-wave velocities. Peridotite and eclogite are based on average xenolith compositions from Fennoscandia.



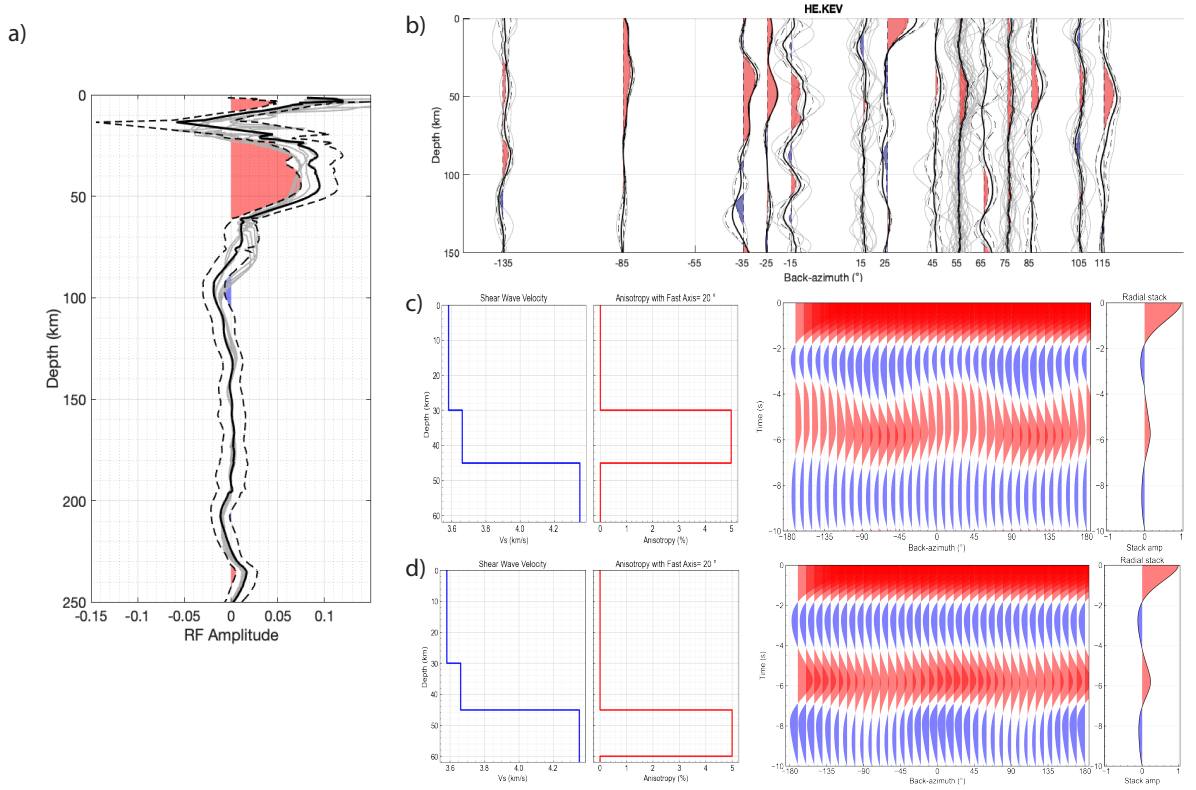
*Figure 3.15: Figure S1. (a) Uncertainty map of Rayleigh wave phase velocity at 5 s from ambient noise (Mauerberger et al., 2022; Köhler et al., 2012, 2015). (b) Uncertainty map of weighted average phase velocity at 20 s combining ambient noise and teleseismic measurements. (c-f) Eikonal phase velocity uncertainty maps at 32-100 s derived from teleseismic earthquakes calculated following Lin et al. (2009).*



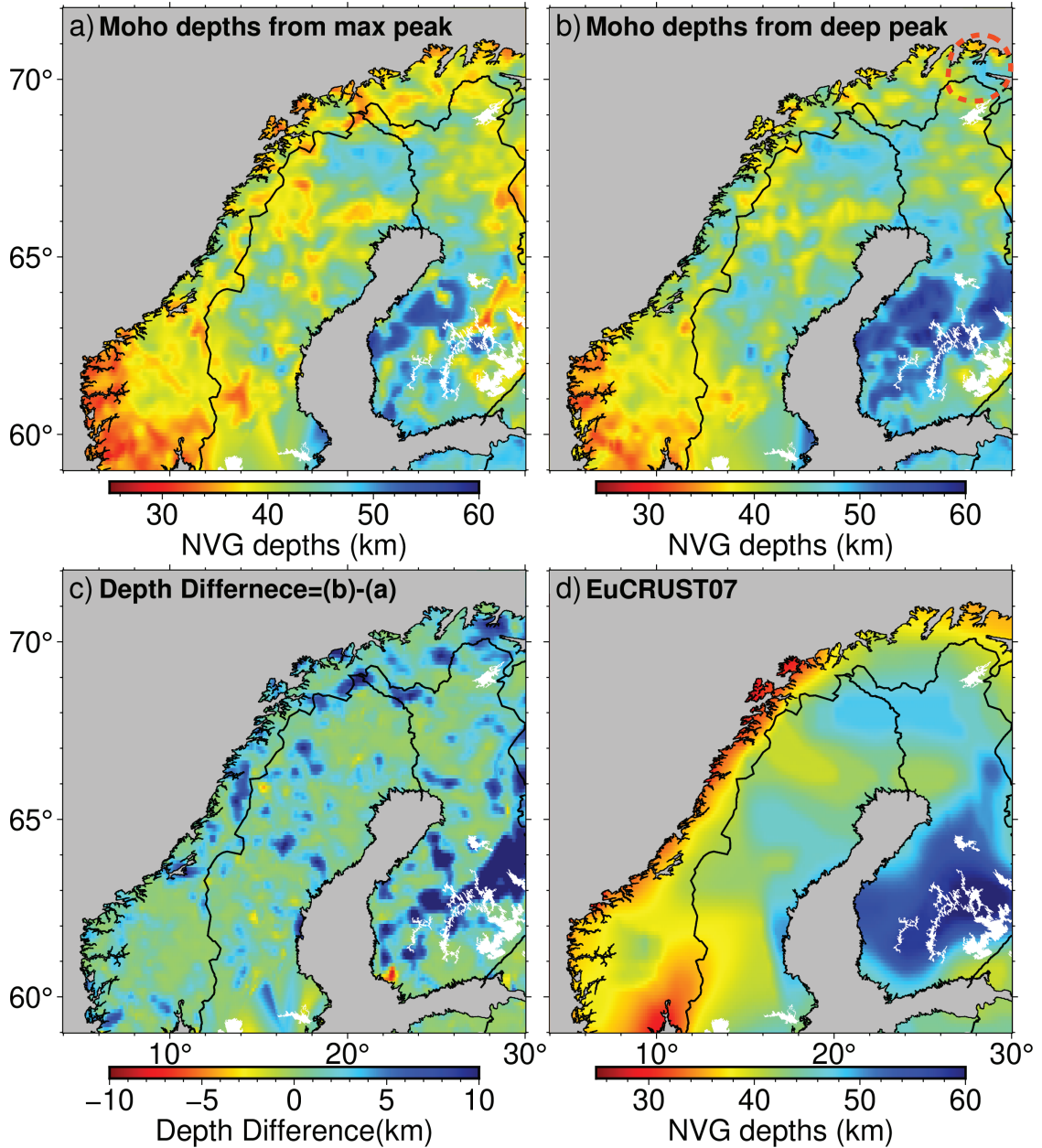
*Figure 3.16: Figure S2.* A comparison of Rayleigh wave phase velocity measurements in subregions with previous studies. (a) Distribution of individual measurements, color-coded by subregions. Green, blue, and red dots indicate subregions mainly covered by LAPNET, SVEKALAPKO, and MAGNUS respectively. (b-d) Comparisons of phase velocity measurements from this study with those from Mauerberger et al. (2022), and Pedersen et al. (2013) in LAPNET, Bruneton et al. (2004b) in SVEKALAPKO, and Maupin (2011) in MAGNUS.



*Figure 3.17: **Figure S3.** Azimuth-weighted average Rayleigh wave Eikonal phase velocity maps derived from teleseismic earthquakes at periods of (a) 32 s, (b) 50 s, (c) 80 s, and (d) 100 s. The azimuth-weighted averages closely resemble the direct stacked results shown in Fig. 2.*

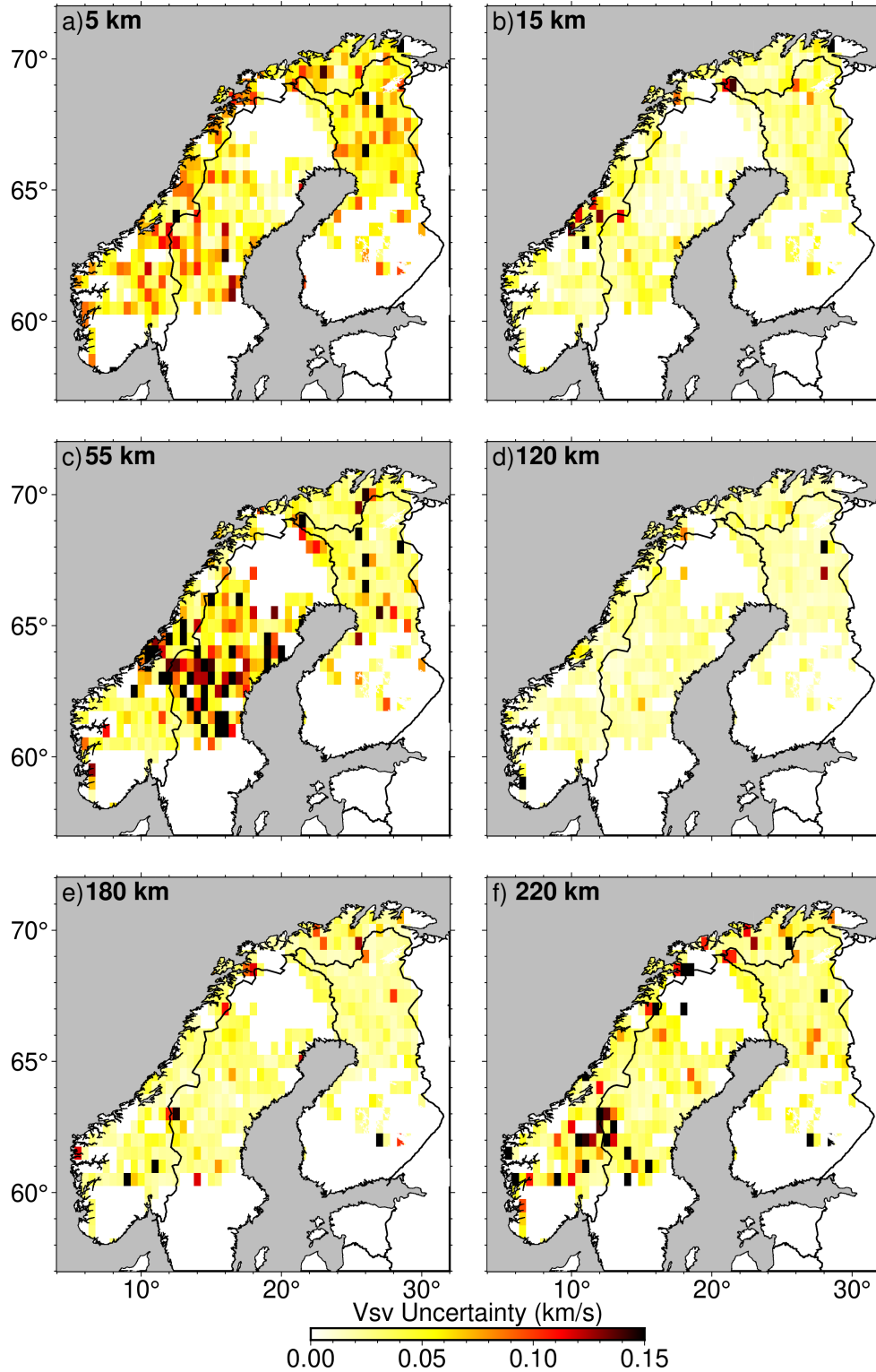


**Figure 3.18: Figure S4.** (a) CCP stack amplitudes at HE.KEV pixel (black solid line), with dashed lines showing one standard deviation of CCP stacks. Sp receiver functions are polarity-flipped to match Ps conventions. Red colors indicate positive velocity gradients and blue values indicate negative velocity gradients. Gray lines show other CCP columns within  $0.1^\circ$  of HE.KEV. Double Moho peaks are visible in some CCP columns, whereas deeper phases are more coherent. (b) Single-station receiver function stacks at HE.KEV sorted by back-azimuth bins. Despite incomplete azimuthal coverage, no clear periodic variation in Moho depth and amplitude is observed. (c) Synthetic test for double Moho peaks caused by azimuthal anisotropy. Left: model with a 15 km-thick anisotropic layer above the Moho (5% peak-to-peak anisotropy, fast-axis orientations of  $10^\circ$ ). Middle: synthetic S-p receiver functions sorted by back-azimuth. Right: stacks of all back-azimuth bins. No clear double Moho peak appears in the stack with the 4-100 s filter band. (d) Same as (c), except with a 15 km-thick anisotropic layer below the Moho.



**Figure 3.19: Figure S5.** Comparison of Moho depths. (a) Moho depth map derived from the maximum negative peaks of CCP stacks in this study. (b) Moho depth map derived from the deeper negative peaks of CCP stacks, where two peaks are present. (c) Difference in Moho depth between panels (b) and (a). (d) Moho depth map from the compilation model EuCRUST-07 (Tesauro et al. 2008). The red circle in (b) highlights the better agreement of our results with Silvennoinen et al. (2014) compared to the compilation model.





*Figure 3.20: **Figure S6.** Map views of horizontal Vsv uncertainty slices extracted from the joint inversion models at depths of (a) 5 km, (b) 15 km, (c) 55 km, (d) 120 km, (e) 180 km, and (f) 220 km.*



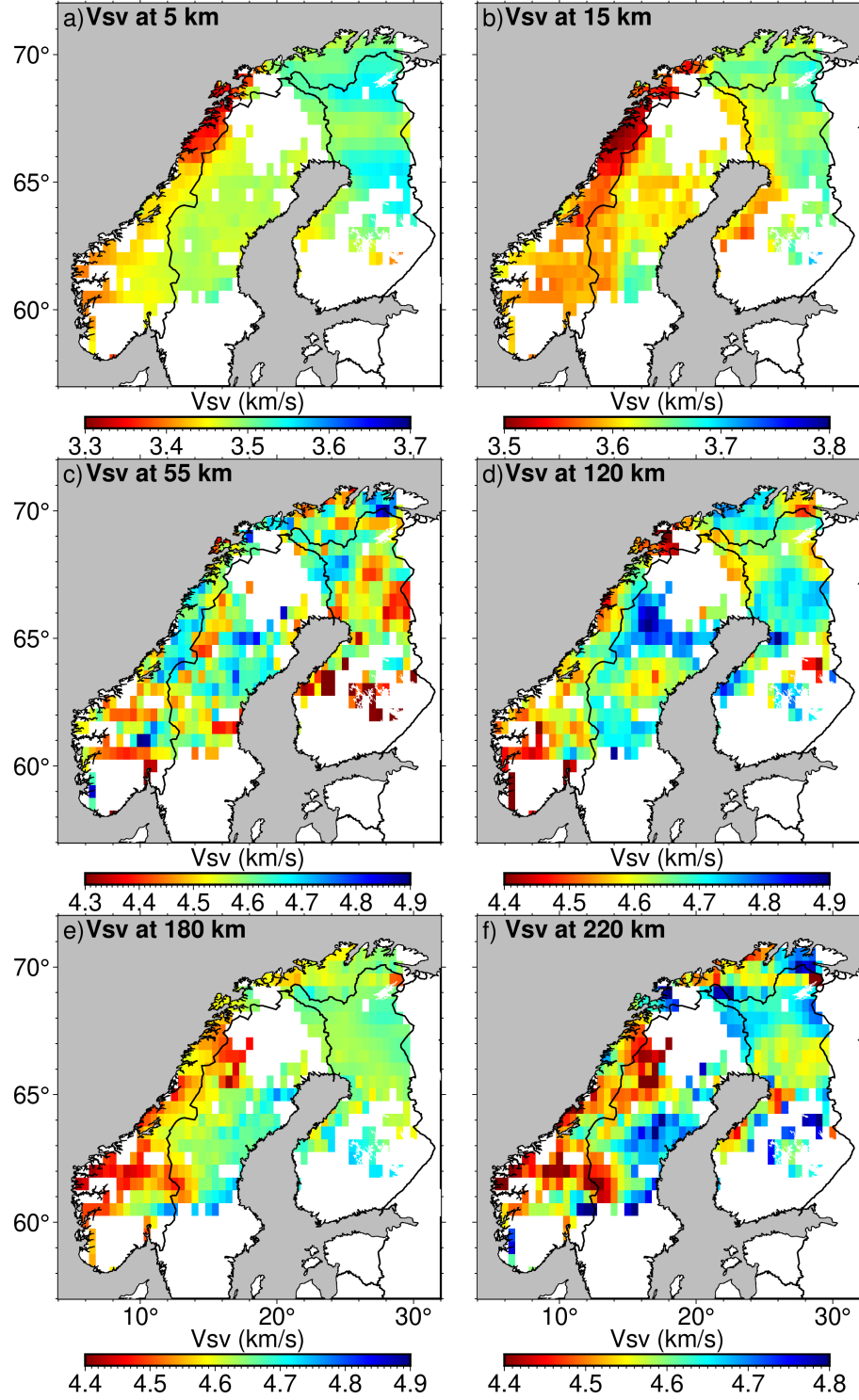
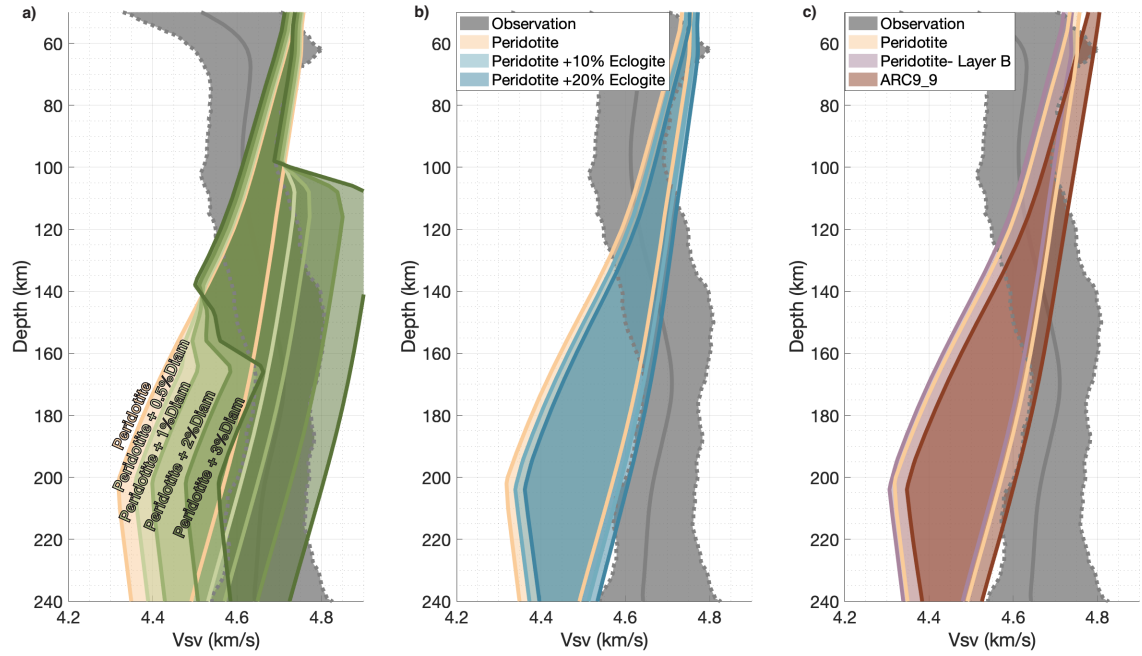
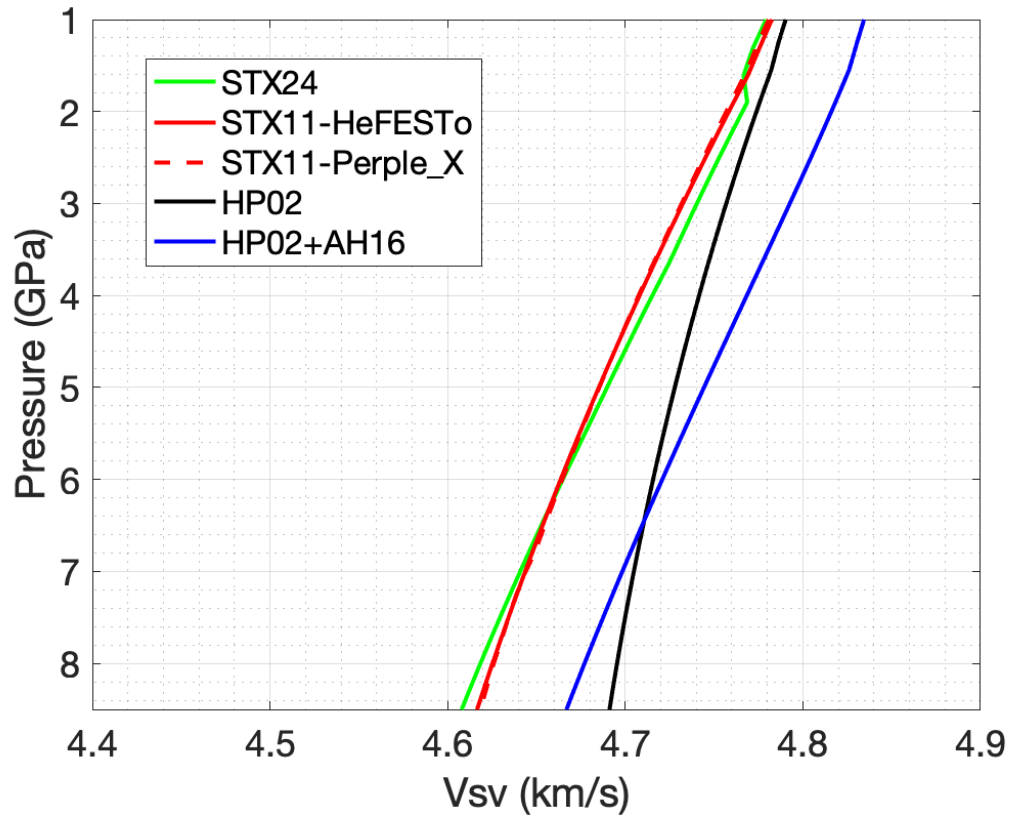


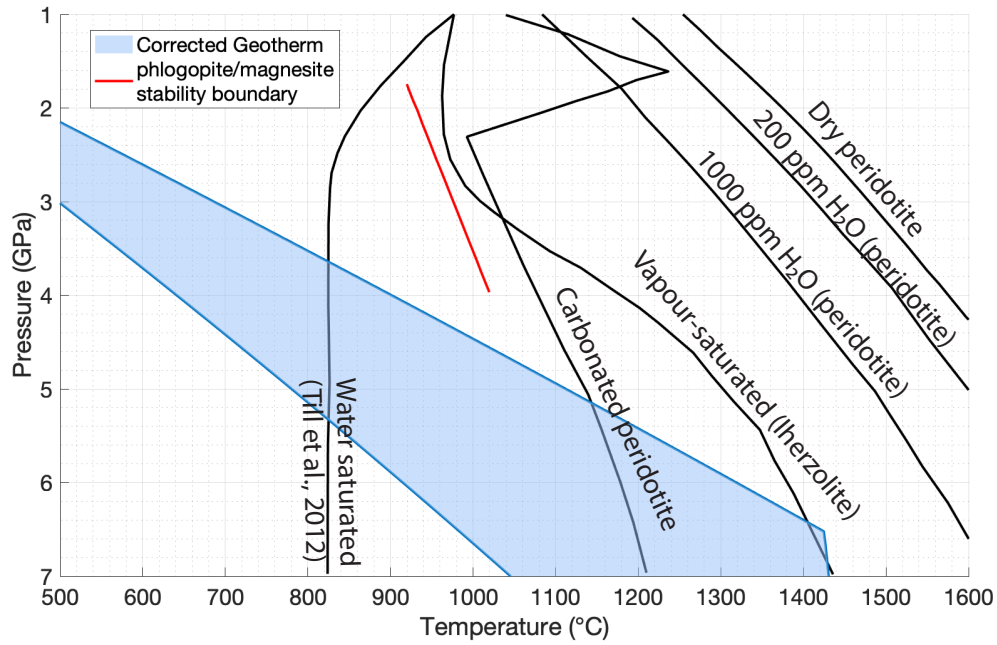
Figure 3.21: **Figure S7.** Map views of horizontal  $V_{sv}$  slices extracted from the phase-velocity-only inversion models at depths of (a) 5 km, (b) 15 km, (c) 55 km, (d) 120 km, (e) 180 km, and (f) 220 km.



**Figure 3. 22: Figure S8.** Predicted Vs under alternative compositional assumptions. Gray shading indicates observed Vs in the Fennoscandian craton interior. Yellow shading shows predictions for isotropic peridotite along the hot geotherm range (Fig. 13a). (a) Diamond. green shading shows predictions for peridotite plus diamond fractions from 0.5 to 3 vol%. (b) Eclogite. Blue shading shows predictions for peridotite containing 10 and 20 vol% eclogite. (c) Periodide composition. pale pink uses the average Layer B peridotite composition (110–180 km) from [Peltonen and Brügmann \(2006\)](#), light red uses the depleted ARC-9 composition (W. L. Griffin et al. 2009). Peridotite and ARC-9 compositions are listed in Table S1.



*Figure 3.23: **Figure S9.** Predicted  $V_s$  using different thermodynamic datasets and modeling approaches for the peridotite composition in Table S1 along the coldest acceptable geotherm in Fig. 13a. Green: STX24 + HeFESTo. Red solid: STX11 + HeFESTo. Red dashed: STX11 + Perple\_X. Blue: HP02 + Perple\_X. Black: HP02 phase assemblages with elastic parameters from Abers and Hacker (2016).*



*Figure 3.24: **Figure S10.** Experimental solidus temperatures and key stability boundaries compared with the acceptable geotherm range for Fennoscandia (blue shading; same as Fig. 13a). The red curve denotes the stability boundary for phlogopite and magnesite. Modified from [Selway et al. \(2015\)](#) and [Saha and Dasgupta \(2019\)](#).*

## **CHAPTER 4| Layered Cratonic Lithosphere in Southern Africa: Insights into Global Craton Evolution**

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## Abstract

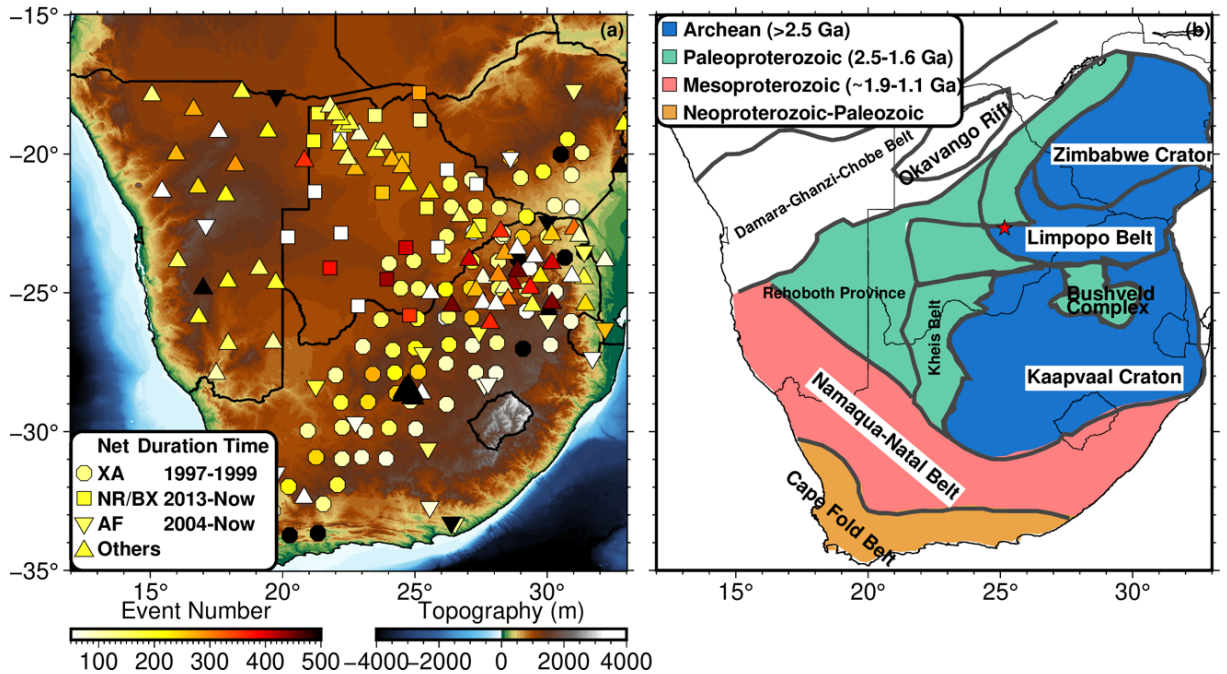
The southern Africa cratons are unique in its relatively low Rayleigh wave phase velocities compared to most other cratons, as well as their elevated topography. Its upper mantle seismic velocity structure, which reflects craton evolution and contributes to uplift, remains controversial. In this study, we develop a 3-D vertically polarized shear wave velocity ( $V_{sv}$ ) model for southern Africa through joint inversion of measured fundamental-mode Rayleigh wave phase velocities from earthquakes in the period range 25-120 s, and  $S_p$  receiver functions from a CCP stack, in order to better constrain complex layered structures. We compare this  $V_{sv}$  model with a recent one obtained for the Fennoscandia craton using similar data sets and joint-inversion approaches. The seismic structure of the shallowest upper mantle above  $\sim 150$  km is similar in both locations, with a high- $V_{sv}$  lid and intermittent layering. However, in the depth range 150-200 km, we find a  $\sim 2\%$  velocity reduction in southern Africa relative to Fennoscandia, suggesting that the southern Africa lithosphere has experienced a thermal perturbation relative to intact Archean cratons. We surveyed average  $V_{sv}$  in the depth range 150-200 km at Archean cratons around the globe and found that the global distribution has two peaks ( $\sim 4.6$  km/s and  $\sim 4.7$  km/s), consistent with the differences observed in southern Africa and Fennoscandia. On this basis, we can divide global cratons into two main classes: those that have been thermally perturbed and those that are largely intact. This conclusion is bolstered by a negative correlation between  $V_{sv}$  and topography across global Archean cratons, which suggests that the thermal perturbation contributes, at least in part, to the elevated topography of craton plateaus.

# 1 Introduction

Cratons are characterized by thick, cold and depleted lithosphere with high seismic velocities and low attenuation, and have remained stable over billions of years (e.g., Dalton et al., 2017; Eaton et al., 2009; Fischer et al. 2020; Lee et al., 2011). However, several common puzzles remain, including the apparent incompatibility between seismic and geochemical constraints (e.g., Garber et al. 2018; Hirsch et al. 2015), suggesting that the cratonic uppermost mantle is more complex than a single high velocity bulk composition. Moreover, global studies have revealed heterogeneous seismic structures in cratons, including variations in the depth and character of the mid-lithospheric discontinuity (e.g., Abt et al. 2010; Chen 2017; Krueger et al. 2021a; T. Liu and Shearer 2021; T. Liu et al. 2023; Carr and Olugboji 2024; Legre and Olugboji 2025), differences in radial anisotropy strength and depth extent (e.g., Boyce et al. 2024), and the need for variable lithospheric temperatures and degrees of metasomatism to reconcile seismic observations (Eeken et al. 2018). These features reflect diverse evolutionary histories from craton formation to their present state.

The southern Africa cratons are key regions to study craton evolution due to its distinctive seismic and geochemical characteristics. First, it exhibits lower Rayleigh wave phase velocities at longer periods compared to most other cratons, implying relatively lower shear-wave velocities at greater depths in the uppermost mantle (Pedersen et al. 2009). Second, the Kaapvaal craton in southern Africa is notable for its excess orthopyroxene for a given Mg#, suggesting metasomatism by Si-rich melts or fluids derived from subducting oceanic crust (Lee et al. 2011 and references therein). These

features indicate that the southern African cratons have undergone modification distinct from typical Archean cratons.



**Figure 1.** (a) Topography map of southern Africa from GEBCO (Weatherall et al. 2015) and locations of seismic stations used for Rayleigh wave phase velocity and Sp receiver function measurements. (b) Geological map showing the main tectonic provinces in southern Africa. Bold black lines represent tectonic boundaries, and thin black lines represent national borders. The red star indicates the magnitude 6.5 earthquake that occurred on 3 April 2017 (USGS, 2017).

In this study, we focus on the Kalahari craton in southern Africa, which is comprised of the Archean Kaapvaal craton (formed between 2.7 and 3.7 Ga; de Wit et al. 1992) and the Zimbabwe craton (formed between 3.6 to 2.5 Ga; Wilson 1990), sutured together by the Limpopo Belt (Fig.1). The Limpopo orogenic belt formed during the collision between these two Archean blocks around 2.7 Ga, trending approximately east-west. The Bushveld Complex, one of the largest layered mafic intrusions in the world, was emplaced in the northern Kaapvaal craton at ~2.05 Ga (de Wit et al. 1992). During the Proterozoic, several younger terranes accreted to the Kalahari craton, including the



Magondi and Okavango terranes, the Kheis Belt, and the Rehoboth Province along the western margin (McCourt 2001; Hanson 2003), followed by the Namaqua-Natal Belt associated with the Namaqua or Kibaran Orogeny (1.0–1.2 Ga; Cornell et al. 2011). In the northwest, the Damara-Ghanzi-Chobe Belt formed during the collision between the Kalahari and Congo Cratons (~870–550 Ma; Kröner & Stern, 2004). The Okavango Rift Zone, located within this belt, is considered an incipient rift system (Modisi et al. 2000).

A characteristic feature of southern Africa is its anomalously high surface topography of 1 to 2 km (Fig. 1a). The mechanisms responsible for this uplift remain debated. Proposed explanations include thermal perturbation in the lithospheric mantle associated with plume activity (Lithgow-Bertelloni and Silver 1998) and sublithospheric mantle convection (e.g., Li and Burke 2006). Seismic observations provide important constraints, but previous studies have yielded conflicting results. A low-velocity zone (LVZ) beneath a high-velocity layer at ~160 km, with  $V_{sv}$  of ~4.45 km/s, was first proposed by Priestley (1999) based on S waveform and Rayleigh wave phase velocities. Subsequent studies also support the presence of a LVZ: Li and Burke (2006) and Li (2011) identified a ~4% velocity reduction from Rayleigh wave phase velocity models, suggesting asthenosphere beneath 160 km; Wang et al. (2008) reported a ~5% LVZ in an SH model beneath 150 km; and Hansen et al. (2009) observed a ~4% reduction at 160 km depth using  $S_p$  receiver functions. Pedersen et al. (2009) further confirmed the existence of a LVZ beneath the Kaapvaal craton through comparison with other cratonic velocity models. Adams and Nyblade (2011) reported a maximum velocity reduction of ~3.9% deeper than ~150 km, but argued against a thermally perturbed upper mantle, noting that this magnitude is comparable to that observed in other cratons. In contrast, several studies

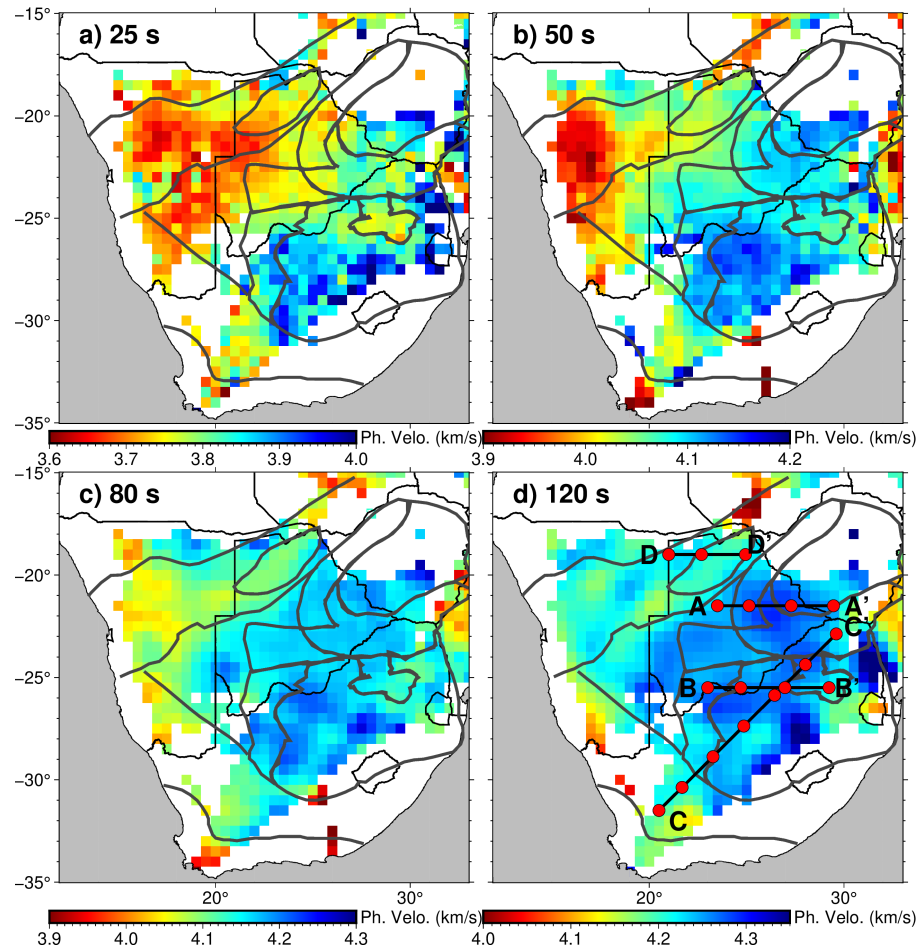
have found little or no evidence for an anomalously low velocity mantle in the southern Africa cratons based on Rayleigh and Love wave phase delays (Freybourger et al. 2001), P and S wave travel times (Fouch 2004; Youssof et al. 2015; White-Gaynor et al. 2020), and Rayleigh wave phase velocities (Marie Larson et al. 2006; White-Gaynor et al. 2020). These contrasting observations highlight the discrepancies among studies arising from differences in datasets and methodologies. Therefore, a new seismic model that integrates complementary datasets, along with comparison to other cratons, is needed to better constrain the origin of uplift in southern Africa.

In this study, we jointly invert Rayleigh wave phase velocities and  $S_p$  receiver functions to constrain the  $V_{sv}$  structure beneath the southern Africa craton. By comparing our results with the Fennoscandian craton, we show that both cratons have layered upper mantle structures with similar shallow features above  $\sim 150$  km, whereas southern Africa is characterized by a lower velocity zone than Fennoscandia at greater depths, likely indicating thermal perturbation. Furthermore, by comparing our results with seismic velocities from global Archean cratons and mantle residual topography, we evaluate the role of thermal modification in the broader context of global craton evolution.

## **2. Data and Method**

We use a transdimensional hierarchical Bayesian joint inversion, introduced by Eilon et al. (2018), to obtain shear wave velocity structure beneath southern Africa from fundamental-mode Rayleigh wave phase velocities and  $S_p$  receiver functions. This method has been successfully applied to resolve seismic velocity structure in Alaska (Gama et al. 2022), the southwestern U.S. (Golos et al. 2024) and Fennoscandia (Huang

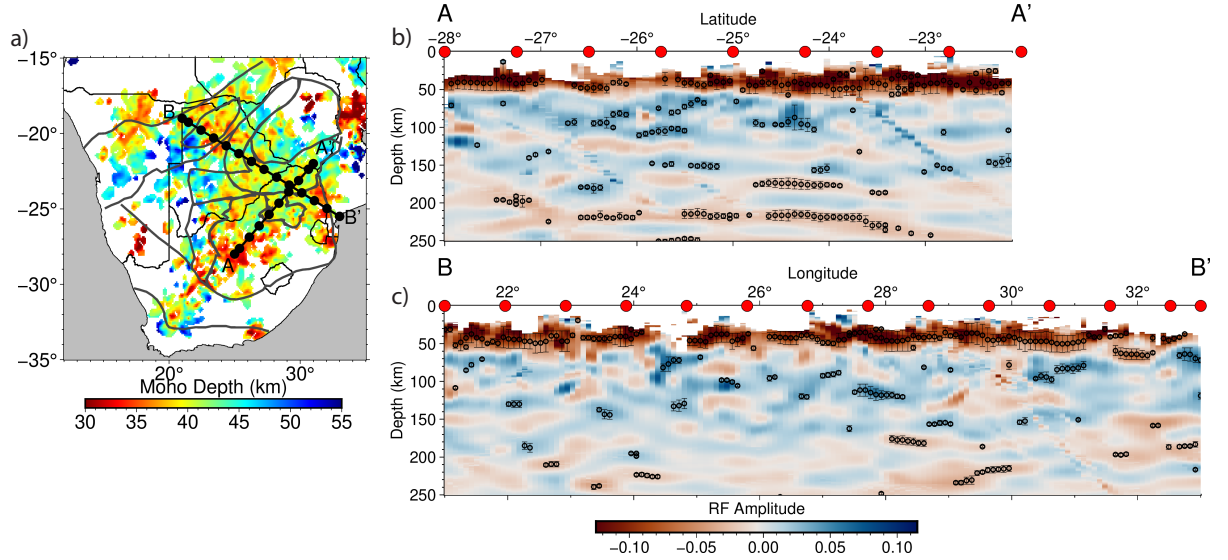
et al. 2024), and with additional data types in the eastern U.S. (Brunsvik et al. 2024). The complementary sensitivities of the two datasets used here, including Rayleigh wave phase velocities to absolute vertically polarized shear-wave speed ( $V_{sv}$ ), and Sp converted waves to detailed vertical velocity gradients, allow for improved constraints on  $V_{sv}$  as a function of depth compared to using a single dataset, as verified through synthetic tests (Huang et al. 2024).



**Figure 4.2: Figure 2.** Eikonal Rayleigh wave phase velocity maps at (a) 25 s, (b) 50 s, (c) 80 s, and (d) 120 s. Country borders and tectonic province boundaries are the same as in Fig.1b. The four lines in (d) indicate the locations of the profiles in Fig. 3.

Fundamental-mode Rayleigh wave phase velocities at eight periods between 25 and 120 s are measured from teleseismic earthquakes with depths shallower than 75 km and  $M_w > 5.5$ , recorded by 337 stations from 16 seismic networks between 1997 to 2025 (Fig. 1a). These include networks 6H (George Helffrich and Fonseca 2011), 8A (Andy Nyblade 2020), AF (Penn State University 2009), BX (Botswana Geoscience Institute 2001), GE (GEOFON Data Centre 1993), GT (Albuquerque Seismological Laboratory (ASL)/USGS 1993), II (EarthScope Consortium 1986), IU (Albuquerque Seismological Laboratory/USGS 1988), MZ (Instituto Nacional de Minas 2025), NR (Utrecht University (UU Netherlands) 1983), XA (Paul Silver 2001), XK (Steve Gao et al. 2016), XV (Michael Wyssession et al. 2015), YB (Frank Vernon et al. 1998), ZQ, ZT (Andy Nyblade 2023). We calculate Rayleigh wave phase velocity maps using Eikonal tomography (Lin et al. 2009), with phase delays measured using the Automated Surface Wave Measuring System (ASWMS) software (Jin and Gaherty 2015). Phase velocity maps are first generated for each event and then averaged to obtain a composite map at each period. Dispersion curves are extracted for each  $0.5^\circ \times 0.5^\circ$  grid cell with at least three measurements. A total of 2397 events are used. The resulting phase velocity maps and their uncertainties are shown in Fig. 2 and Fig. S1, respectively. Higher phase velocities are observed within the Kalahari craton compared to surrounding regions, while the Cape Fold Belt exhibits lower velocities across 25–100 s. Within the Kalahari craton, relatively low phase velocities at 25 s in the central and southern Kaapvaal craton likely reflect variations in crustal thickness. The Bushveld Complex also shows relatively slower velocities at periods shorter than 60 s, consistent with previous studies (e.g., Wang et al. 2019; Li and Burke 2006; Yang et al. 2008).

Sp receiver functions are extracted from a 3-D common conversion point (CCP) stack of Sp receiver functions. Individual receiver functions are computed using a time-domain deconvolution method (Ligorria and Ammon 1999) from the same station dataset used for Rayleigh wave measurements, with a 2–100 s bandpass filter applied. We select earthquakes with  $M_w > 6$  and epicentral distances between  $30^\circ$  and  $90^\circ$ , and Sp receiver functions are obtained for data from a more limited distance range ( $55^\circ$  to  $85^\circ$ ). A total of 18,626 Sp receiver functions are used to calculate the CCP stack using Fresnel-zone-based weighting functions (Hua et al. 2020). A 3D velocity model (GLAD-M25; Lei et al. 2020) is used to migrate the 1D receiver functions to the time domain for joint inversion, and is also used in the generation of CCP stack. To ensure robustness, three quality-control criteria are applied. (1) Prior to CCP stacking, individual Sp receiver function outliers are removed similar to Hua et al. (2020), and Brown and Fischer (2025), based on the coherence of Moho conversions and abnormally mantle phases. Receiver functions with positive amplitudes using the natural polarity of Sp receiver functions, exceeding three times the median amplitude in the 15–60 km depth range are discarded, as these are inconsistent with a clear Moho phase. Additional outliers are removed based on the squared L2 norm  $\|RF - 0.05\|_2^2$  exceeding five times the median value. (2) Only CCP columns with weights above 0.03 for all but 5 km in the 20 and 250 km are retained, where the normalized summed phase weights are defined by Equation 27 in Hua et al. (2020). (3) To reduce the influence of anomalous amplitudes, we use the median of receiver functions within  $0.5^\circ$  of each grid cell (expanded to  $0.75^\circ$  if fewer than 20 receiver functions are available). The resulting Moho depth and two representative profiles through the CCP stack are shown in Fig. 3.

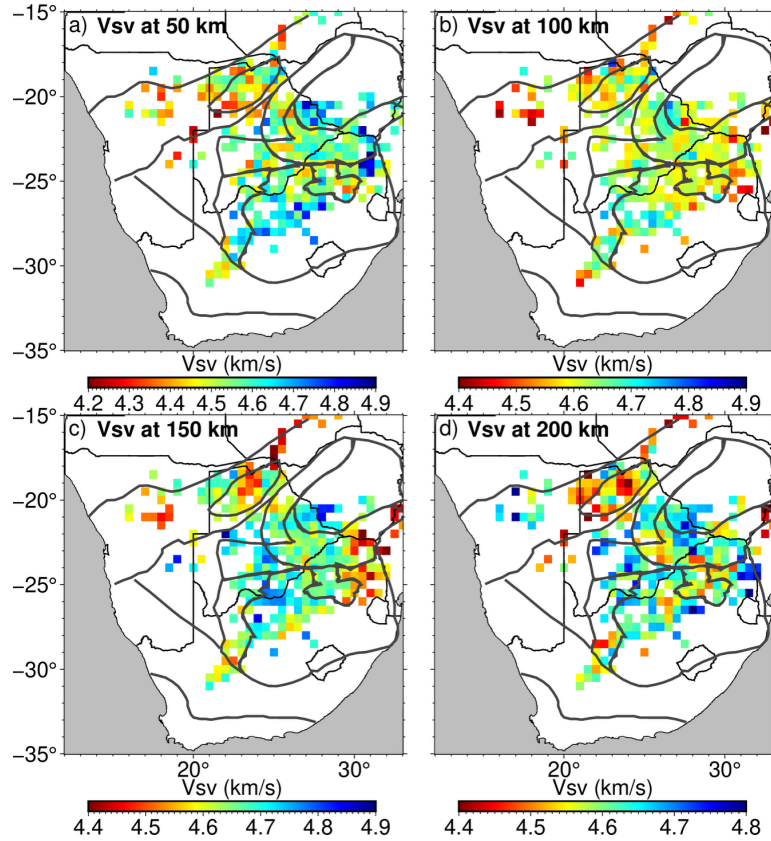


**Figure 4.3: Figure 3.** (a) Moho depth extracted from the Sp receiver function CCP stack. The two lines indicate the locations of two CCP stack amplitude profiles shown in (b) and (c). Sp receiver functions are shown in the natural polarity of Sp phases. Red colors (negative amplitudes) indicate positive velocity gradients and blue values (positive amplitudes) indicate negative velocity gradients. Black circles with error bars represent the most robust phases, with amplitudes exceeding twice their standard deviations.

We jointly invert Rayleigh wave phase velocities and Sp receiver functions using a Monte Carlo Markov Chain (MCMC) Bayesian framework following Eilon et al. (2018) in 348 locations, with the same parameterization as Huang et al. (2024). Here we briefly summarize the method, and we refer the reader to Eilon et al. (2018) for more detail. (1) The inversion begins by randomly perturbing model parameters according to prior probability distributions. The Vsv structure is parameterized using cubic B-spline functions, with both the number and depths of control points allowed to vary between 0 and 250 km. In addition to the spline parameters, Moho depth and data-error hyperparameters are also treated as free parameters. (2) The misfit between observed Rayleigh wave phase velocities and Sp receiver functions and their corresponding predictions from the perturbed model is then calculated. (3) Based on the misfit and

posterior probability, the new model is accepted or rejected according to the Metropolis–Hastings acceptance criterion (Hastings 1970). (4) The model is then updated: if accepted at (3), the new model is further perturbed; if rejected at (3), the last accepted model is perturbed. Steps (2) and (3) are repeated for subsequent iterations. The inversion is parallelized over six chains, each with 15,000 iterations per location (pixel). Acceptable models are saved every 25 iterations after the first 8,000 burn-in models. The final  $V_{sv}$  model is defined as the median of all saved acceptable model ensemble at each depth.

Although the relative weighting between phase velocities and Sp receiver functions is controlled by evolving data error hyperparameters, the bounds imposed on these hyperparameters influence their effective contribution. We find that the fit to the two datasets is not equally good for all of the 348 pixels, particularly for shallow negative velocity gradients required by receiver functions and the shortest-period phase velocities. To address this, we generate two model sets with different relative weightings. Model 1 (Fig. 4) assigns greater weight to phase velocities ( $\text{Phv } \sigma_{\max}=0.25$ ;  $\text{Sp RF } \sigma_{\max}=0.15$ ), resulting in a better fit to phase velocities but a poorer fit to receiver functions. Model 2 (Fig. S2) has a greater weight for the receiver functions ( $\text{Phv } \sigma_{\max}=0.5$ ;  $\text{Sp RF } \sigma_{\max}=0.11$ ), leading to the opposite behavior. The largest differences between the two models occur in the uppermost mantle above 100 km, and do not affect the primary conclusions of this study. We therefore focus on Model 1, while noting that Model 2 yields consistent interpretations. The uncertainty of Model 1 is shown in Fig. S3.



*Figure 4.4: Figure 4.* Map views of horizontal  $V_{sv}$  slices from joint inversion Model 1 (with a relatively higher weight assigned to phase velocities) at depths of (a) 50 km, (b) 100 km, (c) 150 km, (d) 200 km.

### 3. Results

#### 3.1 Layered seismic structures

The CCP stack reveals three primary velocity gradient features, including (a) the Moho, with 95% of Moho depths falling within the 32–50 km depth range; (b) extensive but discontinuous negative velocity gradient (NVGs) from the Moho to ~130 km across the broad central Kalahari Craton; and (c) localized and continuous positive velocity gradients (PVGs) between 190 and 230 km, mainly across the Limpopo Belt, Bushveld Complex and central Kaapvaal Craton from 21°S to 28°S.



Moho depths are identified from the maximum amplitude of the receiver function phase that corresponds to a positive velocity gradient in the depth range from 10 to 70 km. These Moho depths show significant spatial variation across the study region. The heterogeneous Moho depths within Kalahari Craton are consistent with previous studies (e.g., Nguuri et al. 2001; Nair et al. 2006; Yang et al. 2008; Youssof et al. 2013; Wang et al. 2019). The central and southern Kaapvaal Craton has relatively thin crust (~35 km), consistent with interpretations of mafic lower crust removal associated with the deposition of the Ventersdorp Complex (Delph and Porter 2014). In contrast, the Limpopo Belt and Bushveld Complex have thicker crust (>40 km), reflecting tectonic collision (2.7 Ga) and magma intrusion (2.05 Ga) events (Nair et al. 2006). Beneath Botswana, the Okavango Rift Zone is characterized by relatively thin crust (~36 km) in its northeast and central regions, extending to the southeast beyond the rift into central Botswana. High crustal Vp/Vs ratios have been reported in this area (Fadel et al. 2018). Combined with low shear wave velocities in the middle crust, this has been interpreted as evidence for fluid infiltration from the upper mantle beneath the Okavango Rift Zone (Mpuang et al. 2023), with possible links to the 2017 Mw6.5 earthquake in central Botswana and its aftershocks (Fig. 1b). The southwestern tip of the Okavango Rift Zone has thicker crust (~50 km), extending to the southeastern Rehoboth Province. Together with thick crust east of the Zimbabwe Craton, the thick crust in Rehoboth Province is consistent with continental scale gravity-derived crustal models (Tedla et al. 2011).

A striking feature of the upper mantle in the CCP stack is the presence of NVGs from the Moho to ~130 km (Figs. 3b, c), observed across most of the study region without clear correlation to tectonic provinces. These NVGs are discontinuous but

pervasive, and multiple NVGs can occur within a single CCP column. NVGs in the shallow mantle have been reported in previous studies in southern Africa, at depths of 88 to 96 km using Ps receiver functions (Rychert and Shearer 2009), and ~50-100 km from Sp receiver functions (Krueger et al. 2021b), commonly termed mid-lithospheric discontinuities (MLDs). We note that the most significant NVGs in some other early Sp receiver function studies in southern Africa are at larger depths (~150 km) (Hansen et al., 2009; Savage and Silver, 2008). To exclude interference from crustal multiples in Ps receiver functions, Legre and Olugboji (2025) applied the CRISP-RF method (Olugboji et al. 2023) and identified prevalent NVGs between 50-100 km. Although direct comparison between our results and Ps receiver function results is challenging, both approaches indicate pervasive NVGs in a similar depth range. Additional studies (Wittlinger and Farra 2007) also reported widespread NVGs in Kalahari craton above ~150 km, attributing them to radial anisotropy in the uppermost mantle.

At greater depth (210-250 km), a PVG is observed across the Limpopo Belt, Bushveld Complex, and central Kaapvaal craton, for example over ~5° on profile BB' (Fig. 3b). A similar PVG in this depth range has been reported in independent Sp receiver function studies (L. Liu et al. 2021; Sodoudi et al. 2013; Wittlinger and Farra 2007). Wittlinger and Farra (2007) noted that the observed downward increase in velocity contrasts with the expected transition from depleted (faster) to more fertile (slower) mantle compositions.

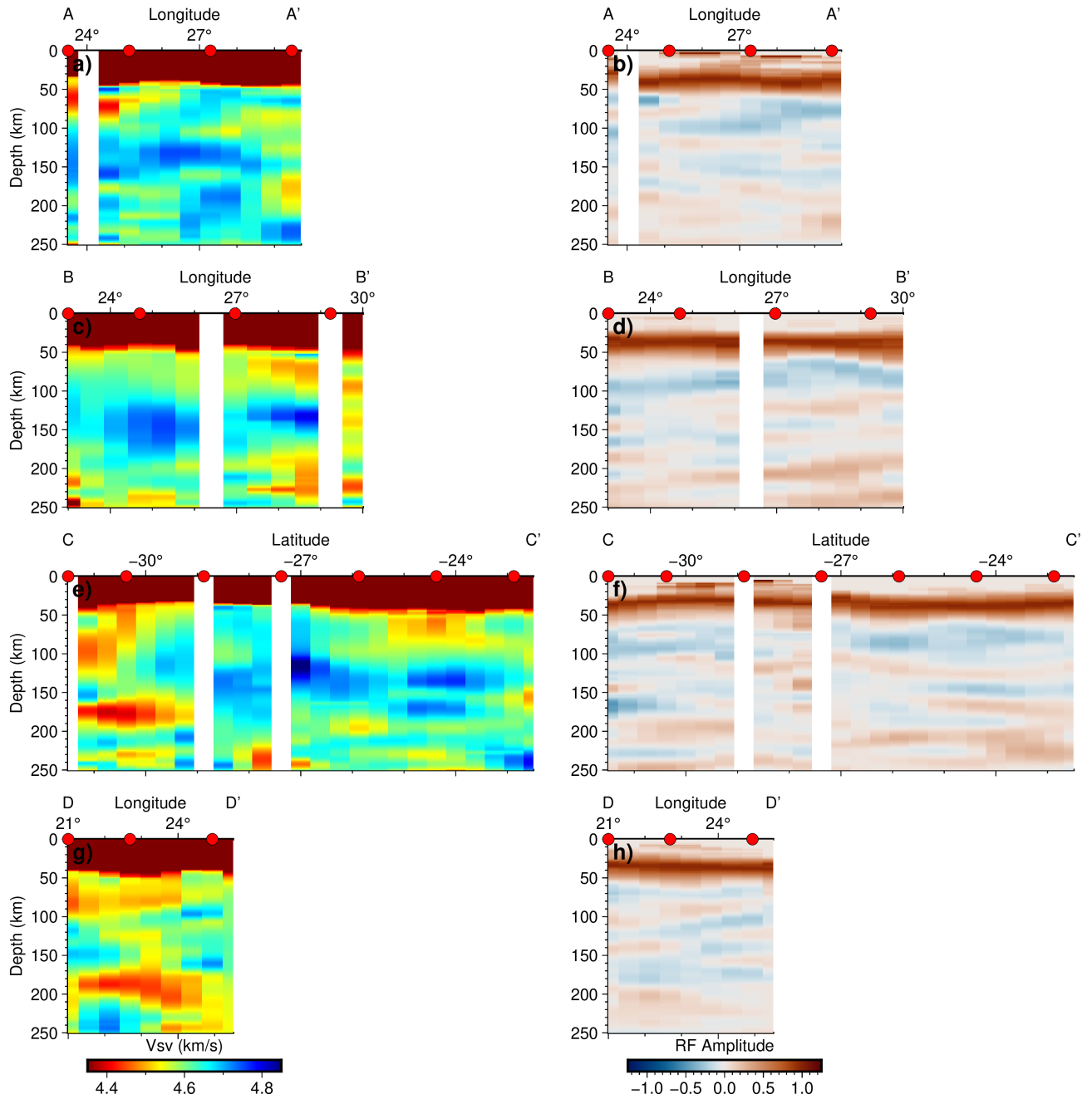
### 3.2 Shear wave velocity structures

The shear wave velocity model in southern Africa study region (Fig. 4) can be broadly divided into two clusters based on uppermost mantle  $V_{sv}$  values: a high velocity region in the Kalahari Craton in the east, and a low velocity region in the northwest, concentrated around the Okavango Rift Zone (Fig. S4). A clear boundary is observed between the Kaapvaal craton and the Namaqua-Natal Belt. Lower mantle  $V_{sv}$  values are also observed along the margins of the Kalahari Craton, which may be due to the limited ray path coverage for phase velocity measurement and larger uncertainties, as indicated by the uncertainty maps in Fig. S1. Model 2, which assigns higher weight to  $S_p$  receiver functions than Model 1 does, has a similar first-order  $V_{sv}$  structures to model, but shows more complicated layering and small-scale heterogeneities due to the improved fit to receiver functions (Fig. S2, S4). Model 2 also shows higher velocities than Model 1 from the Moho to  $\sim 100$  km, consistent with the negative velocity gradients (NVGs) preferred by the receiver functions at these depths. The large-scale velocity pattern in Model 1 and Model 2 is broadly consistent with continental  $V_{sv}$  models (e.g., Afonso et al. 2022), with  $\sim 4.7$  km/s at 150 km in the Kalahari craton,  $\sim 4.6$  km/s in the Okavango Rift, and  $\sim 4.5$  km/s in the western Damara–Ghanzi–Chobe Belt (west of  $\sim 20^\circ\text{E}$ ), although our model reveals additional local heterogeneities.

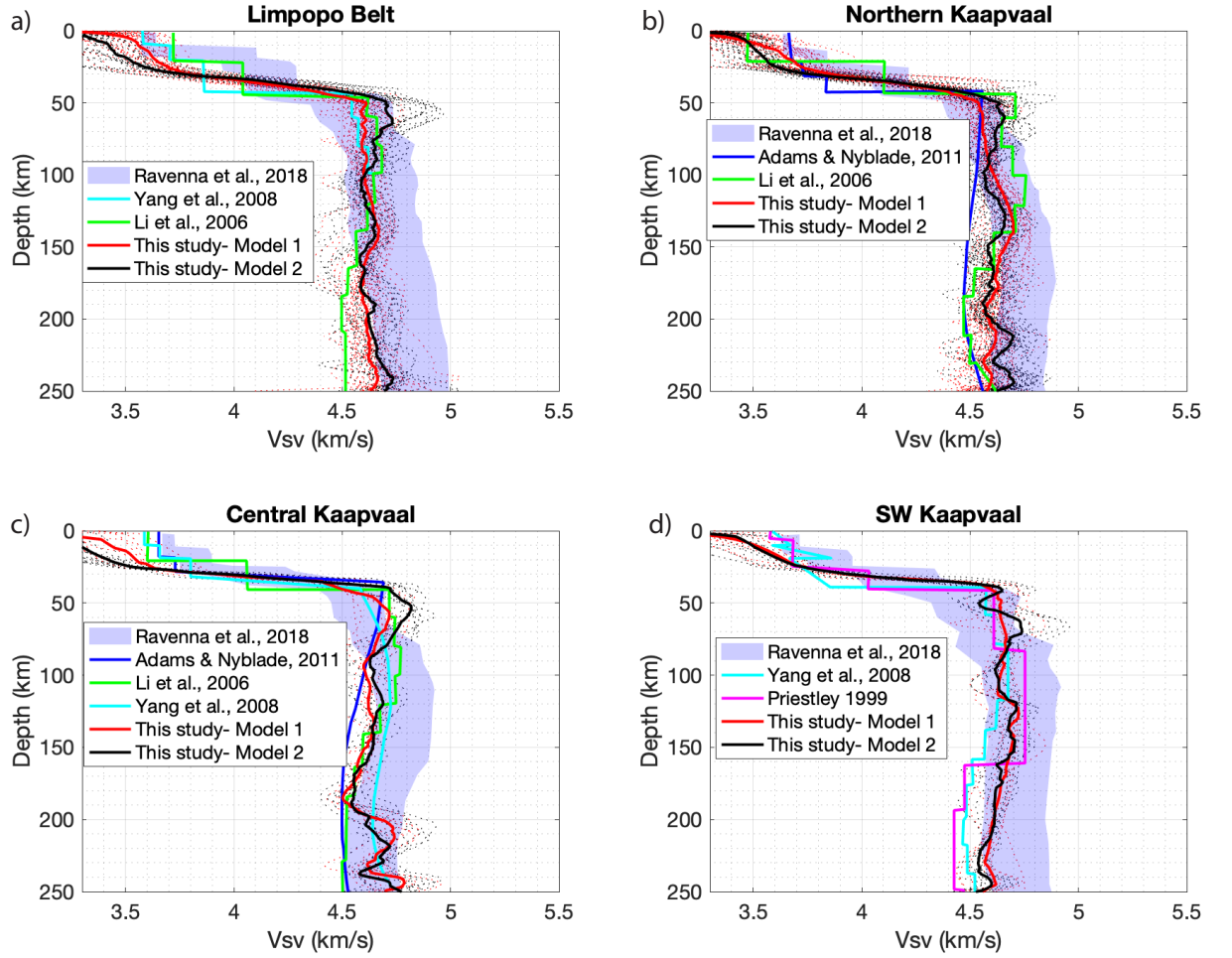
The Kalahari craton is characterized by a high-velocity lid ( $\sim 4.7$  km/s) between 100 and 150 km depth, with slightly lower velocities ( $\sim 4.6$  km/s) both above and below this layer (Fig. 5a-c). No significant differences are observed among the Zimbabwe craton, Limpopo Belt, and Kaapvaal craton, except in the northern Kaapvaal craton. In this region, including the Bushveld Complex, the uppermost mantle exhibits slower velocities

(~4.5 km/s) from the Moho to ~100 km (east of 27°E in BB' profile in Fig. 5c; north of -25°S in CC' profile in Fig. 5e). This slower uppermost mantle has been reported in previous surface wave and body wave studies (Adams and Nyblade 2011; White-Gaynor et al. 2020; Wang et al. 2019; Li and Burke, 2006), interpreted as the result of lithospheric refertilization and metasomatism associated with mafic magmatic intrusion (Fouch 2004), consistent with the thicker crust described in Section 3.1.

In Fig.6, we compare our V<sub>sv</sub> models with previous surface wave studies across four subregions defined by Ravenna et al. (2018) (Adams and Nyblade 2011; Li and Burke 2006; Yang et al. 2008; Ravenna et al. 2018; Priestley 1999). Most models have a high velocity layer at 100-150 km in all subregions, consistent with our results, although absolute V<sub>sv</sub> values vary. However, in most previous studies, velocity generally increases from the Moho to ~100 km, consistent with global craton models (Garber et al. 2018), which some authors have interpreted as the spinel peridotite-garnet peridotite transformation (Ravenna et al. 2018). In contrast, our models have weak (Model 1) or strong (Model 2) NVGs at this depth range, as indicated by the receiver functions. At depths greater than 150 km, our V<sub>sv</sub> values fall near the lower bound of those reported by Ravenna et al. (2018), but remain higher than other models. Our model contains a pronounced velocity increase at ~210 km in the northern and central Kaapvaal craton, which is also constrained by receiver functions. The differences between Model 1 and Model 2 are evident in this comparison: Model 2 exhibits stronger vertical gradients and more detailed structure, while Model 1 is smoother, but both models show consistent V<sub>sv</sub> structures between 100 and 200 km.



**Figure 5.** Vsv cross-sections from Model 1. Left column: joint-inversion models; Right column: Sp receiver function CCP stack amplitudes for profiles AA', BB', CC', and DD' where profile locations are shown in Fig. 2d. The sign of the amplitude is flipped to the Ps convention, so that a positive phase corresponds to a velocity increase with depth.



**Figure 4.6: Figure 6.** Comparison of mean Vsv profiles (solid lines) from Model 1 (red) and Model 2 (black), along with individual Vsv models (dotted lines) in this study, and previously published models from Ravenna et al. (2018), Yang et al. (2008), Li and Burke (2006), Adams and Nyblade (2011), and Priestley (1999) for different subregions, following the geographic divisions of Ravenna et al. (2018).

Velocities in off-craton regions also vary significantly. Lower velocities ( $\sim 4.35$  km/s) in the uppermost mantle ( $\sim 50$  km) are observed at the southwestern tip of the Okavango Rift Zone, whereas at  $\sim 150$  km, the northeastern tip has lower velocities (4.5 km/s) (Fig. 4). This shift in low-velocity anomalies has also been observed in a previous surface wave study (Fadel et al. 2020), which suggested that the Okavango Rift Zone represents the southern termination of the East African Rift System and that ascending fluids related

to rifting could be linked to the 3 April 2017 Mw 6.5 intraplate earthquake (Fig. 1b). However, our phase velocity uncertainties in the northeast of Okavango Rift Zone are relatively large (Fig. S1), and we therefore cannot confirm this interpretation. A low-velocity zone below ~150 km depth, observed on both flanks, appears to uplift beneath the center of the Okavango Rift Zone at ~23.5°E (Fig. 5g, h) along the E–W profile. This feature likely reflects a local shallowing of the lithosphere–asthenosphere boundary, consistent with lithospheric stretching inferred from P-wave travel time studies (Yu et al. 2017), and associated with the relative motion between the Congo and Kalahari cratons. A slow and weak uppermost mantle combined with faster mantle below ~100 km beneath the epicenter of the 3 April 2017 Botswana earthquake (A end in Fig. 5a) is consistent with a top-down extension model. In this scenario, extension is driven by large scale extensional forces acting on a mechanically weak uppermost mantle, without involving deeper mantle upwelling, as proposed by Moorkamp et al. (2019).

## **4. Discussion**

### **4.1 Layered seismic structure comparison with Fennoscandia**

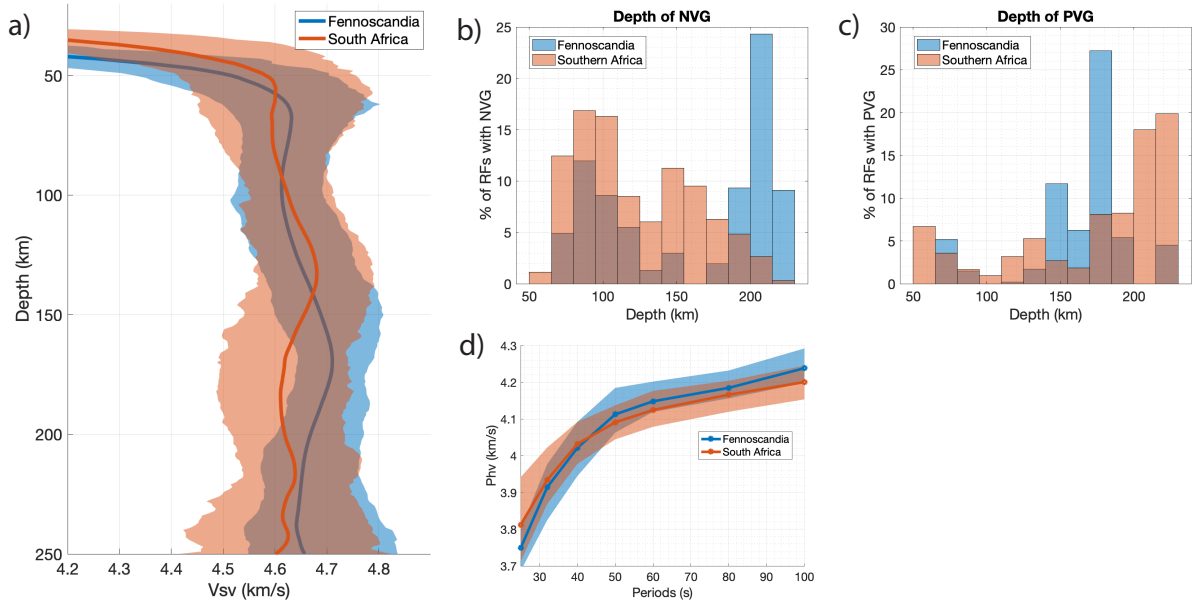
Comparing shear wave velocity models derived from the same inversion framework is essential to minimize differences introduced by varying data sensitivities and model parameterizations. For example, Sodoudi et al. (2013) showed that different methods, such as body wave tomography, surface wave tomography, and Sp and Ps receiver functions, can produce large discrepancies in lithosphere–asthenosphere boundary (LAB) depth beneath the Kalahari craton. Moreover, even when using the same Rayleigh wave

phase velocity dispersion data, different inversion approaches and parameterizations can yield significantly different  $V_s$  variations in depth due to the inherent non-uniqueness of the inversion problem (Pedersen et al. 2009). Therefore, comparisons between models derived from different inversion frameworks may include methodological artifacts. Here we compare the southern Africa craton  $V_{sv}$  model in this study with a  $V_{sv}$  model for the Fennoscandia craton interior (Huang et al. 2024), both derived using the same joint inversion of Rayleigh wave phase velocities and  $S_p$  receiver functions. This approach allows consistent recovery of both absolute  $V_{sv}$  and vertical velocity gradients, enabling a more direct comparison between the two cratons.

In Fig. 7, we compare our  $V_{sv}$  model in the southern Africa craton with  $V_{sv}$  profiles from the interior of the Fennoscandian craton, which Huang et al. (2024) identified using a k-means clustering analysis applied to their  $V_{sv}$  model. Here we apply a similar k-means clustering analysis to our southern Africa craton  $V_{sv}$  model. However, unlike Fennoscandia, where Huang et al. (2024) identified three distinct clusters (craton interior, craton edge, and non-craton), we cluster the southern Africa model into two groups (Fig. S4), as we found no systematic differences when a third cluster was added. For the southern Africa clustering analysis, we associated one cluster with the craton interior and the other cluster with non-craton region. Fig. 7 highlights similarities and key differences between the two cratons. First, both models exhibit layered structures (Fig. 7a), with a  $V_{sv}$  maximum at  $\sim 160$  km in Fennoscandia and at  $\sim 120$  km in southern Africa, consistent with the depths of the velocity maxima reported by Boyce et al. (2024). The upper mantle structure from the Moho to  $\sim 140$  km is broadly similar in both regions. However, below 160 km Fennoscandia is characterized by a fast layer with negative



velocity gradient, whereas southern Africa shows a pronounced velocity reduction in the depth range 140-200 km beneath the high-velocity lid. At depths greater than ~210 km, the  $V_{sv}$  in southern Africa increases and approaches that of Fennoscandia. However, this depth range is relatively weakly constrained by Rayleigh wave phase velocities due to the limited range of measured periods. These differences persist regardless of how the phase velocity and receiver function datasets are weighted in the joint inversion (Fig. S4b, c), indicating that they are robust features of the data and not overly sensitive to choices made in the model development. Further insight is provided by the depth distribution of robust velocity gradient phases in CCP stacks (Figs. 7b, c), identified as phases where amplitudes exceed twice the standard deviation. In southern Africa, NVGs are present from the Moho to ~130 km, with an additional peak near ~150 km, corresponding to the velocity decrease beneath the high-velocity lid. In contrast, the Fennoscandia craton interior also shows shallow NVGs, but the dominant peak occurs deeper than 200 km, reflecting a deeper velocity reduction. PVGs are observed at depths >200 km in southern Africa and at ~150–190 km in Fennoscandia. Figure 7d compares Rayleigh wave phase velocities between the two cratons. Fennoscandia shows lower velocities at periods shorter than 40 s due to its thicker crust, but higher velocities (by ~0.15 km/s) at periods longer than 50 s, reflecting faster mantle velocities compared to southern Africa, particularly below ~150 km.



**Figure 4.7: Figure 7.** Comparison of seismic structures between Fennoscandia craton interior (blue) and the southern Africa cratons (red), including (a) average  $V_{sv}$  profiles (line) and full  $V_{sv}$  distributions (shaded); depth distributions of mantle (b) NVG and (c) PVG for Fennoscandia from 70 to 250 km and southern Africa cratons from 50 to 250 km respectively; and (d) average Rayleigh wave phase velocities (bold lines) and their distributions (shaded).

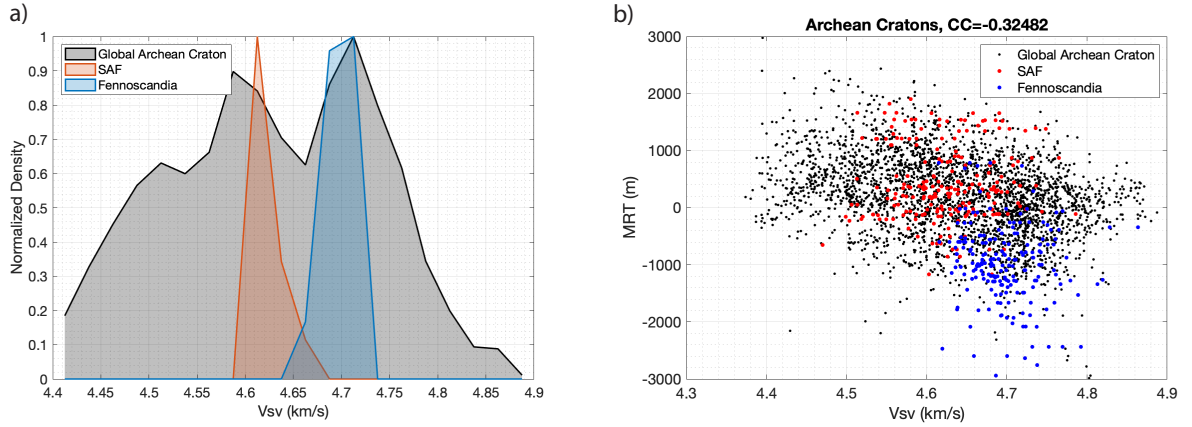
These observations are broadly consistent with Pedersen et al. (2009), in that they found similar  $V_{sv}$  in the shallowest mantle but lower  $V_{sv}$  in the Kaapvaal craton than the Fennoscandia craton (by  $\sim 0.1$  km/s) at greater depths, between 150 and 200 km. Their model suggested even lower velocities at greater depths than observed in this study. This discrepancy likely arises from the limited long-period ( $\leq 120$  s) constraints in the southern Africa phase velocities and the influence of the strong PVG at  $\sim 210$  km in the southern Africa receiver functions. The  $V_{sv}$  structure above  $\sim 210$  km in our model is therefore better constrained than in Pedersen et al. (2009). The presence of NVGs near  $\sim 150$  km and reduced velocities below this depth in the southern African craton has also been reported in previous studies as noted in section 1. Early 1D models identified a low-

velocity zone deeper than ~150 km (Y. Wang et al. 2008; Priestley, K. 1999). Savage and Silver (2008), using Sp and Ps receiver functions, reported velocity reductions near 150 km in the northeastern Kalahari craton, which they interpreted as lithospheric refertilization due to melt or metasomatic infiltration. Similarly, Li and Burke (2006) observed a low-velocity zone (~4.5 km/s) between 160 and 260 km, attributing it to elevated temperatures that may contribute to the high topography of southern Africa. In section 4.2, we further discuss the implications of these low velocities below ~150 km in southern Africa and their significance for global craton evolution.

## **4.2 Thermal modification of the upper mantle in the southern Africa cratons and implications for global Archean cratons**

The similar uppermost mantle V<sub>sv</sub> profiles in Fennoscandia and southern Africa may reflect a common depleted peridotite composition, combined with the effects of radial anisotropy and presence of low-velocity carbonate or hydrous minerals introduced through metasomatism (e.g. Huang et al. 2024). However, the lower-velocity layer (slower by ~0.1 km/s) below ~150 km in the southern Africa craton, compared to the Fennoscandian craton interior (which we consider a benchmark for intact cratonic lithosphere) likely reflects thermal perturbation of the deeper southern Africa lithosphere, as compositional variations alone are insufficient to produce a V<sub>sv</sub> reduction of ~0.1 km/s (Garber et al. 2018). This interpretation is consistent with previous studies suggesting that elevated temperature and/or reduced density in the upper mantle contribute to the high topography of southern Africa (e.g., Li and Burke 2006; Y. Wang et al. 2008), as noted in section 1, although some studies argued that the observed velocity reduction is not

significantly different from that in other cratons (e.g., Adams and Nyblade 2011; White-Gaynor et al. 2020).



**Figure 4.8: Figure 8.** (a) Comparison of average Vsv (150-200 km) between global Archean craton and two regional cratons. Black= distribution of average Vsv from the SEMum2 model (French and Romanowicz 2014) for Archean cratons defined using CRUST1.0 model (Laske et al. 2013), excluding Fennoscandia and southern Africa. Red = southern Africa cratons (this study). Blue = Fennoscandian craton interior (Huang et al., 2024). (b) Correlation between average Vsv (150-200 km) and mantle residual topography. Black=global Archean cratons; red= southern Africa cratons (this study); blue= Fennoscandia craton interior (Huang et al., 2024).

The concept of thermally modified lithosphere beneath the southern African craton has been explored in several geodynamic models. Hu et al. (2018) proposed that the high topography of South America and Africa cratons results from lithospheric modification by mantle plumes, which also produced Cretaceous uplift and widespread kimberlite magmatism. Based on high seismic velocity observations and additional constraints, they suggested that dense lower lithosphere was removed from these cratons and replaced by newly formed lithosphere through subsequent cooling. Celli et al. (2020) argued that such lithospheric removal is permanent and irreversible, based on seismic evidence of low uppermost mantle velocities in some Archean cratons. Gernon et al. (2023) proposed that

continental rifting and breakup can trigger lithospheric keel removal and kimberlite magmatism through conductive instabilities propagating from rift zones into craton interiors. Although our  $V_{sv}$  model comparison cannot distinguish among these mechanisms, it provides clear evidence for thermal perturbation in the mantle beneath cratonic southern Africa at depths greater than 150 km.

Geochemical constraints further support the presence of mechanical layering and thermal modification in the southern African lithosphere. Mantle xenolith-derived geotherms show an inflection between ~140 and 180 km depth, separating low-temperature, coarse-grained depleted lithologies above from higher-temperature, more fertile and strongly sheared lithologies below (Carlson et al. 2005). These sheared peridotites are commonly interpreted as products of deformation associated with mantle upwelling or metasomatism (Ehrenberg 1979; Katayama et al. 2009), and are thought to represent a thermally disturbed transition layer overlying the asthenosphere (Eaton et al. 2009). This transition, observed at ~50 kbar in the Kaapvaal craton (Eaton et al. 2009; Tharimena et al. 2017), corresponds well with the NVG at ~150 km in our model, supporting the interpretation of thermal modification at this depth. In addition, the evolution of lithospheric geotherms during Cretaceous kimberlite magmatism also supports thermal modification of the upper mantle. Nimis et al. (2024) suggested that the Kaapvaal craton may have initially been characterized by a cold geotherm (surface heat flow  $< 35 \text{ mW/m}^2$ ) that evolved, along with mantle uplift, and thermal re-equilibration, toward a warmer geotherm ( $\sim 40 \text{ mW/m}^2$ ), based on xenolith P–T estimates and changes in Cr-saturated garnet depths with kimberlite age. This would indicate progressive thinning of the refractory lithospheric mantle. Additionally, abrupt shifts in the Nd

isotopic compositions of Kaapvaal xenoliths from values typical of the lithosphere to values expected for asthenospheric mantle at ~115 Ma suggest the removal of ~30–40 km of lithosphere during kimberlite emplacement (Gernon et al. 2023). These geochemical and thermal changes are consistent with the reduced velocities observed below ~150 km in our model.

Comparison of our results with global Archean cratons suggests that thermal modification is a widespread feature of craton evolution. Fig. 8 shows the distribution of average  $V_{sv}$  between 150 and 200 km depth from the SEMum2 model (French and Romanowicz 2014) for global Archean cratons, where we identify Archean crust using the CRUST1.0 model (Laske et al. 2013). Excluding southern Africa and Fennoscandia, the global Archean dataset shows two dominant  $V_{sv}$  peaks at ~4.6 km/s and ~4.7 km/s. Our southern Africa model and the Fennoscandian craton interior correspond well to these two peaks, respectively. Interpreting these peaks as thermally modified and intact cratons, the global distribution suggests two primary classes of Archean lithosphere: intact cratons with higher velocities (~4.7 km/s), such as the central Superior craton, São Francisco craton in southern America, central Siberia, and southwestern Australia; and thermally modified cratons with moderately reduced velocities (~4.6 km/s), such as the northeastern Superior craton. Even lower velocities may reflect strongly modified or destroyed cratons, such as the North China craton. Furthermore, the Fennoscandian craton edge model (Huang et al, 2024), including much of the Caledonides and northernmost Archean regions of Sweden and Finland, shows similar velocities to southern Africa above 200 km. In this tectonic environment, at the margins of the thick Fennoscandian cratonic lithosphere, edge-driven convection may be the dominant mode

of uppermost mantle modification. Therefore, the  $\sim 4.6$  km/s peak in global Archean  $V_{sv}$  likely reflects a range of thermal modification processes, including plume interaction and edge-driven convection.

Finally, we explore the relationship between thermal modification and topography in Archean cratons. Mantle residual topography is calculated by removing Airy isostatic contributions from observed topography (GEBCO; Weatherall et al. 2015), using crustal thickness from CRUST1.0 model (Laske et al. 2013), with an assumed density of  $2800 \text{ kg/m}^3$  for the crust and  $3300 \text{ kg/m}^3$  for the mantle. This approach removes the effect of crustal thickness variations on surface topographic variations so that the remaining topographic signal can be interpreted in terms of variations in the density of the shallow mantle. In Fig. 8b we compare mantle residual topography with  $V_{sv}$  averaged over the depth range 150–200 km. Under the first-order assumption that variations in  $V_{sv}$  reflect changes in mantle temperature, this comparison allows us to assess, in a qualitative way, whether elevated regions (high residual topography) are supported by anomalously warm upper mantle (low  $V_{sv}$ ), and vice versa.

In the southern African cratons, only a very weak negative correlation is observed between  $V_{sv}$  at 150–200 km and residual topography (Fig. 8b). However, this relationship is more clearly defined in the global dataset, which shows a moderate negative correlation (cross correlation coefficient  $CC = -0.32$ ). This moderate negative correlation is also present when data from the Fennoscandia craton are included. Regions with higher topography and lower  $V_{sv}$  at these depths include southern Africa, São Francisco craton and eastern Siberia, in contrast to the lower elevations and higher  $V_{sv}$  values of Fennoscandia. This relationship suggests that thermal modification of the

upper mantle contributes to the elevation of some Archean cratons. Although deeper mantle processes may also influence the surface via dynamic topography, several observations argue against a dominant role for deep mantle convection in controlling the surface topography of southern Africa. The mantle transition zone thickness beneath the Kaapvaal craton is close to the global average ( $246 \pm 7$  km; Sun et al. 2018), indicating limited temperature anomalies at these depths. In addition, the observed residual topography has a wavelength of  $<500$  km, whereas dynamic topography associated with deep mantle flow typically occurs at wavelengths  $>1000$  km (Hu et al. 2018). Therefore, these observations suggest that upper mantle thermal modification is an important contributor to the present-day topography of southern Africa and likely plays a broader role in shaping global Archean cratons.

## 5 Conclusions

We develop a 3D Vsv model of the southern Africa cratons using Bayesian joint inversion of fundamental Rayleigh wave phase velocities (25 - 120 s) and Sp receiver functions extracted from CCP stacks. Higher phase velocities at all periods are clearly observed in the Kalahari craton compared to the Rehoboth Provinces, Namaqua-Natal belt, and Cape Fold Belt. The Bushveld Complex has relatively lower phase velocities than other cratonic regions at periods shorter than 60s. Moho depths derived from Sp phases mainly fall within the 32–50 km range, with relatively thin crust in the central and southern Kaapvaal craton. Widespread negative velocity gradients (NVGs) from the Moho to  $\sim 130$  km are observed across most of the study region, and a positive velocity gradient (PVG) at  $\sim 210$  km is observed in a narrower but coherent region with NNE-



SSW orientation across the Limpopo Belt, Bushveld Complex, and central Kaapvaal Craton.

V<sub>sv</sub> models of the Kalahari craton show a layered seismic structure, characterized by a high-velocity lid ( $\sim 4.7$  km/s) at 100–150 km depth. Compared with the Fennoscandian craton interior, the southern African cratons show similar velocities in the shallow upper mantle above  $\sim 140$  km, but have a low-velocity zone ( $\sim 2\%$  reduction) beneath this layer, consistent with thermal perturbation of the deeper lithosphere. We also identify lithospheric thinning to  $\sim 150$  km beneath the Okavango Rift. Comparison with global Archean cratons (average V<sub>sv</sub> in the 150–200 km depth range) shows that both the Fennoscandian and southern African cratons fall within two distinct V<sub>sv</sub> peaks at  $\sim 4.65$  km/s and  $\sim 4.55$  km/s, respectively, representing intact and thermally modified lithosphere. A clear negative correlation between V<sub>sv</sub> (150–200 km) and mantle residual topography further suggests that upper mantle thermal state contributes to the elevation of Archean cratons. The southern African cratons therefore provide a key example of how thermal modification of cratonic lithosphere influences both its internal structure and surface uplift, with broader implications for global craton evolution.

## 6 References

- Abt, David L., Karen M. Fischer, Scott W. French, Heather A. Ford, Huaiyu Yuan, and Barbara Romanowicz. 2010. "North American Lithospheric Discontinuity Structure Imaged by *Ps* and *Sp* Receiver Functions." *Journal of Geophysical Research: Solid Earth* 115 (B9): 2009JB006914.  
<https://doi.org/10.1029/2009JB006914>.
- Adams, Aubreya, and Andrew Nyblade. 2011. "Shear Wave Velocity Structure of the Southern African Upper Mantle with Implications for the Uplift of Southern Africa: Upper Mantle Velocities in Southern Africa." *Geophysical Journal International* 186 (2): 808–24. <https://doi.org/10.1111/j.1365-246X.2011.05072.x>.
- Afonso, Juan C., Walid Ben-Mansour, Suzanne Y. O'Reilly, et al. 2022. "Thermochemical Structure and Evolution of Cratonic Lithosphere in Central and Southern Africa." *Nature Geoscience* 15 (5): 405–10.  
<https://doi.org/10.1038/s41561-022-00929-y>.
- Albuquerque Seismological Laboratory (ASL)/USGS. 1993. "Global Telemetered Seismograph Network (USAF/USGS)." International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/SN/GT>.
- Albuquerque Seismological Laboratory/USGS. 1988. "Global Seismograph Network (GSN - IRIS/USGS)." International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/SN/IU>.

Andy Nyblade. 2020. “AfricaArray - Namibia.” With Namibia Geological Survey, University of the Witwatersrand, and Pennsylvania State University. IRIS, December 31. SEED data, 500 MB. [https://doi.org/10.7914/SN/8A\\_2015](https://doi.org/10.7914/SN/8A_2015).

Andy Nyblade. 2023. “REU: Imaging the Bushveld Complex, South Africa.” International Federation of Digital Seismograph Networks, December 31. SEED data, 2000 MB. [https://doi.org/10.7914/SN/ZT\\_2015](https://doi.org/10.7914/SN/ZT_2015).

Botswana Geoscience Institute. 2001. “Botswana Seismological Network.” With i.e.a.m.fadel@utwente.nl, Utrecht University, and Botswana Geoscience Institute. International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/SN/BX>.

Boyce, Alistair, Thomas Bodin, Stéphanie Durand, Dorian Soergel, and Eric Debayle. 2024. “Seismic Evidence for Craton Formation by Underplating and Development of the MLD.” *Geophysical Research Letters* 51 (4): e2023GL106170. <https://doi.org/10.1029/2023GL106170>.

Brown, Sarah E., and Karen M. Fischer. 2025. “Investigating the Antarctic Lithosphere Through Sp Receiver Function Analysis.” *Geochemistry, Geophysics, Geosystems* 26 (10): e2025GC012268. <https://doi.org/10.1029/2025GC012268>.

Brunsvik, Brennan, Zachary Eilon, and Colton Lynner. 2024. “Plate-Scale Imaging of Eastern US Reveals Ancient and Ongoing Continental Deformation.” *Geophysical Research Letters* 51 (12): e2024GL109041. <https://doi.org/10.1029/2024GL109041>.

- Carlson, Richard W., D. Graham Pearson, and David E. James. 2005. “Physical, Chemical, and Chronological Characteristics of Continental Mantle.” *Reviews of Geophysics* 43 (1): 2004RG000156. <https://doi.org/10.1029/2004RG000156>.
- Carr, Steve A. B., and Tolulope Olugboji. 2024. “A Taxonomy of Upper-Mantle Stratification in the US.” *Journal of Geophysical Research: Solid Earth* 129 (5): e2024JB028781. <https://doi.org/10.1029/2024JB028781>.
- Celli, Nicolas Luca, Sergei Lebedev, Andrew J. Schaeffer, and Carmen Gaina. 2020. “African Cratonic Lithosphere Carved by Mantle Plumes.” *Nature Communications* 11 (1): 92. <https://doi.org/10.1038/s41467-019-13871-2>.
- Chen, Ling. 2017. “Layering of Subcontinental Lithospheric Mantle.” *Science Bulletin* 62 (14): 1030–34. <https://doi.org/10.1016/j.scib.2017.06.003>.
- Cornell, David H., Valby Van Schijndel, Olafur Ingolfsson, et al. 2011. “Evidence from Dwyka Tillite Cobbles of Archaean Basement beneath the Kalahari Sands of Southern Africa.” *Lithos* 125 (1–2): 482–502. <https://doi.org/10.1016/j.lithos.2011.03.006>.
- De Wit, Maarten J., Cornel E. J. De Ronde, Marian Tredoux, et al. 1992. “Formation of an Archaean Continent.” *Nature* 357 (6379): 553–62. <https://doi.org/10.1038/357553a0>.
- Delph, J. R., and R. C. Porter. 2014. “Crustal Structure beneath Southern Africa: Insight into How Tectonic Events Affect the Mohorovičić Discontinuity.” *Geophysical Journal International* 200 (1): 254–64. <https://doi.org/10.1093/gji/ggu376>.

- EarthScope Consortium. 1986. “Global Seismograph Network - II.” International Federation of Digital Seismograph Networks. SEED data.  
<https://doi.org/10.7914/SN/II>.
- Eaton, David W., Fiona Darbyshire, Rob L. Evans, Herman Grütter, Alan G. Jones, and Xiaohui Yuan. 2009. “The Elusive Lithosphere–Asthenosphere Boundary (LAB) beneath Cratons.” *Lithos* 109 (1–2): 1–22.  
<https://doi.org/10.1016/j.lithos.2008.05.009>.
- Eeken, Thomas, Saskia Goes, Helle A. Pedersen, Nicholas T. Arndt, and Pierre Bouilhol. 2018. “Seismic Evidence for Depth-Dependent Metasomatism in Cratons.” *Earth and Planetary Science Letters* 491 (June): 148–59.  
<https://doi.org/10.1016/j.epsl.2018.03.018>.
- Ehrenberg, Stephen N. 1979. “Garnetiferous Ultramafic Inclusions in Minette from the Navajo Volcanic Field.” In *The Mantle Sample: Inclusion in Kimberlites and Other Volcanics*, edited by F. R. Boyd and Henry O. A. Meyer. American Geophysical Union. <https://doi.org/10.1029/SP016p0330>.
- Eilon, Zachary, Karen M. Fischer, and Colleen A. Dalton. 2018. “An Adaptive Bayesian Inversion for Upper-Mantle Structure Using Surface Waves and Scattered Body Waves.” *Geophysical Journal International* 214 (1): 232–53.  
<https://doi.org/10.1093/gji/ggy137>.
- Fadel, Islam, Hanneke Paulssen, Mark van der Meijde, et al. 2020. “Crustal and Upper Mantle Shear Wave Velocity Structure of Botswana: The 3 April 2017 Central

- Botswana Earthquake Linked to the East African Rift System.” *Geophysical Research Letters* 47 (4): e2019GL085598.  
<https://doi.org/10.1029/2019GL085598>.
- Fadel, Islam, Mark Van Der Meijde, and Hanneke Paulssen. 2018. “Crustal Structure and Dynamics of Botswana.” *Journal of Geophysical Research: Solid Earth* 123 (12).  
<https://doi.org/10.1029/2018JB016190>.
- Fischer, Karen M., Catherine A. Rychert, Colleen A. Dalton, Meghan S. Miller, Caroline Beghein, and Derek L. Schutt. 2020. “A Comparison of Oceanic and Continental Mantle Lithosphere.” *Physics of the Earth and Planetary Interiors* 309 (December): 106600. <https://doi.org/10.1016/j.pepi.2020.106600>.
- Fouch, M. J. 2004. “Mantle Seismic Structure beneath the Kaapvaal and Zimbabwe Cratons.” *South African Journal of Geology* 107 (1–2): 33–44.  
<https://doi.org/10.2113/107.1-2.33>.
- Frank Vernon, David James, and Paul Silver. 1998. “Broad Band Array Study in South Africa.” International Federation of Digital Seismograph Networks. SEED data.  
[https://doi.org/10.7914/SN/YB\\_1998](https://doi.org/10.7914/SN/YB_1998).
- French, S. W., and B. A. Romanowicz. 2014. “Whole-Mantle Radially Anisotropic Shear Velocity Structure from Spectral-Element Waveform Tomography.” *Geophysical Journal International* 199 (3): 1303–27. <https://doi.org/10.1093/gji/ggu334>.

Freybourger, Marion, James B. Gaherty, Thomas H. Jordan, and Kaapvaal Seismic Group.

2001. "Structure of the Kaapvaal Craton from Surface Waves." *Geophysical Research Letters* 28 (13): 2489–92. <https://doi.org/10.1029/2000GL012436>.

Gama, Isabella, Karen M. Fischer, Colleen A. Dalton, and Zachary Eilon. 2022.

"Variations in Lithospheric Thickness Across the Denali Fault and in Northern Alaska." *Geophysical Research Letters* 49 (24): e2022GL101256. <https://doi.org/10.1029/2022GL101256>.

Garber, Joshua M., Satish Maurya, Jean-Alexis Hernandez, et al. 2018.

"Multidisciplinary Constraints on the Abundance of Diamond and Eclogite in the Cratonic Lithosphere." *Geochemistry, Geophysics, Geosystems* 19 (7): 2062–86. <https://doi.org/10.1029/2018GC007534>.

GEOFON Data Centre. 1993. "GEOFON Seismic Network." With GEOFON Data

Centre, [geofon@gfz.de](mailto:geofon@gfz.de), and Deutsches GeoForschungsZentrum GFZ. GFZ Data Services. ...Mseed, XML, Approximately 80 active stations; greater than 440MB/day. <https://doi.org/10.14470/TR560404>.

George Helffrich, and J. F. B. D. Fonseca. 2011. "Mozambique Rift Tomography." With

National Directorate of Geology, Council for Geoscience, and University of Lisbon. International Federation of Digital Seismograph Networks. SEED data, 2000 MB. [https://doi.org/10.7914/SN/6H\\_2011](https://doi.org/10.7914/SN/6H_2011).

- Gernon, Thomas M., Stephen M. Jones, Sascha Brune, et al. 2023. “Rift-Induced Disruption of Cratonic Keels Drives Kimberlite Volcanism.” *Nature* 620 (7973): 344–50. <https://doi.org/10.1038/s41586-023-06193-3>.
- Golos, E. M., B. Brunsvik, Z. Eilon, K. M. Fischer, J. Byrnes, and J. Gaherty. 2024. “A New View of Shear Wavespeed and the Lithosphere-Asthenosphere Boundary in the Southwestern United States.” *Journal of Geophysical Research: Solid Earth* 129 (8): e2024JB029220. <https://doi.org/10.1029/2024JB029220>.
- Green, H. W., and Y. Gueguen. 1974. “Origin of Kimberlite Pipes by Diapiric Upwelling in the Upper Mantle.” *Nature* 249 (5458): 617–20. <https://doi.org/10.1038/249617a0>.
- Hansen, Samantha E., Andrew A. Nyblade, Jordi Julià, Paul H. G. M. Dirks, and Raymond J. Durrheim. 2009. “Upper-Mantle Low-Velocity Zone Structure beneath the Kaapvaal Craton from *S*-Wave Receiver Functions.” *Geophysical Journal International* 178 (2): 1021–27. <https://doi.org/10.1111/j.1365-246X.2009.04178.x>.
- Hanson, R. E. 2003. “Proterozoic Geochronology and Tectonic Evolution of Southern Africa.” *Geological Society, London, Special Publications* 206 (1): 427–63. <https://doi.org/10.1144/GSL.SP.2003.206.01.20>.
- Hastings, W. K. 1970. “Monte Carlo Sampling Methods Using Markov Chains and Their Applications.” *Biometrika* 57 (1): 97–109. <https://doi.org/10.1093/biomet/57.1.97>.



- Hirsch, Aaron C., Colleen A. Dalton, and Jeroen Ritsema. 2015. “Constraints on Shear Velocity in the Cratonic Upper Mantle from R Ayleigh Wave Phase Velocity.” *Geochemistry, Geophysics, Geosystems* 16 (11): 3982–4005.  
<https://doi.org/10.1002/2015GC006066>.
- Hu, Jiashun, Lijun Liu, Manuele Faccenda, et al. 2018. “Modification of the Western Gondwana Craton by Plume–Lithosphere Interaction.” *Nature Geoscience* 11 (3): 203–10. <https://doi.org/10.1038/s41561-018-0064-1>.
- Hua, J., K. M. Fischer, M. Wu, and N. A. Blom. 2020. “New Approaches to Multifrequency *Sp* Stacking Tested in the Anatolian Region.” *Journal of Geophysical Research: Solid Earth* 125 (11): e2020JB020313.  
<https://doi.org/10.1029/2020JB020313>.
- Huang, Yiran, Colleen A. Dalton, and Karen M. Fischer. 2024. “Investigating Layered Cratonic Lithosphere in the Fennoscandian Shield.” 2024 (December): T23D-07.  
<https://ui.adsabs.harvard.edu/abs/2024AGUFMT23D...07H>.
- Instituto Nacional de Minas. 2025. “National Seismic Network of Mozambique.” International Federation of Digital Seismograph Networks, November 30. SEED data. <https://doi.org/10.7914/ZD74-X973>.
- Jin, Ge, and James B. Gaherty. 2015. “Surface Wave Phase-Velocity Tomography Based on Multichannel Cross-Correlation.” *Geophysical Journal International* 201 (3): 1383–98. <https://doi.org/10.1093/gji/ggv079>.

- Katayama, Ikuo, Yuka Suyama, Jun-ichi Ando, and Tsuyoshi Komiya. 2009. “Mineral Chemistry and P–T Condition of Granular and Sheared Peridotite Xenoliths from Kimberley, South Africa: Origin of the Textural Variation in the Cratonic Mantle.” *Lithos* 109 (3–4): 333–40. <https://doi.org/10.1016/j.lithos.2008.05.004>.
- Krueger, Hannah E., Isabella Gama, and Karen M. Fischer. 2021a. “Global Patterns in Cratonic Mid-Lithospheric Discontinuities From Sp Receiver Functions.” *Geochemistry, Geophysics, Geosystems* 22 (6): e2021GC009819. <https://doi.org/10.1029/2021GC009819>.
- Krueger, Hannah E., Isabella Gama, and Karen M. Fischer. 2021b. “Global Patterns in Cratonic Mid-Lithospheric Discontinuities From Sp Receiver Functions.” *Geochemistry, Geophysics, Geosystems* 22 (6): e2021GC009819. <https://doi.org/10.1029/2021GC009819>.
- Laske, Gabi, Guy Masters, Zhitu Ma, and Mike Pasyanos. 2013. “Update on CRUST1.0 - A 1-Degree Global Model of Earth’s Crust.” April 1, EGU2013-2658. <https://ui.adsabs.harvard.edu/abs/2013EGUGA..15.2658L>.
- Lee, Cin-Ty A., Peter Luffi, and Emily J. Chin. 2011. “Building and Destroying Continental Mantle.” *Annual Review of Earth and Planetary Sciences* 39 (1): 59–90. <https://doi.org/10.1146/annurev-earth-040610-133505>.
- Legre, Jean-Joel, and Tolulope Olugboji. 2025. “The Origins of Africa’s Upper Mantle Discontinuities.” *Geochemistry, Geophysics, Geosystems* 26 (7): e2025GC012315. <https://doi.org/10.1029/2025GC012315>.

- Lei, Wenjie, Youyi Ruan, Ebru Bozdağ, et al. 2020. “Global Adjoint Tomography—Model GLAD-M25.” *Geophysical Journal International* 223 (1): 1–21.  
<https://doi.org/10.1093/gji/ggaa253>.
- Li, Aibing. 2011. “Shear Wave Model of Southern Africa from Regional Rayleigh Wave Tomography with 2-D Sensitivity Kernels: Shear Wave Model of Southern Africa.” *Geophysical Journal International* 185 (2): 832–44.  
<https://doi.org/10.1111/j.1365-246X.2011.04971.x>.
- Li, Aibing, and Kevin Burke. 2006. “Upper Mantle Structure of Southern Africa from Rayleigh Wave Tomography.” *Journal of Geophysical Research: Solid Earth* 111 (B10): 2006JB004321. <https://doi.org/10.1029/2006JB004321>.
- Ligorria, Juan Pablo, and Charles J. Ammon. 1999. “Iterative Deconvolution and Receiver-Function Estimation.” *Bulletin of the Seismological Society of America* 89 (5): 1395–400. <https://doi.org/10.1785/BSSA0890051395>.
- Lin, Fan-Chi, Michael H. Ritzwoller, and Roel Snieder. 2009. “Eikonal Tomography: Surface Wave Tomography by Phase Front Tracking across a Regional Broad-Band Seismic Array.” *Geophysical Journal International* 177 (3): 1091–110.  
<https://doi.org/10.1111/j.1365-246X.2009.04105.x>.
- Lithgow-Bertelloni, Carolina, and Paul G. Silver. 1998. “Dynamic Topography, Plate Driving Forces and the African Superswell.” *Nature* 395 (6699): 269–72.  
<https://doi.org/10.1038/26212>.

Liu, Lin, Simon L. Klemperer, and Alexander R. Blanchette. 2021. “Western Gondwana Imaged by S Receiver-Functions (SRF): New Results on Moho, MLD (Mid-Lithospheric Discontinuity) and LAB (Lithosphere-Asthenosphere Boundary).” *Gondwana Research* 96 (August): 206–18.  
<https://doi.org/10.1016/j.gr.2021.04.009>.

Liu, Tianze, Emily J. Chin, and Peter Shearer. 2023. “Strong Physical Contrasts Across Two Mid-Lithosphere Discontinuities Beneath the Northwestern United States: Evidence for Cratonic Mantle Metasomatism.” *AGU Advances* 4 (6): e2023AV001014. <https://doi.org/10.1029/2023AV001014>.

Liu, Tianze, and Peter M. Shearer. 2021. “Complicated Lithospheric Structure Beneath the Contiguous US Revealed by Teleseismic S-Reflections.” *Journal of Geophysical Research: Solid Earth* 126 (5): e2020JB021624.  
<https://doi.org/10.1029/2020JB021624>.

Marie Larson, Angela, J. Arthur Snoke, and David E. James. 2006. “S -Wave Velocity Structure, Mantle Xenoliths and the Upper Mantle beneath the Kaapvaal Craton.” *Geophysical Journal International* 167 (1): 171–86.  
<https://doi.org/10.1111/j.1365-246X.2006.03005.x>.

McCourt, S. 2001. “SHRIMP U-Pb Zircon Geochronology of the Hurungwe Granite Northwest Zimbabwe: Age Constraints on the Timing of the Magondi Orogeny and Implications for the Correlation between the Kheis and Magondi Belts.” *South African Journal of Geology* 104 (1): 39–46.  
<https://doi.org/10.2113/104.1.39>.

- Michael Wyssession, Douglas Wiens, and Andy Nyblade. 2015. “INVESTIGATION OF SOURCES OF INTRAPLATE VOLCANISM USING PASSCAL BROADBAND INSTRUMENTS IN MADAGASCAR, THE COMORES, AND MOZAMBIQUE.” International Federation of Digital Seismograph Networks, October 1. SEED data, 200000 MB. [https://doi.org/10.7914/SN/XV\\_2011](https://doi.org/10.7914/SN/XV_2011).
- Modisi, M. P., E. A. Atekwana, A. B. Kampunzu, and T. H. Ngwisanyi. 2000. “Rift Kinematics during the Incipient Stages of Continental Extension: Evidence from the Nascent Okavango Rift Basin, Northwest Botswana.” *Geology* 28 (10): 939. [https://doi.org/10.1130/0091-7613\(2000\)28%3C939:RKDTIS%3E2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28%3C939:RKDTIS%3E2.0.CO;2).
- Moorkamp, Max, Stewart Fishwick, Richard J. Walker, and Alan G. Jones. 2019. “Geophysical Evidence for Crustal and Mantle Weak Zones Controlling Intra-Plate Seismicity – the 2017 Botswana Earthquake Sequence.” *Earth and Planetary Science Letters* 506 (January): 175–83. <https://doi.org/10.1016/j.epsl.2018.10.048>.
- Mpuang, Admore Phindani, Takuo Shibutani, and Onkgopotse Ntibinyane. 2023. “New Insights into the Tectonics and Geodynamics of Botswana from Intracrustal Structures Revealed by Receiver Function Analysis.” *Geophysical Journal International* 235 (2): 1086–101. <https://doi.org/10.1093/gji/ggad291>.
- Nair, Shaji K., Stephen S. Gao, Kelly H. Liu, and Paul G. Silver. 2006. “Southern African Crustal Evolution and Composition: Constraints from Receiver Function Studies.” *Journal of Geophysical Research: Solid Earth* 111 (B2): 2005JB003802. <https://doi.org/10.1029/2005JB003802>.

- Nguuri, T. K., J. Gore, D. E. James, et al. 2001. “Crustal Structure beneath Southern Africa and Its Implications for the Formation and Evolution of the Kaapvaal and Zimbabwe Cratons.” *Geophysical Research Letters* 28 (13): 2501–4.  
<https://doi.org/10.1029/2000GL012587>.
- Nimis, Paolo, Herman Grütter, and Fabrizio Nesola. 2024. “Multistage Diamond Formation, Mantle Uplift and Changing Geothermal Regimes Recorded by Inclusions in Kimberley Diamonds.” Paper presented at International Kimberlite Conference. *International Kimberlite Conference Extended Abstracts: 2024*, July 8. <https://doi.org/10.29173/ikc4085>.
- Nyblade, Andrew A., and Norman H. Sleep. 2003. “Long Lasting Epeirogenic Uplift from Mantle Plumes and the Origin of the Southern African Plateau.” *Geochemistry, Geophysics, Geosystems* 4 (12): 2003GC000573.  
<https://doi.org/10.1029/2003GC000573>.
- Olugboji, Tolulope, Ziqi Zhang, Steve Carr, Canberk Ekmekci, and Mujdat Cetin. 2023. “On the Detection of Upper Mantle Discontinuities with Radon-Transformed Receiver Functions (CRISP-RF).” *Geophysical Journal International* 236 (2): 748–63. <https://doi.org/10.1093/gji/ggad447>.
- Paul Silver. 2001. “Anatomy of an Archean Craton, South Africa, A Multidisciplinary Experiment across the Kaapvaal Craton.” International Federation of Digital Seismograph Networks, July 11. SEED data.  
[https://doi.org/10.7914/SN/XA\\_1997](https://doi.org/10.7914/SN/XA_1997).

- Pedersen, H. A., S. Fishwick, and D. B. Snyder. 2009. "A Comparison of Cratonic Roots through Consistent Analysis of Seismic Surface Waves." *Lithos* 109 (1–2): 81–95. <https://doi.org/10.1016/j.lithos.2008.09.016>.
- Penn State University. 2009. "AfricaArray." International Federation of Digital Seismograph Networks, October 22. SEED data. <https://doi.org/10.7914/SN/AF>.
- Priestley, K. 1999. "Velocity Structure of the Continental Upper Mantle: Evidence from Southern Africa." In *Developments in Geotectonics*, vol. 24. Elsevier. [https://doi.org/10.1016/S0419-0254\(99\)80004-4](https://doi.org/10.1016/S0419-0254(99)80004-4).
- Ravenna, Matteo, Sergei Lebedev, Javier Fullea, and Joanne M. -C. Adam. 2018. "Shear-Wave Velocity Structure of Southern Africa's Lithosphere: Variations in the Thickness and Composition of Cratons and Their Effect on Topography." *Geochemistry, Geophysics, Geosystems* 19 (5): 1499–518. <https://doi.org/10.1029/2017GC007399>.
- Rychert, Catherine A., and Peter M. Shearer. 2009. "A Global View of the Lithosphere-Asthenosphere Boundary." *Science* 324 (5926): 495–98. <https://doi.org/10.1126/science.1169754>.
- Savage, Brian, and Paul G. Silver. 2008. "Evidence for a Compositional Boundary within the Lithospheric Mantle beneath the Kalahari Craton from S Receiver Functions." *Earth and Planetary Science Letters* 272 (3–4): 600–609. <https://doi.org/10.1016/j.epsl.2008.05.026>.

- Sodoudi, Forough, Xiaohui Yuan, Rainer Kind, et al. 2013. “Seismic Evidence for Stratification in Composition and Anisotropic Fabric within the Thick Lithosphere of Kalahari Craton.” *Geochemistry, Geophysics, Geosystems* 14 (12): 5393–412. <https://doi.org/10.1002/2013GC004955>.
- Steve Gao, Kelly Liu, Mohamed Abdelsalam, and John Hogan. 2016. “Seismic Arrays for African Rift Initiation.” International Federation of Digital Seismograph Networks, August 10. SEED data. [https://doi.org/10.7914/SN/XK\\_2012](https://doi.org/10.7914/SN/XK_2012).
- Sun, Muchen, Xiaofei Fu, Kelly H. Liu, and Stephen S. Gao. 2018. “Absence of Thermal Influence from the African Superswell and Cratonic Keels on the Mantle Transition Zone beneath Southern Africa: Evidence from Receiver Function Imaging.” *Earth and Planetary Science Letters* 503 (December): 108–17. <https://doi.org/10.1016/j.epsl.2018.09.012>.
- Tedla, Getachew E., M. Van Der Meijde, A. A. Nyblade, and F. D. Van Der Meer. 2011. “A Crustal Thickness Map of Africa Derived from a Global Gravity Field Model Using Euler Deconvolution: Map of Africa Using Euler Deconvolution.” *Geophysical Journal International* 187 (1): 1–9. <https://doi.org/10.1111/j.1365-246X.2011.05140.x>.
- Tharimena, Saikiran, Catherine Rychert, and Nicholas Harmon. 2017. “A Unified Continental Thickness from Seismology and Diamonds Suggests a Melt-Defined Plate.” *Science* 357 (6351): 580–83. <https://doi.org/10.1126/science.aan0741>.



- Utrecht University (UU Netherlands). 1983. “NARS.” International Federation of Digital Seismograph Networks. SEED data. <https://doi.org/10.7914/SN/NR>.
- Wang, Tuo, Stephen S. Gao, Yuhang Dai, Qiuyue Yang, and Kelly H. Liu. 2019. “Lithospheric Structure and Evolution of Southern Africa: Constraints From Joint Inversion of Rayleigh Wave Dispersion and Receiver Functions.” *Geochemistry, Geophysics, Geosystems* 20 (7): 3311–27. <https://doi.org/10.1029/2019GC008259>.
- Wang, Yi, Lianxing Wen, and Donald Weidner. 2008. “Upper Mantle SH- and P-Velocity Structures and Compositional Models beneath Southern Africa.” *Earth and Planetary Science Letters* 267 (3–4): 596–608. <https://doi.org/10.1016/j.epsl.2007.12.010>.
- Weatherall, Pauline, K. M. Marks, Martin Jakobsson, et al. 2015. “A New Digital Bathymetric Model of the World’s Oceans.” *Earth and Space Science* 2 (8): 331–45. <https://doi.org/10.1002/2015EA000107>.
- White-Gaynor, A. L., A. A. Nyblade, R. Durrheim, et al. 2020. “Lithospheric Boundaries and Upper Mantle Structure Beneath Southern Africa Imaged by *P* and *S* Wave Velocity Models.” *Geochemistry, Geophysics, Geosystems* 21 (10): e2020GC008925. <https://doi.org/10.1029/2020GC008925>.
- Wilson, J. F. 1990. “A Craton and Its Cracks: Some of the Behaviour of the Zimbabwe Block from the Late Archaean to the Mesozoic in Response to Horizontal Movements, and the Significance of Some of Its Mafic Dyke Fracture Patterns.”

*Journal of African Earth Sciences (and the Middle East)* 10 (3): 483–501.

[https://doi.org/10.1016/0899-5362\(90\)90101-J](https://doi.org/10.1016/0899-5362(90)90101-J).

Wittlinger, Gérard, and Véronique Farra. 2007. “Converted Waves Reveal a Thick and Layered Tectosphere beneath the Kalahari Super-Craton.” *Earth and Planetary Science Letters* 254 (3–4): 404–15. <https://doi.org/10.1016/j.epsl.2006.11.048>.

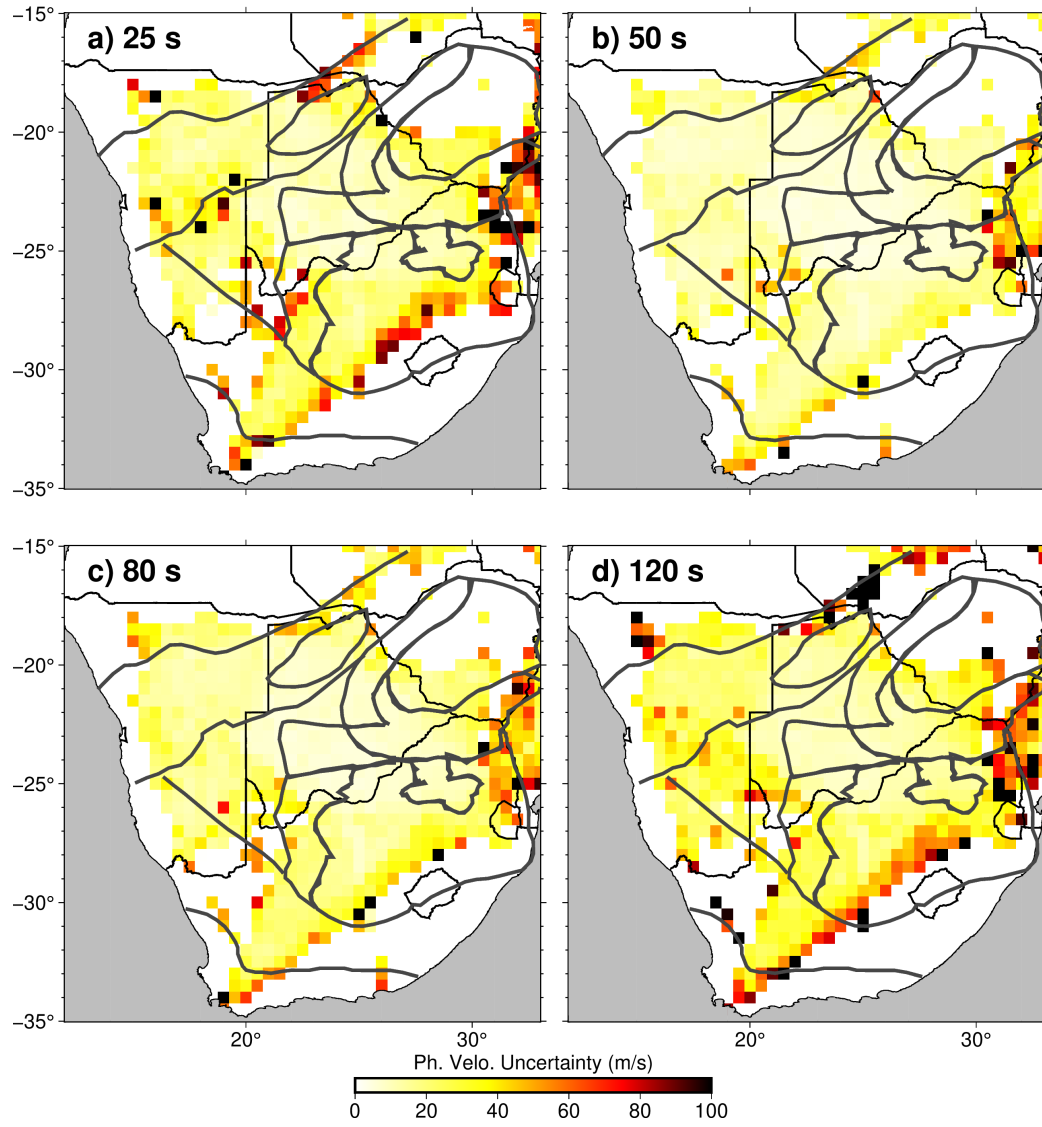
Yang, Yingjie, Aibing Li, and Michael H. Ritzwoller. 2008. “Crustal and Uppermost Mantle Structure in Southern Africa Revealed from Ambient Noise and Teleseismic Tomography.” *Geophysical Journal International* 174 (1): 235–48. <https://doi.org/10.1111/j.1365-246X.2008.03779.x>.

Youssof, M., H. Thybo, I. M. Artemieva, and A. Levander. 2013. “Moho Depth and Crustal Composition in Southern Africa.” *Tectonophysics* 609 (December): 267–87. <https://doi.org/10.1016/j.tecto.2013.09.001>.

Youssof, M., H. Thybo, I. M. Artemieva, and A. Levander. 2015. “Upper Mantle Structure beneath Southern African Cratons from Seismic Finite-Frequency P- and S-Body Wave Tomography.” *Earth and Planetary Science Letters* 420 (June): 174–86. <https://doi.org/10.1016/j.epsl.2015.01.034>.

Yu, Youqiang, Kelly H. Liu, Zhouchuan Huang, et al. 2017. “Mantle Structure beneath the Incipient Okavango Rift Zone in Southern Africa.” *Geosphere* 13 (1): 102–11. <https://doi.org/10.1130/GES01331.1>.

## 7 Supplemental Material



*Figure 4.9: Figure S1. Eikonal Rayleigh wave phase velocity uncertainty maps at (a) 25 s, (b) 50 s, (c) 80 s, and (d) 120 s.*

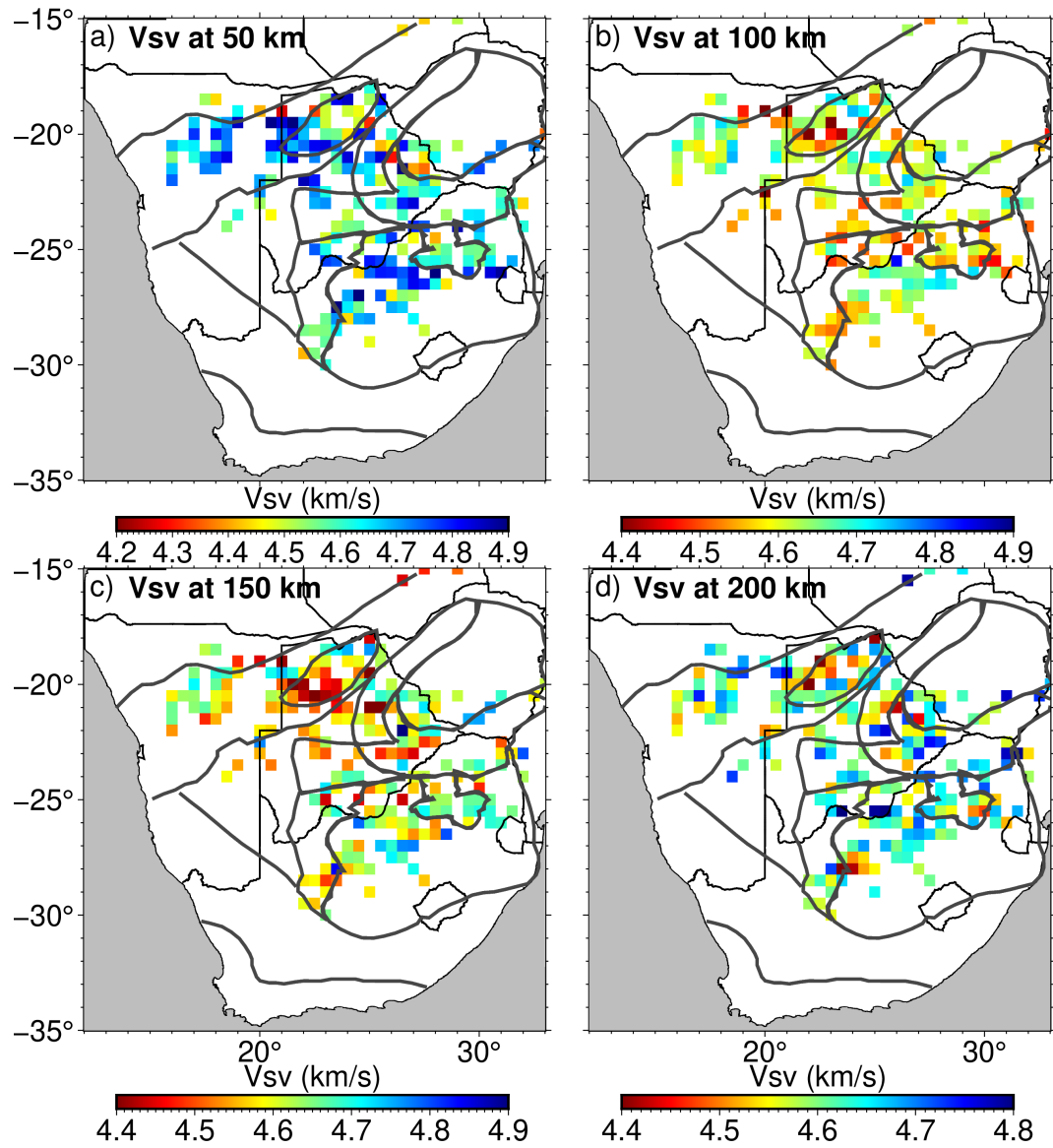
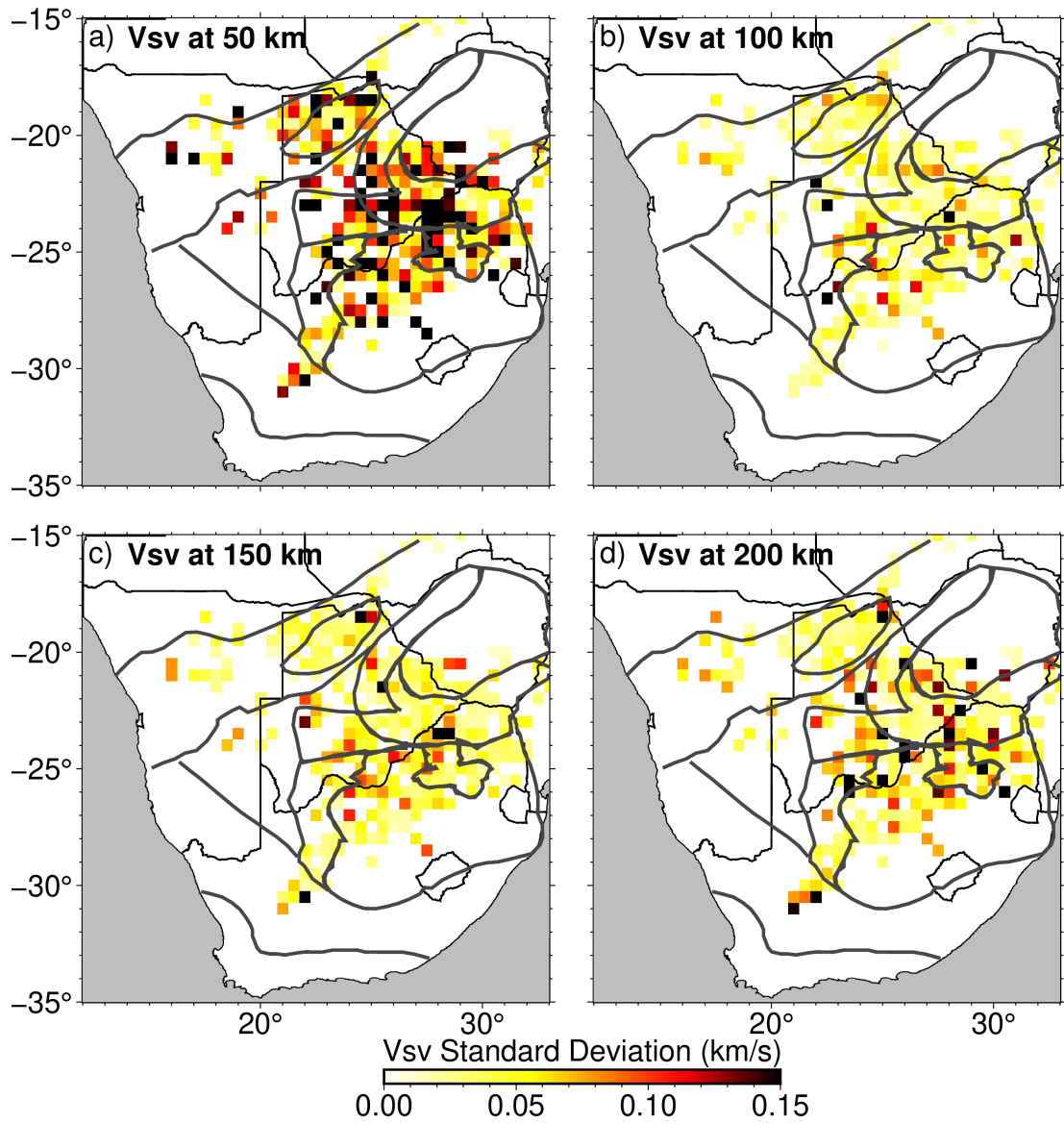
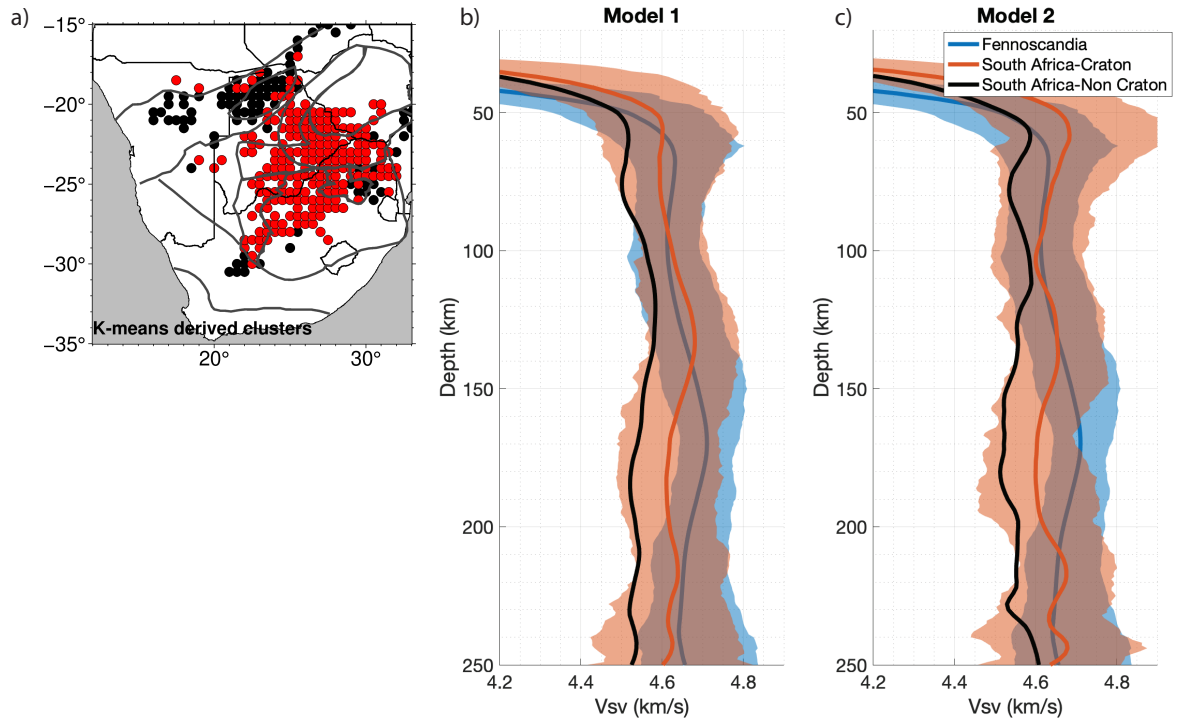


Figure 4.10: **Figure S2.** Map views of horizontal  $V_{sv}$  slices from joint inversion Model 2 (with relatively higher weight assigned to Sp receiver functions) at depths of (a) 50 km, (b) 100 km, (c) 150 km, (d) 200 km.



*Figure 4.11: Figure S3.* Map views of horizontal Vsv uncertainty slices extracted from joint inversion Model 1 at depths of (a) 50 km, (b) 100 km, (c) 150 km, (d) 200 km.



*Figure 4.12: Figure S4.* (a) Spatial distribution of two clusters obtained from k-means classification of 1-D Vsv Model 1 between 50 km and 250 km depth. (b) Comparison of Vsv profiles: black = southern Africa non-craton (Model 1), red = southern Africa cratons (Model 1), blue = Fennoscandian craton interior. Solid lines indicate average Vsv, and shaded regions show the distribution of models. (c) Same as (b), but for southern Africa Vsv from Model 2.

# **APPENDIX| Estimating the Local Amplification of Rayleigh Waves with Regional Broadband Seismic Arrays**

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## Abstract

Because of the different characteristics of seismic velocity and density sensitivity kernels for Rayleigh wave amplification and phase velocity, combining observations of surface-wave amplification with surface-wave phase could help to constrain seismic models. We are developing a new method to determine the local amplification of the Rayleigh wave on a small regional scale. The amplitudes of 20-100 s Rayleigh waves are extracted from 677 teleseismic earthquakes measured at the SVEKALAPKO, LAPNET, and ScanArray with other regional networks in Fennoscandia. We use the amplitude ratio approach (Eddy and Ekström, 2014) to isolate the receiver effect from the source, focusing, and attenuation effects. In order to separate the effects of local elastic structure from erroneous instrument response in the amplification values, we firstly use three different references to identify obviously constant and time-dependent erroneous instrument response. Then by comparing amplitudes from every two nearby stations, tinier time-dependent erroneous instrument responses will be removed. Lastly, outlier amplitude data, amplitude ratio data, and stations with larger deviations from median values are removed. Because seismic networks did not overlap in time, we force the amplification of one or more geographically adjacent station pairs from time-non-overlapping networks with the same value to get a coherent set of relative local amplification for the entire study area. The final observed local Rayleigh wave amplification factors at eight periods correlate well with some known main tectonic and geologic features and agree with two predictions from regional models in Fennoscandia.



# 1 Introduction

Compared with phase, amplitude, as another observable from surface waves, is less used in developing seismic models. While surface wave amplitude carries a wealth of information from the source, path, and receiver effects (Ferreira & Woodhouse 2006). Amplitude is affected by propagation through attenuation and elastic focusing. Previous studies used surface wave amplitude measurements to image both intrinsic attenuation (e.g., Dalton & Ekström 2006b) and elastic structures (Dalton & Ekström 2006a). Receiver effects include instrument response and local elastic structure beneath stations, which is studied by data products of surface wave amplitude, such as Rayleigh-wave ellipticity (e.g., Lin et al. 2012a) and local site amplification (e.g., Lin et al. 2012b; Schardong et al. 2019). Because of the different characteristics of phase velocity and local site amplification sensitivity kernels to  $V_p$ ,  $V_s$ , and density, local site amplification is believed to be more sensitive to shallow  $V_p/V_s$  ratio and density structure with high depth resolution than the phase velocity (Eddy & Ekström 2014; Schardong et al. 2019; Lin et al. 2012b), and consequently an independent and complementary tool with phase velocity to constrain elastic structure better.

However, there are two main challenges to making use of local site amplification measured from amplitude data. The first challenge is problematic stations with wrong instrument gain. Both time-dependent and constant for the entire operational period calibration errors at seismic stations are observed in previous studies (e.g., Ekström et al. 2006; Ma et al. 2016). The absolute instrument gain is difficult to measure and correct,

which will cause significant deviations in measured surface wave amplitudes and deduced local site amplification.

The second challenge is, with well-calibrated instruments, to separate the receiver effect from other contributing effects in amplitude data. There are two approaches to obtaining local site amplifications. Lin et al. (2012b) used wavefront tracking to separate attenuation and local site amplification, accounting for focusing and defocusing effects in Helmholtz tomography. Bao et al. (2016) also use this method to study attenuation structure beneath North America and Bowden et al. (2017) applied the Helmholtz tomography-based method to ambient noise at shorter periods. Eddy and Ekström (2014;2020) develop another method to isolate local site amplification is amplitude ratio approach, by calculating the amplitude ratios of two nearby stations to cancel out the other effects except for the receiver effect. The local site amplification maps measured by these studies show consistent features with known geologic and tectonic features at a large scale. Schardong et al. (2019) develop a 3-D shear wave velocity model inversed from sole amplification measured by amplitude ratio approach, which agrees with some previously reported features but has more detail features at a small scale, implying amplification has the potential to constrain the elastic structure of the crust and upper mantle better combining with phase velocity.

Estimating the local site surface wave amplification using the amplitude ratio approach with regional broadband seismic arrays faces other challenges. The first is the uncertainty of amplitude recorded by seismometers. Metadata issues, noisy site locations, and degrading instrument responses will introduce bias in the amplitude measurement of sensors (Ringler et al. 2021). Without instrument upgrades, investigation of instrument

gain, and yearly calibrations like the Global Seismographic Network (GSN), some temporary regional seismic arrays are likely to be exposed to measurement uncertainty. Secondly, the amplitude measurements of temporary regional seismic arrays may lack uniform event azimuthal coverage, which may cause the assumption of amplitude ratio approach, the cancellation of source and path effect, fails as discussed in Bao et al. (2016). Sparse deployment of regional seismic arrays can possibly lead to the failure of the assumption. In addition, due to the absence of long-term existing stations, regional temporary networks may bring trouble in the inversion of amplification because the amplitude ratio approach can only constrain relative amplification between time-overlapping stations.

In this study, we estimate the local amplification of Rayleigh waves with regional broadband seismic arrays in Fennoscandia. we apply the amplitude-ratio approach of Eddy and Ekström (2014; 2020) and develop a method to control data quality and generate the entail amplification maps by combining separate temporary networks. Additionally, we compare our amplification measurement with amplification predictions with local 3-D shear-wave velocity models.

## **2 Data and Methods**

We use waveforms from 283 broadband stations mainly from three temporary time-non-overlapping networks in Fennoscandia (Fig. A1a): SVEKALAPKO (ZB network), LAPNET (XK network), and ScanArray Core (1G network). In addition to them, we include some stations present between 2012 and 2017 from SCANLIPS3D (ZR network), NEONOR2 (2D network), NORSAR (NO network), and NNSN (NS network), which

together with ScanArray Core (1G) we refer to as 1G+ for the sake of convenience in the following text. The operation times of the temporary networks differ significantly (ZB: 1998.10-1999.02, XK: 2007-2009, 1G+: 2012-2019). Additionally, we include two permanent stations from the Global Seismographic Network (IU network code). We obtain data from 669 teleseismic earthquakes with magnitudes  $M_w > 6$  ( $M_w > 5.5$  for network ZB) and epicentral distances  $\leq 120^\circ$  (Fig. A1b) recorded by the three main networks during their respective operation times. We measure fundamental-mode Rayleigh wave using the Automated Surface Wave Measurement System (ASWMS; Jin and Gaherty, 2015) at eight periods between 20-100 s.

In order to isolate local amplification in amplitude data, we use the amplitude ratio approach developed by Eddy and Ekström (2014). With this approach, a measured amplitude is considered to be a product of four factors: earthquake source, receiver, focusing and geometric spreading during propagation, and attenuation during propagation. The receiver factor includes both local site amplification, which is the focus on this study, and the instrument response. For a given earthquake recorded by two spatially adjacent stations, because of the similar take-off angles and ray paths, the amplitude effects associated with the source and propagation can be assumed as the same. Therefore, taking the ratio of amplitudes from the same earthquake at two adjacent stations  $i$  and  $j$  can isolate the receiver factor because the ratio should cancel out the other three effects. For a given event  $k$ , logarithmic amplitude ratio  $d_{ij}^k(\omega)$  is calculated from the measured amplitudes  $A_i^k$  and  $A_j^k$  recorded by two neighboring stations  $i$  and  $j$  (within  $2^\circ$  in our dataset) at a given frequency  $\omega$ ,

$$d_{ij}^k(\omega) = \ln \left( \frac{A_i^k(\omega)}{A_j^k(\omega)} \right) = \ln \left( A_i^k(\omega) \right) - \ln \left( A_j^k(\omega) \right) \quad (1)$$

The average logarithmic ratio  $\bar{d}_{ij}(\omega)$  for each pair of stations  $i$  and  $j$  is computed from the average of logarithmic amplitude ratios of all  $N_e$  events recorded by the two stations. To determine local amplification  $A_{R,i}(\omega)$  at each station  $i$  from the average logarithmic ratio  $\bar{d}_{ij}(\omega)$ , a least-squares inversion is used to minimize the following misfit function weighted by the standard deviation  $\sigma_{ij}(\omega)$  of logarithmic amplitude ratios  $d_{ij}^k(\omega)$ .

$$\chi^2 = \sum_{ij} \frac{1}{\sigma_{ij}^2(\omega)} [(\ln(A_{R,i}(\omega)) - \ln(A_{R,j}(\omega)) - \bar{d}_{ij}(\omega))]^2 \quad (2)$$

From equation (2), the relative amplification  $A_{R,i}(\omega)$  can only be solved by logarithmic amplitude ratios  $d_{ij}^k(\omega)$ , which requires the station pairs should be overlapping in time. Separate seismic networks not existing at the same time is one of a challenge for the amplitude ratio approach as mentioned before, which will be discussed in Section 2.2. Receiver effects can include both the instrument response and the Earth structure beneath the station. Since we remove the instrument response from the waveform data prior to measuring amplitudes, Equations 1 and 2 should effectively isolate the local site amplification due to the structure beneath the receiver if the instrument is well calibrated—that is, if the reported response matches the actual response. However, errors in the reported response will bias estimates of local site amplification and any inferences about elastic structure drawn from them. Here we develop a method to identify instrument response problems and estimate more accurate local amplification with data selection using regional broadband seismic arrays in Fennoscandia. Fig A2 shows the main five steps we used to

deal with our dataset and generate local Rayleigh wave amplification factors. We use the amplitude data at 50 s as an example to illustrate the method to process data.

## 2.1 Amplitude data selection

Like what is found about amplitude data recorded by seismic stations in Ma et al. (2016) and Ekström et al. (2006), our data set also shows three types. Most of the stations with correct instrument responses can provide consistent amplitude data with others. Some stations may have incorrect instrument gain recording amplitude with a large deviation from others for their entire operational period. Other problematic stations behave time-dependent questionably, which means their recorded amplitude data systematically increase or decrease starting from a time, with obviously incorrect instrument response for a certain time period. We first perform amplitude data selection to identify and remove the amplitude data recorded by the stations with erroneous instrument responses.

If we set criteria of absolute amplitude to identify problematic amplitude data recorded by constant or time-dependent erroneous stations, floating absolute values due to source and propagation effect will make the constraint ineffective. In order to exclude the effect from source and propagation from absolute amplitude data and narrow the criteria of data selection as much as possible, we use three references to compute relative amplitude: amplitude of synthetic waveforms, fundamental-mode Rayleigh wave source excitation values, and amplitude of permanent stations. We generate three types of relative amplitude data by calculating the logarithmic ratio of observed amplitude data and respective references ( $\ln (A/A_{synthetics})$ ,  $\ln (A/A_{excitation})$ , and  $\ln (A/A_{KEV})$ ).

Synthetic seismograms are generated using normal-mode summation (Gilbert, 1971) for a 1-D earth model ATL2A (James et al. 2014). The MINEOS code (Masters et al. 2011) is used to calculate mode eigenfunctions and eigenfrequencies. Rayleigh wave amplitudes are measured by ASWMS (Jin and Gaherty, 2015) from synthetic waveforms of every event for each station. Using the amplitude of synthetic waveforms as a reference takes to source and attenuation in propagation into account and thus can constrain the relative amplitude to a narrower range, but it can only be used for the events with moment tensors.

Fundamental-mode Rayleigh wave source excitation values are calculated following Hariharan et al. (2022) for every event. Source effects are reflected in the source excitation values including radiation patterns and earthquake magnitudes. Because separating the source effect from the amplitude, using source excitation values can help constrain the range of amplitude data better than the absolute amplitude only, but it can also be used for earthquakes with moment tensors. Using Rayleigh wave amplitudes measurement from two high-quality permanent stations (IU.KONO and IU.KEV) as reference is the most maneuverable choice among these three, and it does not need to know the moment tensors of earthquakes. It can help eliminate the source and propagation effects especially when permanent stations are close to the other stations within the study region. However, the uncertainty of amplitude measurement of permanent stations may introduce errors in calculating relative amplitudes.

### ***2.1.1 Step 1: Use time series to delete obvious bad stations and time-dependent erroneous stations***

In the first step, we aim to identify obviously always and time-dependent erroneous instrument responses for every single station. The relative amplitude data of each event is

treated as a 2D variable including the event index and relative amplitude value. Using the difference between the 90th and 10th percentiles of relative amplitude in the dataset a given frequency  $\omega$  as a normalized number to normalize the relative amplitude of every station. We also normalize the event index of every station because the dataset sizes differ a lot for the stations from different networks and we expect to have uniform parameters when we identify outlier clusters in the following clustering processes.

We apply the density-based spatial clustering of applications with noise (DBSCAN) algorithm (Ester et al., 1996) to classify the normalized relative amplitude data with time information for every station. Moving median values with a moving 5-events-length window are calculated for the stability of the DBSCAN algorithm performance. The DBSCAN algorithm can separate the amplitude data into several time-continuous clusters, and we will identify the outlier clusters according to their median values. We set the median number of all relative amplitudes data at a given frequency  $\omega$  as a reference number and the standard deviation of relative amplitudes data as a limit number. If the median number of a given cluster is larger or smaller than one standard deviation than the reference number, we will tag the cluster as an outlier cluster and remove it. In Fig. A3, stations VRF and SA02 are classified as only one cluster with a median value within our acceptable amplitude range, so all events will be kept for these two stations. Although being separated into three clusters with some noise data points for station SA37, all three clusters have a lower median relative amplitude value than our acceptable range, which means the SA37 station is an “always-bad” station and will be removed in this step. Some stations, like station FF13, have a sudden and significant change in relative amplitude along the event index. DBSCAN algorithm can separate them into several clusters and remove the outlier



clusters whose median relative amplitudes are outside our acceptable range. Such kinds of stations are “time-dependent erroneous” stations.

Although with different relative values, the clustering results and estimation of outlier clusters using the three references are similar (Fig. A2). Permanent stations are not as so stable as the other two references because the access and measurement of all events amplitudes are not always available for one permanent station. We will use amplitudes from synthetic waveforms as an example in the following illustration.

***2.1.2 Step 2: Use every two adjacent stations to identify tinier time-dependent erroneous stations.***

After removing the obvious “always-bad” and “time-dependent erroneous” stations using relative amplitudes data with synthetic waveforms as references, we seek to identify tinier time-dependent erroneous stations which are difficult to be distinguished in step 1, by using amplitudes from other adjacent stations as references. Because the two neighboring stations are close enough, we should expect them to have the same amplitudes for the same event with well-calibrated instruments. A sensible time-dependent change in relative amplitudes of a station is more likely to be contributed by erroneous instrument response. For every given station, we compare its amplitudes with the amplitudes of other spatially nearby stations (within  $2^\circ$ ) manually. If more than one sensible time-dependent relative amplitude change is found, we will suppose this station with tinier time-dependent errors. Same as in step 1, choosing one of the nearby stations as a reference, we apply the DBSCAN algorithm to separate relative amplitude data into several clusters and remove the cluster with an absolute median relative amplitude larger than 0.1.

Fig. A4 shows an example of step 2. Relative to amplitude from synthetic waveforms, it is difficult to sense the change of amplitude for station ARC2. If we compare the amplitude of ARC2 with other nearby stations (ARA0, ARC7, and KEV), an apparent sharp change in relative amplitudes can be seen starting from the 582nd event systematically. Because the event cluster after the 582nd event has closer amplitude values with other stations, we will remove the first cluster which is more likely to include inaccurate instrument response. When the distance between two stations is larger and larger, the distribution of relative amplitude is more and more scattered. In this case, until the distance of  $\sim 0.5592^\circ$ , the relative amplitude is still concentrated enough to be separated as two clusters.

***2.1.3 Step 3: Use relative amplitudes to form acceptable value ranges and delete single outlier events and outlier stations.***

After removing obvious outlier stations and time-dependent outlier clusters, we use relative amplitudes compared with synthetic waveforms to remove outlier events and outlier stations further. For every single station, we set values, triple standard deviation away from median relative amplitudes, as thresholds and will reject the event outside this range. For all stations, the thresholds are set as triple standard deviation away from the median number of all stations' median relative amplitudes, and we will discard the stations with a median relative amplitude outside this range. The acceptable value range is narrowed after the data selection of steps 1 and 2. The numbers of amplitude data and stations involved in steps 1-3 using synthetic waveforms amplitudes as reference are shown in Table A1.

Following the event- and station-based data selection, we form a datum of station-pair observation using equation (1). Before calculating the average logarithmic ratio  $\bar{d}_{ij}(\omega)$  for each pair of stations  $i$  and  $j$ , we use the same strategy to select amplitude ratio data: setting one standard deviation away from median logarithmic amplitude ratios as an acceptable range and rejecting the ratios outside this limit. When calculating the misfit function in least-square inversion, we also use the reciprocal of standard deviation as a weight matrix to weigh every station-pair observation.

## 2.2 Combining separate networks

The amplitude ratio approach can only constrain relative amplification between stations because only the stations linked together by station-pair observations can be included in the inversion, which means this approach can only form local amplitudes factors with time-overlapping stations.

Since the seismic networks in our study did not overlap in time, a strategy must be developed in order to solve for relative amplification within the entire study region. The method we used is after inverting every subregional local amplification, we force the amplification of one or more geographically adjacent station pairs from time-non-overlapping networks with the same value to get a coherent set of relative local amplification for the entire study area.

When forcing the station pair with the same amplification factor, the choice of the station pair should be very cautious. Fig. A5a and b show the influence of different connecting station pairs on the whole amplification map. When we use stations SA10 and LP72 to link networks 1G+ and XK together, the amplification of stations from 1G+ will

be systematically smaller (Fig. A5a). However, if we use stations SA09 and LP75 to link networks 1G+ and XK together, the change of amplification factors from 1G+ to XK is smoother. Fortunately, permanent station KEV from the international network IU can connect the 1G+ and XK naturally (Fig. A5c). The amplification map connected by two nearby stations SA09 and LP75 has great consistency with the amplification map linked by KEV naturally, which means our strategy can work in generating an amplification map within the entire study region. The bias of the entire amplification map in Fig. A5a may be caused by the inaccurate amplification factor of station LP72, which has an abnormal value with other nearby stations from network XK. Except for connecting two separate networks by only linking one station pair, we also try more than one station pair. In Fig A5d, we connect networks XK and ZB with three station pairs by forcing two stations within every pair with the same amplification factors. The selection criteria for station pairs are: the distance between two stations should be less than  $0.7^{\circ}$  and for every station, whose amplification difference with the other two closest stations within its network should be less than 0.2 in order to avoid the abnormal station, like LP72. The entire amplification maps connected by only one station pair and more than one pair are consistent in the first order (Fig. A5 c and d), which means forcing neighboring stations from separate networks can connect subregional relative amplification maps together effectively only if with careful choice of station pairs. We fix the amplification of one station as 1 and generate the local amplification map for the entire study region by connecting one station pair.

Finally, after generating the amplification maps, we identify and remove outlier stations based on amplification factors following Eddy and Ekström (2014). Because of the uneven deployment of seismic arrays, instead of defining the outlier stations as the stations

with amplification  $> 2\sigma$  different from at least five neighboring stations, we use a more flexible strategy in this study. Here  $\sigma$  is the standard deviation of amplification factors. For a given station, if the number of neighboring stations (distance within  $2^\circ$ ) with amplification  $> 1.5\sigma$  different from this station over the number of all neighboring stations is larger than  $1/6$ , we will identify this station as an outlier station, and remove it. We apply this method to other periods and can get local amplification factors at all periods.

### 3 Results

We apply the data selection method to all eight periods and only the station pairs with more than 30 events will be used to do the amplitude ratio data selection. Fig. A6 shows the distribution of  $d_{ij}^k$  at every period which we will use to inverse local site amplification. Both the two datasets are generated with amplitude ratio data selection (only accept  $d_{ij}^k$  within one standard deviation in every station pair), and the only difference is with and without amplitude data selection. Compared with datum without amplitude data selection, the distribution of  $d_{ij}^k$  after amplitude data selection is more concentrated on the value of 0 with smaller standard deviations at all periods. The standard deviation of  $d_{ij}^k$  is smaller in intermediate periods (40 s) and then increases with increasing periods, which may be dominated by instrument calibration. We calculate the features of the datum used for inversion with and without amplitude data selection and report them in Table A2. The average station-pair uncertainty  $\bar{\sigma}$  ( $\bar{d}_{ij}$  uncertainty) also shows the same pattern as  $d_{ij}^k$  uncertainty. The higher uncertainty at shorter periods may reflect the larger scatter and the

interstation distance is larger than the focusing and defocusing wavefront at a wavelength. The higher uncertainty at longer periods is more likely to be because of the small number of high-quality amplitude measurements.

Fig. A7 shows all local amplification factor maps at eight periods in our study using the two datasets to the inversion respectively: datum processed with and without amplitude data selection, and both of them are processed with amplitude ratio data selection and outlier stations removal. Networks 1G+ and XK are connected by the permanent station IU.KEV naturally at 25-80 s, and because of obvious deviation amplitude measurement of IU.KEV at 100 s with other nearby stations, we connect the two separate networks using neighboring stations 1G.SA09 and XK.LP75 at 100 s. We fix the amplification factor of station KEV from network XK at 1 as a reference station and generate the relative amplification factors of all stations at all periods. The amplification maps with amplitude data selection are smoother with fewer spatial anomaly stations than those without amplitude data selection. For example, the amplification factors of stations from high-density NO network in east Norway is much scattered without amplitude data selection. Even with the final outlier stations removal step, the ability to remove outlier stations of amplification maps without amplitude data selection is limited because of the high uncertainties in amplification.

In order to access how our inversed amplification factors can explain the observations, we calculate variance reduction and root mean square (RMS) of residuals (Fig. A8). Here the straight variance reduction is calculated as  $V = 100 * 1 - \frac{\sum(\bar{d}_{ij}^{obs} - d_{ij}^{pre})^2}{\sum(\bar{d}_{ij}^{obs} - \mu)^2}$ , where  $\bar{d}_{ij}^{obs}$  is the observed average logarithmic amplitude ratio between station  $i$  and  $j$ ,  $d_{ij}^{pre}$  is the

logarithmic amplitude ratio predicted by the amplification factors  $A_i$  and  $A_j$ , and  $\mu$  is the average of  $\bar{d}_{ij}^{obs}$ . By adding a weighting matrix, we also calculate the weighted variance

reduction as  $V = 100 * 1 - \frac{\sum (\frac{1}{\sigma_{ij}^2} * (\bar{d}_{ij}^{obs} - d_{ij}^{pre}))^2}{\sum (\frac{1}{\sigma_{ij}^2} * (\bar{d}_{ij}^{obs} - \mu))^2}$ , where the weighting matrix  $\frac{1}{\sigma_{ij}^2}$  is same as

that in misfit function. Similarly, we calculate the straight and weighted RMS of residuals ( $\bar{d}_{ij}^{obs} - d_{ij}^{pre}$ ). We set 0.001 as a minimum threshold of  $\sigma_{ij}^2$ . When the uncertainty of  $d_{ij}^k$  is smaller, the difference between straight and weighting variance reduction (also for RMS of residuals) is smaller. For the dataset with amplitude data selection, the weighting variance reduction peaked at 32 s and is lower at shorter and longer periods. The difference between weighting and straight variance reduction is smaller at intermediate periods and larger at shorter and longer periods. The RMS of residuals is also the smallest at intermediate periods and larger at shorter and longer periods, which is consistent with the variation in the data, suggesting a higher standard deviation of amplitude ratio data leads to a lower variance reduction and higher RMS of residuals. For the datasets without amplitude data selection, the difference between straight and weight variance reduction is larger. The higher absolute variance reduction values are caused by the larger deviation with  $\mu$  (higher uncertainty in data). The RMS of residuals is much higher than that with amplitude data selection, which means our data selection method is working.

Because of the finite-frequency effect, we expect the surface wave amplification to be smooth and coherent spatially (Eddy and Ekström, 2020). Following Eddy and Ekström (2014), we divide the interstation distances into  $0.5^\circ$  bins and calculate the median absolute amplification difference for pairs of stations within the interstation distance range at every period. Fig. A9 shows the median amplification differences as a function of interstation

distance for the Rayleigh wave amplification factors processed with and without amplitude data selection. The amplification difference at a minimum distance can be seen as a reflection of the average quality of calibration of stations. The value of the median amplification difference at the minimum distance with amplitude data selection ( $\sim 3\text{--}4\%$ ) is smaller than the variation of amplification factors. Compared with amplification factors without amplitude data selection, the amplification differences at the minimum distance with our amplitude data selection are much smaller at all periods, suggesting the effect of removing outliers. We don't observe obvious patterns of amplification differences change between periods, which may be caused by the small-scale local elastic structure in this study region.

## 4 Discussion

The variation range of amplification factors ( $\sim -20\% - 20\%$ ), is larger than the median amplification differences at the minimum distance ( $\sim 4\%$ ) at all periods (Fig. A9), which are assumed to reflect the quality of instrument calibration. As a result, we believe the observed amplification variations can largely be attributed to elastic structure variations beneath stations. Our observed amplification factor maps show good spatial coherence with known tectonic and geologic structures (Fig. A10). High amplification at short periods in the Nordland province in north Norway coincides with slow phase velocity in this region, which is associated with low-density Trans-Scandinavian Igneous Belt magmatic intrusions (Ebbing 2007; Breivik et al. 2017). High amplification in the south and central Finland at 20 – 50 s correlates with thick Paleoproterozoic crust, which agrees with deep



Moho in Finnish Svecofennian terrane suggested by previous studies (Alinaghi et al. 2003; Kozlovskaya et al. 2008). Higher amplification in southern Norway at longer periods also agrees with the expected high amplification from the observed low S-wave velocity in the upper mantle in this region (Maupin 2011; Kolstrup et al. 2015).

In order to evaluate the degree to which our amplification measurement correlates with known elastic crust and upper mantle structure in the study region, we compare our amplification factor maps with the predicted amplification factors of two local tomographic models in Fennoscandia. A predicted relative Rayleigh wave receiver amplitude factor on the vertical component can be expressed as  $A_R = \frac{A}{A_0} = \sqrt{\frac{G_0}{G}} \frac{U_R}{U_{R0}}$ , where  $A$  is the Rayleigh wave receiver amplitude factor in the receiver location and can be calculated as  $A = \frac{U_R}{\sqrt{G}}$  (e.g., Tromp and Dahlen, 1992), where  $U_R$  is the radial eigenfunction and  $G$  is local group velocity of the model in the receiver location.  $A_0$ ,  $U_{R0}$ , and  $G_0$  are reference Rayleigh wave receiver amplitude factor, radial eigenfunction, and local group velocity in a reference receiver location. We used the location of the same reference station as in our measurement to calculate the predicted relative amplification factors. For a given pixel, four values are calculated using MINEOS software package (Masters et al. 2011) by extracting 1-D velocity models in this location.

Two 3-D shear-wave velocity models are used to generate predicted Rayleigh wave amplification factors. Mauerberger et al. (2022) developed a 3-D shear wave velocity model in Scandinavia down to 250 km inversed from phase dispersion curves from ambient noise and teleseismic Rayleigh waves, referred to as LithoScan2022. Zhu et al. (2014) determined an anisotropic seismic model of the crust and upper mantle beneath the

European continent and the North Atlantic Ocean, EU60, based on P-SV body waves, SH body waves, Rayleigh waves, and Love waves recorded on three components. Elastic wave speed variations, including radial and azimuthal anisotropy, and anelastic attenuation are inverted in EU60. We sample and extract the 1-D shear-wave velocity model from two 3-D models on the same grid size with  $0.1^\circ$  by  $0.1^\circ$  and generate the relative amplification factor maps using MINEOS.

Fig. A10 shows the comparison of our measured amplification factors with predictions from LithoScan2022 and EU60. In the first order, they show spatial coherency in main tectonic and geologic structure anomalies. At short periods (20-32 s), high amplification factors in Nordland province can be observed in all three maps. The continuous high amplification in Norland at longer periods in our measurement is consistent with EU60. High amplification in southern Finland at intermediate periods (32-60 s), which may be associated with a thick crust, is a striking feature in all three maps. At longer periods, the smaller difference in amplification between western and eastern Fennoscandia is consistent in three maps. For amplification factors in southern Norway, except for 20 s, the change from low to high amplification factors can also be observed in all maps. However, there are also some disagreements between the three maps. For example, the distinct low-velocity arc between Trondheim and the Norwegian-Swedish border at 32 s generates high values in amplification maps in both LithoScan2022 and EU60, but such high amplification factors are not observed in our measurement.

Associating Rayleigh wave amplification factors with elastic structure is more difficult than phase velocity because of the different characteristics of seismic velocity and density sensitivity kernels for amplification and phase velocity (Eddy and Ekström, 2014). The

dominantly positive phase velocity sensitivity kernels are more straightforward to interpret the variation of elastic wave speed than the amplification sensitivity kernels with a changing sign with depth. At a given period, a positive  $V_s$  perturbation at shallow depth will lead to an increase in amplification but a positive  $V_s$  perturbation at greater depth will lead to a decrease in amplification. For example, a striking feature in the phase velocity map at 50 s of LithoScan2022 is the sharp west-east velocity increase below the Oslo Graben (Fig. A8 in Mauerberger et al. 2022), but the transformation in predicted amplification maps is smoother at 50 s (Fig. A10). At the same time, it is worth noticing the difference between the two predicted amplification maps. The variation range of LithoScan2022 is less than EU60, and there are some small-scale disagreements between the two predicted amplifications from two models based on different data types. Considering the amplification is more sensitive to shallow structure, density, and  $V_p/V_s$  ratio (Lin et al. 2012b), the different predicted amplification maps may imply combining observations of surface-wave amplification with surface-wave phase velocity could help to better constrain elastic structure models.

## 5 Conclusion

The frequency-dependent local site amplification contains rich information on the elastic structure beneath stations. We apply the amplitude ratio method developed by Eddy and Ekström (2014) in Fennoscandia to estimate the local amplification of Rayleigh waves at periods 20 -100 s with regional broadband seismometers mainly from three non-time-overlapping networks. We develop a method to control data quality by selecting amplitude

data and amplitude ratio data, constant bad stations and time-dependent erroneous stations are identified and removed from the datum. By forcing two or more stations from separate networks with the same amplification factors, we can combine non-time-overlapping networks together to generate the amplification maps in the entire study region. Our amplitude data selection method decreases standard deviations of amplitude ratio significantly at all eight periods and also decreases the median amplification differences between two stations at the minimum distance, which is an indication of the quality of instrument calibration.

The measured amplification factor maps show good spatial agreement with known tectonic and geologic structures, like high amplification at short periods in the Nordland province correlated with slow phase velocity in this region, high amplification in the south and central Finland at 20 – 50 s coinciding with a thick crust. Our observed amplification maps agree with the predictions of two local 3-D tomographic models in the first order. The differences between them may imply the potential of amplification to refine elastic structure in the crust and upper mantle.

## 6 References

- Anthony, R.E., Ringler, A.T. and Wilson, D.C., 2022. Seismic background noise levels across the Continental United States from USArray transportable array: The influence of geology and geography. *Bulletin of the Seismological Society of America*, 112(2), pp.646-668.
- Alinaghi, A., Bock, G., Kind, R., Hanka, W., Wylegalla, K. and Tor and Svekalapko Working Groups, 2003. Receiver function analysis of the crust and upper mantle from the North German Basin to the Archaean Baltic Shield. *Geophysical Journal International*, 155(2), pp.641-652.
- Bao, X., Dalton, C.A., Jin, G., Gaherty, J.B. and Shen, Y., 2016. Imaging Rayleigh wave attenuation with USArray. *Geophysical Journal International*, 206(1), pp.241-259.
- Bowden, D.C., Tsai, V.C. and Lin, F.C., 2017. Amplification and attenuation across USArray using ambient noise wavefront tracking. *Journal of Geophysical Research: Solid Earth*, 122(12), pp.10-086.
- Breivik, A.J., Faleide, J.I., Mjelde, R., Fluch, E.R. and Murai, Y., 2017. A new tectono-magmatic model for the Lofoten/Vesterålen Margin at the outer limit of the Iceland Plume influence. *Tectonophysics*, 718, pp.25-44.
- Dalton, C.A. and Göran, E., 2006a. Constraints on global maps of phase velocity from surface-wave amplitudes. *Geophysical Journal International*, 167(2), pp.820-826.

- Dalton, C.A. and Ekström, G., 2006b. Global models of surface wave attenuation. *Journal of Geophysical Research: Solid Earth*, 111(B5).
- Ebbing, J., 2007. Isostatic density modelling explains the missing root of the Scandes. *Norwegian Journal of Geology/Norsk Geologisk Forening*, 87.
- Eddy, C.L. and Ekström, G., 2014. Local amplification of Rayleigh waves in the continental United States observed on the USArray. *Earth and Planetary Science Letters*, 402, pp.50-57.
- Eddy, C.L. and Ekström, G., 2020. Comparisons between measurements and predictions of Rayleigh wave amplification across the contiguous United States. *Physics of the Earth and Planetary Interiors*, 299, p.106407.
- Ekström, G., Dalton, C.A. and Nettles, M., 2006. Observations of time-dependent errors in long-period instrument gain at global seismic stations. *Seismological Research Letters*, 77(1), pp.12-22.
- Ester, M., Kriegel, H.P., Sander, J. and Xu, X., 1996, August. A density-based algorithm for discovering clusters in large spatial databases with noise. In *kdd* (Vol. 96, No. 34, pp. 226-231).
- Ferreira, A.M. and Woodhouse, J.H., 2007. Source, path and receiver effects on seismic surface waves. *Geophysical Journal International*, 168(1), pp.109-132.
- Gilbert, F., 1971. Excitation of the normal modes of the Earth by earthquake sources. *Geophysical Journal International*, 22(2), pp.223-226.

- Hariharan, A., Dalton, C.A., Babikoff, J. and Ekström, G., 2022. Controls on surface wave overtone interference. *Geophysical Journal International*, 228(3), pp.1665-1683.
- James, E.K., Dalton, C.A. and Gaherty, J.B., 2014. Rayleigh wave phase velocities in the Atlantic upper mantle. *Geochemistry, Geophysics, Geosystems*, 15(11), pp.4305-4324.
- Jin, G. and Gaherty, J.B., 2015. Surface wave phase-velocity tomography based on multichannel cross-correlation. *Geophysical Journal International*, 201(3), pp.1383-1398.
- Kolstrup, M.L., Hung, S.H. and Maupin, V., 2015. Multiscale, finite-frequency P and S tomography of the upper mantle in the southwestern Fennoscandian Shield. *Geophysical Journal International*, 202(1), pp.190-218.
- Kozlovskaya, E., Kosarev, G., Aleshin, I., Riznichenko, O. and Sanina, I., 2008. Structure and composition of the crust and upper mantle of the Archean–Proterozoic boundary in the Fennoscandian shield obtained by joint inversion of receiver function and surface wave phase velocity of recording of the SVEKALAPKO array. *Geophysical Journal International*, 175(1), pp.135-152.
- Lin, F.C., Schmandt, B. and Tsai, V.C., 2012a. Joint inversion of Rayleigh wave phase velocity and ellipticity using USArray: Constraining velocity and density structure in the upper crust. *Geophysical Research Letters*, 39(12).

- Lin, F.C., Tsai, V.C. and Ritzwoller, M.H., 2012b. The local amplification of surface waves: A new observable to constrain elastic velocities, density, and anelastic attenuation. *Journal of Geophysical Research: Solid Earth*, 117(B6).
- Ma, Z., Masters, G. and Mancinelli, N., 2016. Two-dimensional global Rayleigh wave attenuation model by accounting for finite-frequency focusing and defocusing effect. *Geophysical Journal International*, 204(1), pp.631-649.
- Masters, G., Woodhouse, J.H. and Freeman, G., 2011. Mineos v1. 0.2 [software], computational infrastructure for geodynamics. Available from: [geodynamics.org](http://geodynamics.org), doi: NoDOI, url: <https://geodynamics.org/cig/software/mineos>.
- Mauerberger, A., Sadeghisorkhani, H., Maupin, V., Gudmundsson, Ó. and Tilmann, F., 2022. A shear-wave velocity model for the Scandinavian lithosphere from Rayleigh waves and ambient noise-Implications for the origin of the topography of the Scandes mountain range. *Tectonophysics*, 838, p.229507.
- Maupin, V., 2011. Upper-mantle structure in southern Norway from beamforming of Rayleigh wave data presenting multipathing. *Geophysical Journal International*, 185(2), pp.985-1002.
- Ringler, A.T., Anthony, R.E., Dalton, C.A. and Wilson, D.C., 2021. Rayleigh-Wave Amplitude Uncertainty across the Global Seismographic Network and Potential Implications for Global Tomography. *Bulletin of the Seismological Society of America*, 111(3), pp.1273-1292.



Schardong, L., Ferreira, A.M., Berbellini, A. and Sturgeon, W., 2019. The anatomy of uppermost mantle shear-wave speed anomalies in the western US from surface-wave amplification. *Earth and Planetary Science Letters*, 528, p.115822.

Tromp, J. and Dahlen, F.A., 1992. Variational principles for surface wave propagation on a laterally heterogeneous Earth—II. Frequency-domain JWKB theory. *Geophysical Journal International*, 109(3), pp.599-619.

Zhu, H., Bozdağ, E. and Tromp, J., 2015. Seismic structure of the European upper mantle based on adjoint tomography. *Geophysical Journal International*, 201(1), pp.18-52.

## Tables

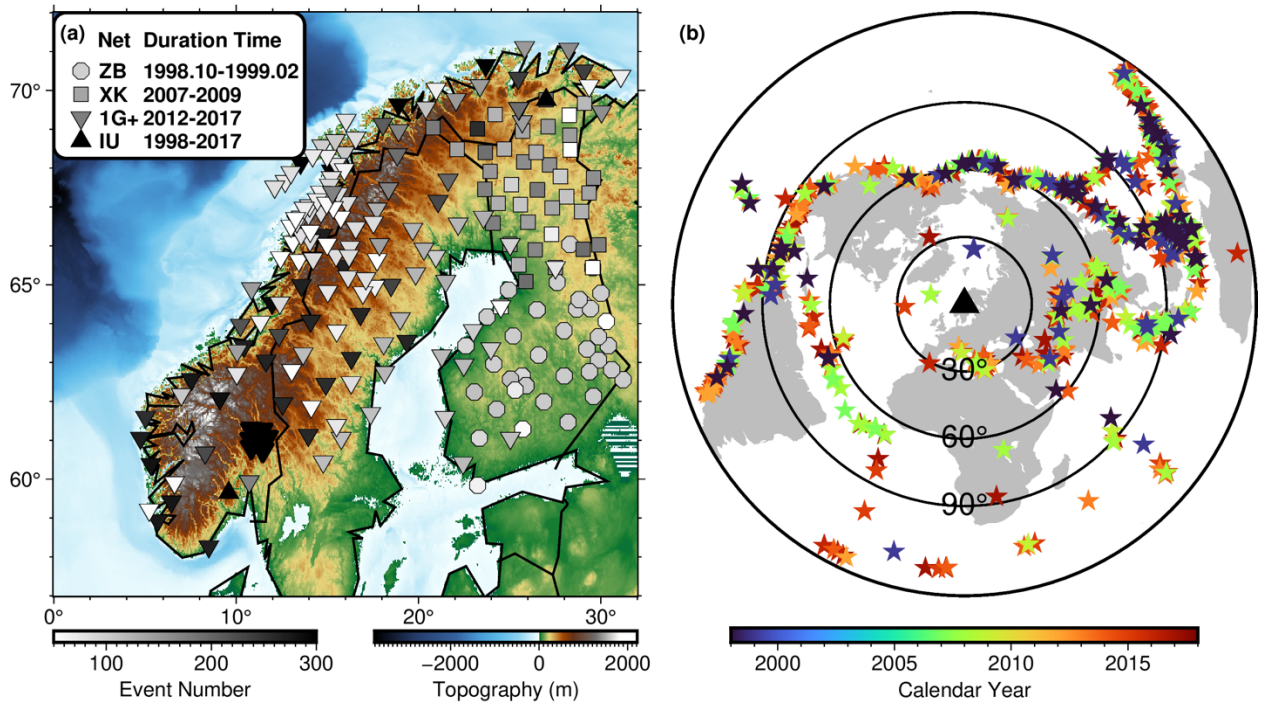
| Periods | Before Step 1 |     |      | After Step 1 |     |      | After Step 2 |     |      | After Step 3 |     |      |
|---------|---------------|-----|------|--------------|-----|------|--------------|-----|------|--------------|-----|------|
|         | #A            | #S  | #A/S | #A           | #S  | #A/S | #A           | #S  | #A/S | #A           | #S  | #A/S |
| 20 s    | 40625         | 267 | 152  | 36277        | 259 | 140  | 31640        | 259 | 124  | 31152        | 251 | 124  |
| 25 s    | 40628         | 267 | 152  | 37169        | 256 | 145  | 32546        | 256 | 129  | 32192        | 248 | 130  |
| 32 s    | 40630         | 267 | 152  | 37268        | 257 | 145  | 32783        | 257 | 129  | 32332        | 251 | 129  |
| 40 s    | 40631         | 267 | 152  | 37347        | 258 | 145  | 32890        | 258 | 128  | 32005        | 248 | 129  |
| 50 s    | 40636         | 267 | 152  | 36731        | 256 | 143  | 32949        | 256 | 129  | 31965        | 248 | 129  |
| 60 s    | 40649         | 267 | 152  | 36385        | 256 | 142  | 32619        | 256 | 128  | 31564        | 247 | 128  |
| 80 s    | 40652         | 267 | 152  | 34358        | 257 | 134  | 32089        | 257 | 125  | 31543        | 251 | 126  |
| 100 s   | 40652         | 267 | 152  | 32229        | 257 | 125  | 31434        | 257 | 122  | 30764        | 249 | 124  |

**Table A1.** Size of data metrics before and after 3 steps in amplitude data selection. #A: number of amplitude data. #S: number of stations. #A/S: average number of amplitude data per station.

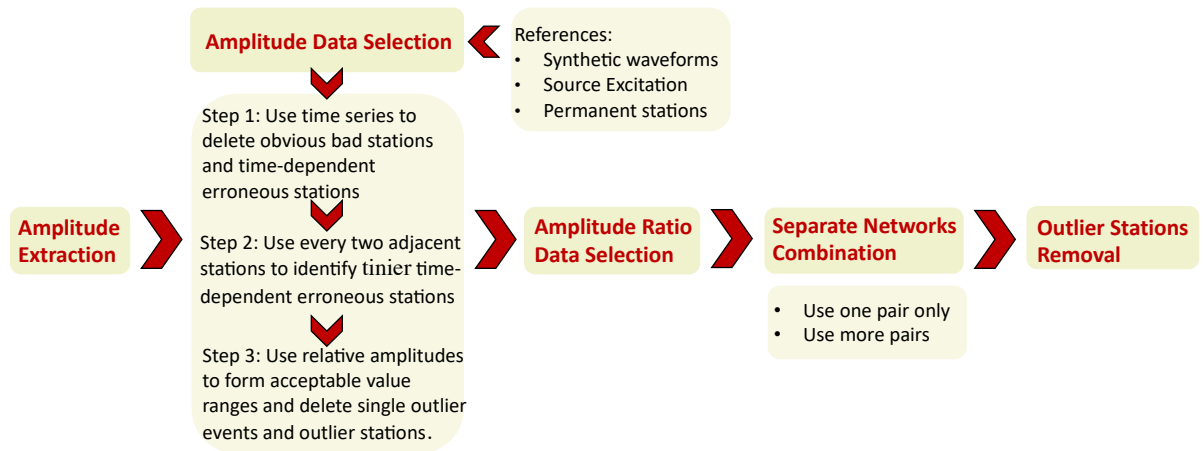
| Period (s) | $N_s$ (B) | $N_s$ (A) | $N_p$ (B) | $N_p$ (A) | $N_e$ (B) | $N_e$ (A) | $\bar{\sigma}$ (B) | $\bar{\sigma}$ (A) |
|------------|-----------|-----------|-----------|-----------|-----------|-----------|--------------------|--------------------|
| 20 s       | 260       | 234       | 3485      | 2850      | 110.78    | 72.96     | 0.2973             | 0.1664             |
| 25 s       | 260       | 237       | 3485      | 2935      | 113.88    | 76.75     | 0.2731             | 0.1481             |
| 32 s       | 260       | 237       | 3485      | 2882      | 119.14    | 80.15     | 0.2663             | 0.1369             |
| 40 s       | 260       | 234       | 3485      | 2872      | 119.04    | 80.02     | 0.2757             | 0.1300             |
| 50 s       | 260       | 234       | 3485      | 2892      | 119.20    | 83.90     | 0.3314             | 0.1534             |
| 60 s       | 260       | 234       | 3485      | 2742      | 119.42    | 86.52     | 0.3842             | 0.1664             |
| 80 s       | 260       | 235       | 3485      | 2684      | 116.91    | 86.58     | 0.4419             | 0.1924             |
| 100 s      | 260       | 230       | 3485      | 2573      | 114.70    | 84.61     | 0.4916             | 0.1937             |

**Table A2.** Data metrics before (B) and after (A) amplitude data selection which were used for inversion.  $N_s$ : number of stations.  $N_p$ : number of station pairs.  $N_e$ : average number of events per station pair.  $\bar{\sigma}$ : average station-pair uncertainty.

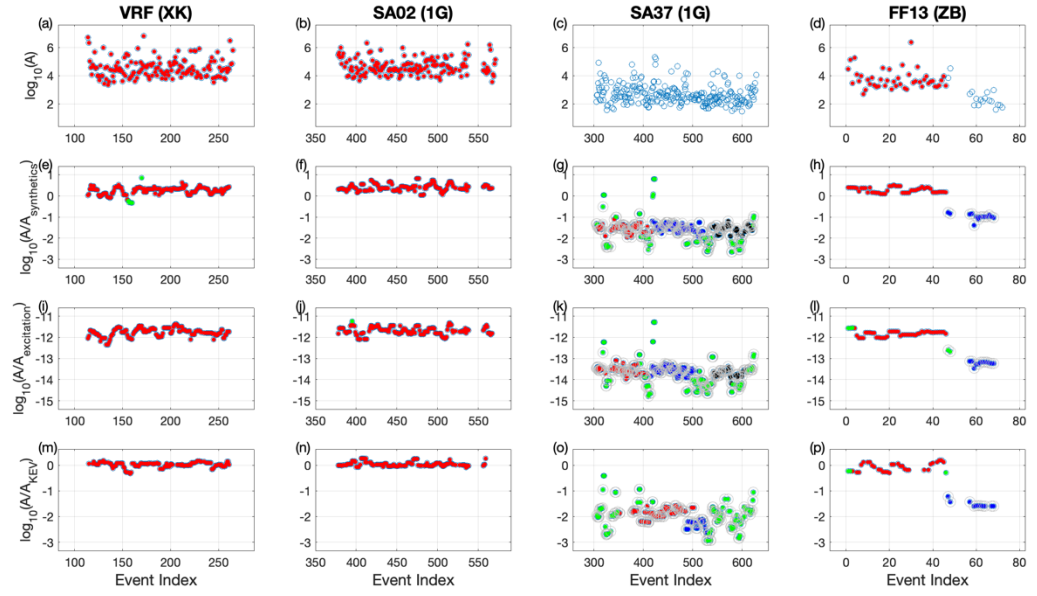
## Figures



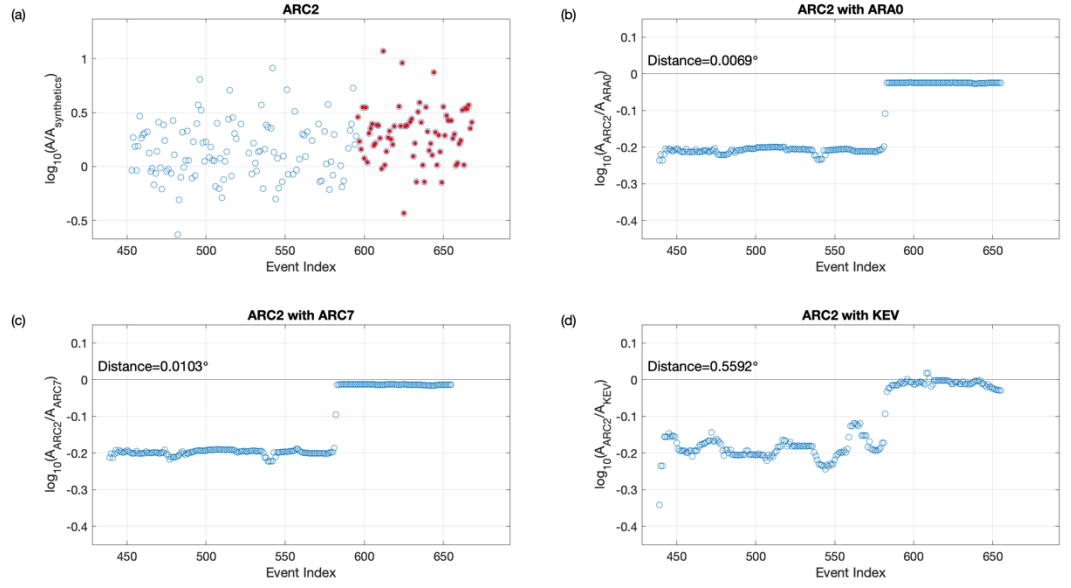
**Figure A1.** (a) Topography map of Fennoscandia with 283 seismic stations from the networks used in this study. The stations are colored according to the number of earthquakes that each one contributes to the amplitude data set. (b) Locations of 669 teleseismic earthquakes used in this study, colored by event origin time.



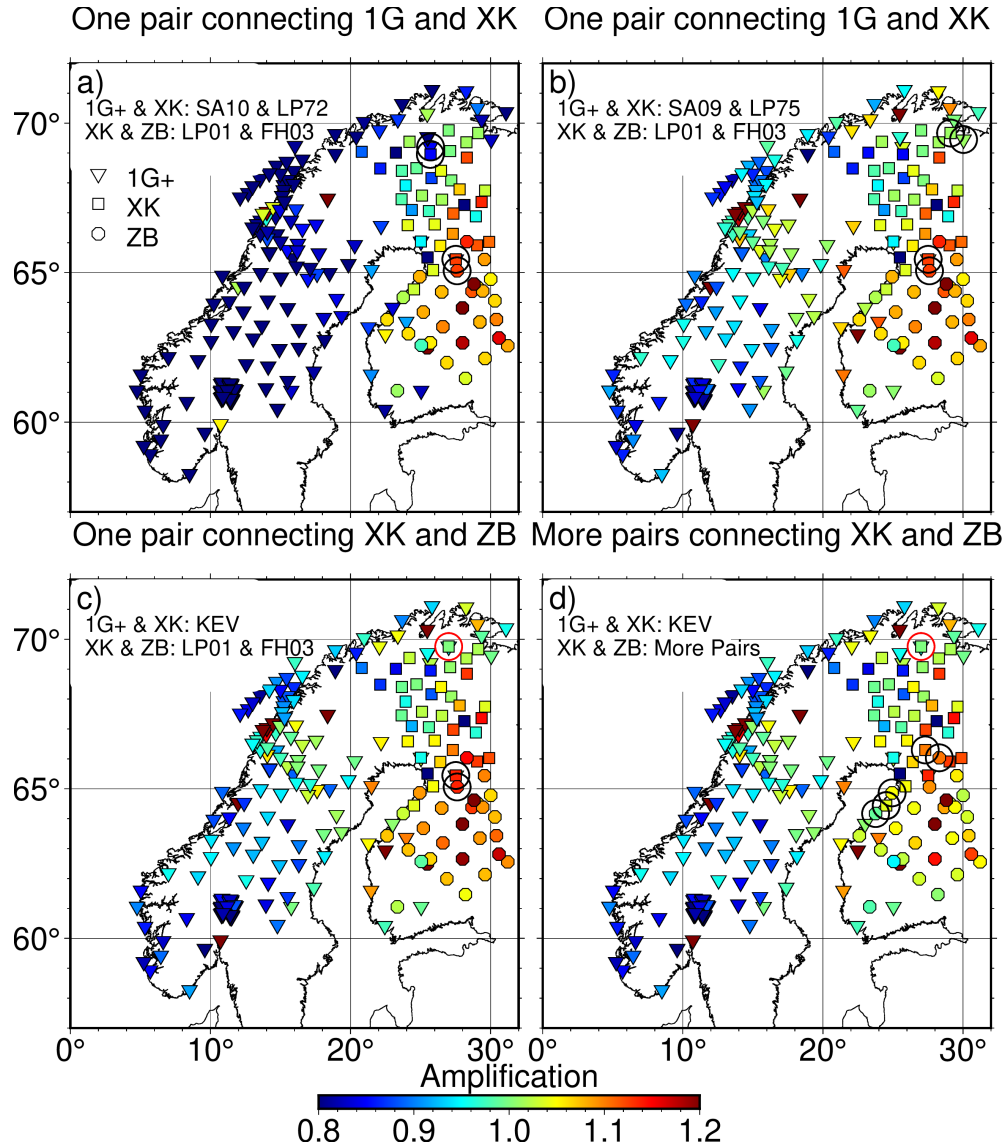
**Figure A2.** Workflow with the five primary steps used for processing data.



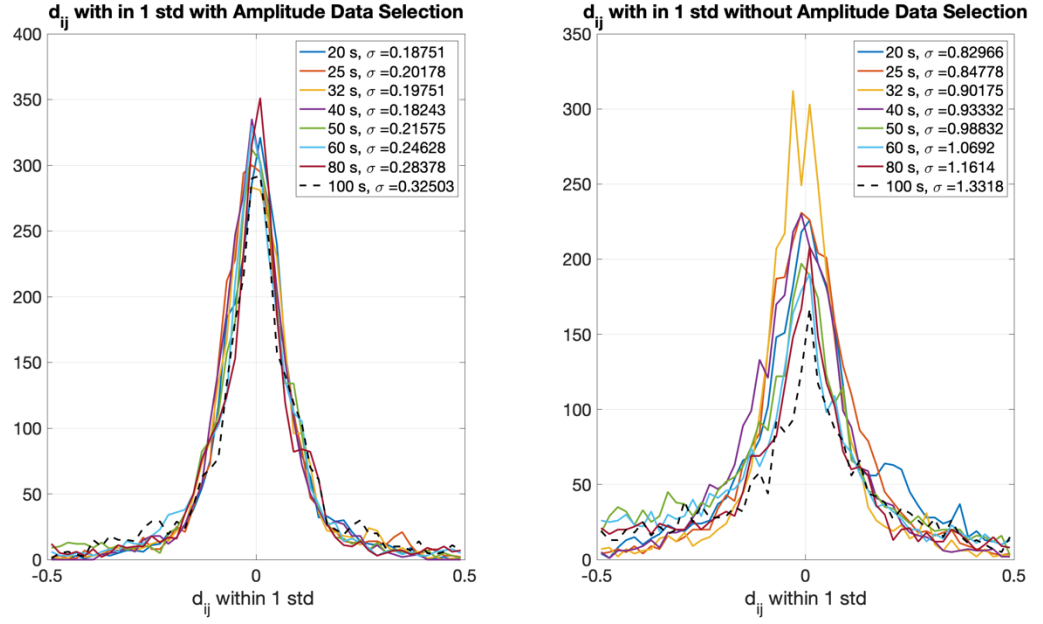
**Figure A3.** Example of step 1 in amplitude data selection. Amplitude data of four example stations: VRF-XK, SA02-1G, SA37-1G, and FF13-ZB. In (a)-(d) blue circles are all measured amplitude data by event and red dots are quality-selected amplitudes after step 1 using synthetic waveforms as references. (e)-(h), (i)-(l), and (m)-(p) are relative amplitudes with synthetic waveforms, source excitation, and station KEV-IU respectively for four stations. Dots with various colors in (e)-(p) mean various clusters using a density-based clustering algorithm. Circled dots are the outlier event groups and will be removed in step 1.



**Figure A4.** Example of step 2 in amplitude data selection. (a) Relative amplitude data with synthetic waveforms of station ARC2. Blue circles are measured amplitude data by the event after step 1 and red dots are quality-selected amplitudes after step 2. (b)-(d) Amplitude comparison of ARC2 and three adjacent stations, which shows a consistent pattern along the event index: there is a sudden increase in amplitude data recorded by station ARC2 from the 582nd event systematically. The distances between station ARC2 and the other three nearby stations are labeled in (b)-(d).

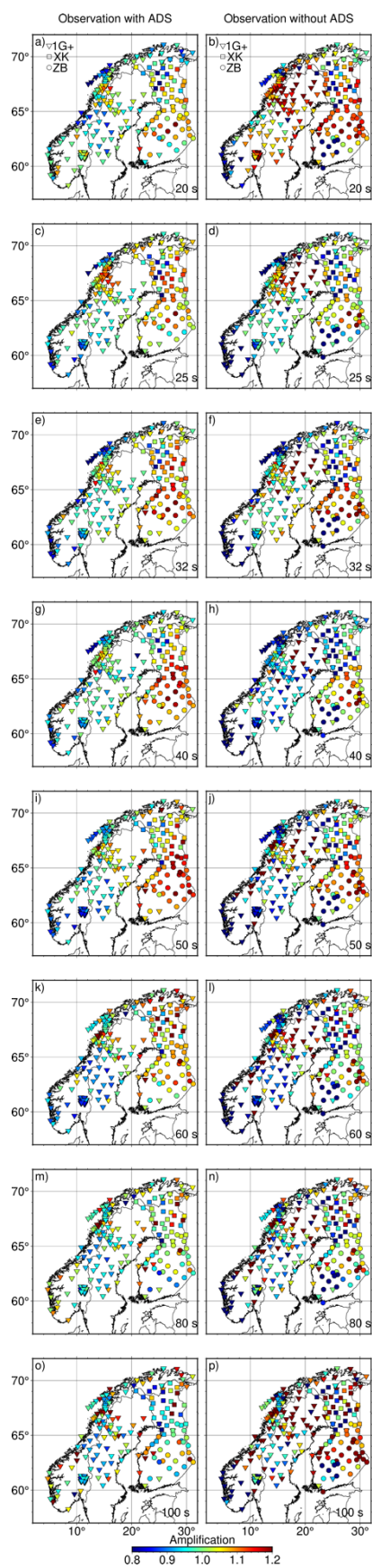


**Figure A5.** Observed local Rayleigh wave Amplification factor maps at the period of 50 s with different connecting station pairs when connecting separate networks. (a) Amplification factor map by using station SA10 and LP72 to connect network 1G+ and XK, and using LP01 and FH03 to connect network XK and ZB. Connected station pairs are circled. (b) Same as (a) except for connecting network 1G+ and XK using station SA09 and LP75. The different patterns between (a) and (b) mean the selection of connecting station pairs should be cautious. (c) Local Rayleigh wave amplification factor map using permanent station KEV (red circle) to connect network 1G+ and XK and station LP01 and FH03 (black circles) to connect XK and ZB. (d) same with (c) but using three station pairs (black circles) simultaneously to connect networks XK and ZB.

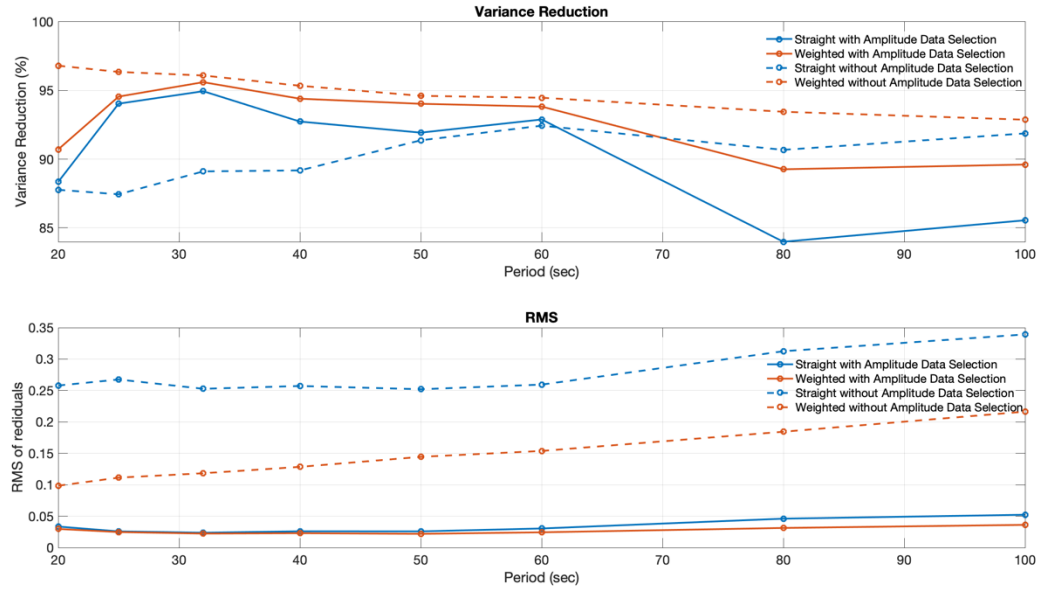


**Figure A6.** Distributions of  $d_{ij}^k$  values generated from the dataset proceeded with amplitude data selection (left) and without amplitude data selection (right) at periods 20-100 s, which will be used in the following local site amplification factors inversion.

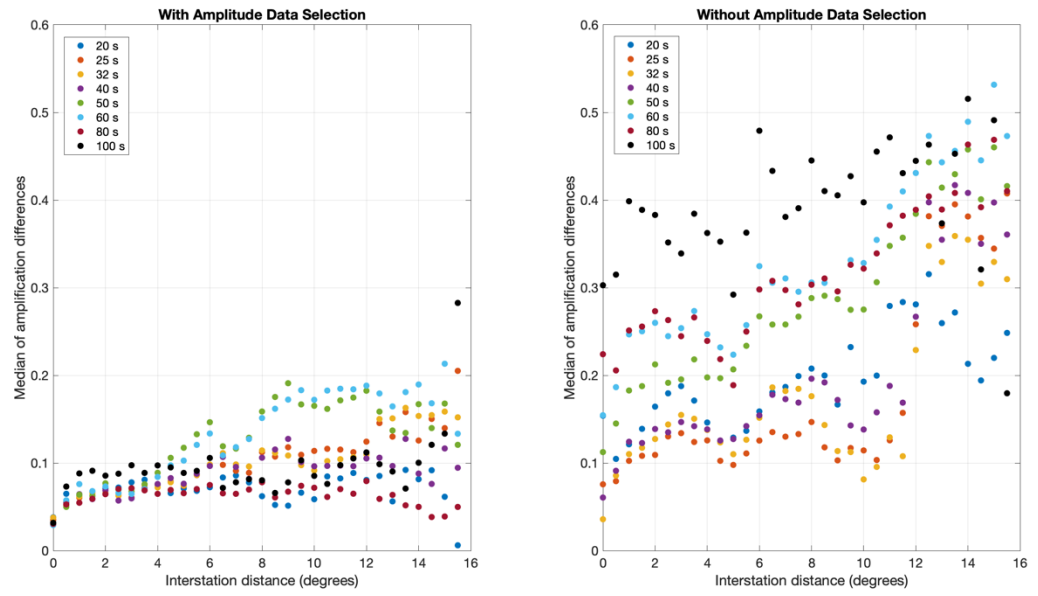




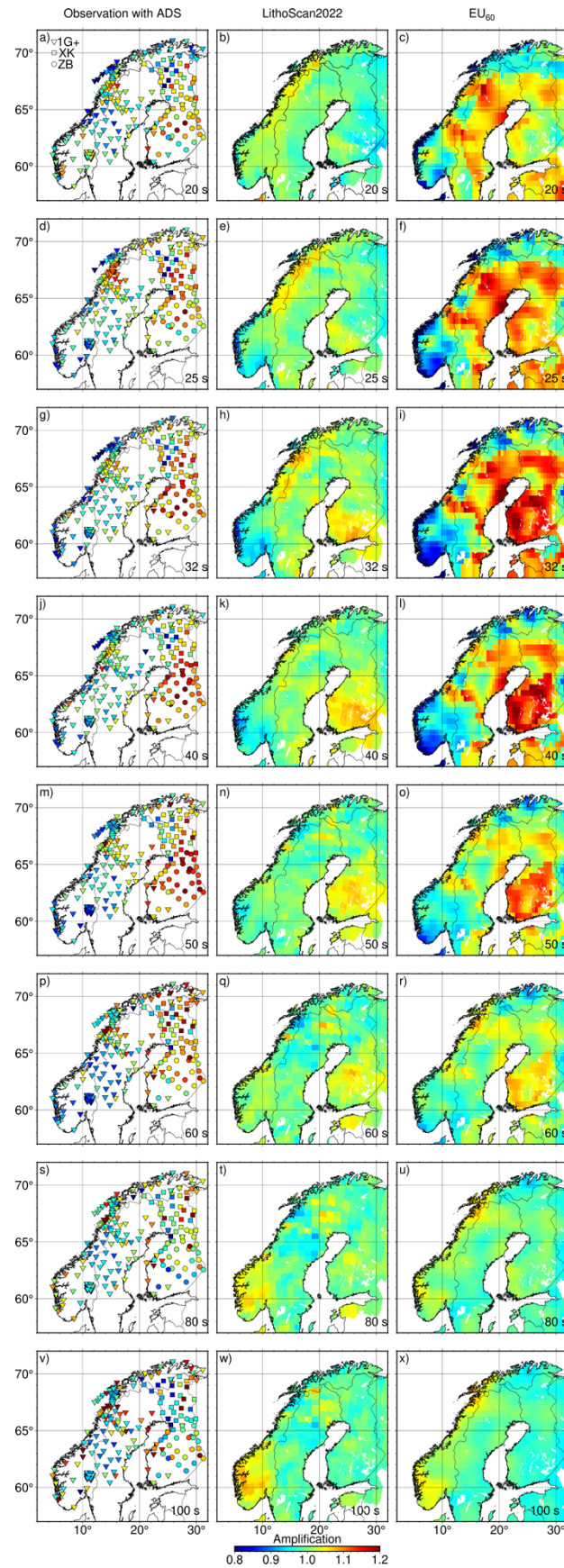
**Figure A7.** Maps of local site amplification at periods 20-100 s using a dataset with amplitude data selection in the left column and without amplitude data selection in the right column.



**Figure A8.** Variance reduction (top) and RMS of residuals of Rayleigh wave amplification factors at periods 20 – 100 s. Straight (blue lines) and weighted (red lines) variance reduction and RMS of residuals are plotted. The variance reduction and RMS of residuals of amplification factors proceeded with and without amplitude data selection are plotted as continuous lines and dashed lines respectively.



**Figure A9.** Median absolute differences between amplification factors at stations as a function of interstation distance from the dataset proceeded with amplitude data selection (left) and without amplitude data selection (right) at periods 20 – 100 s.



**Figure A10.** Observed local Rayleigh wave amplification factors in the left column, and predicted amplification factors from local shear-wave velocity models LithoScan2022 (Mauerberger et al., 2022) and EU60 (Zhu et al., 2015) in the center and right column respectively.