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University of California
Santa Barbara

Distributed Extension across the Ethiopian Rift and Plateau Illuminated by Joint Inversion of Surface Waves and Scattered Body Waves

A dissertation submitted in partial satisfaction
of the requirements for the degree

Master of Science
in
Earth Sciences

by

Jon Mitchell Petruska

Committee in charge:

Professor Zachary Eilon, Chair
Professor Toshiro Tanimoto
Professor Robin Matoza

March 2022

The Dissertation of Jon Mitchell Petruska is approved.

Professor Toshiro Tanimoto

Professor Robin Matoza

Professor Zachary Eilon, Committee Chair

December 2022

Distributed Extension across the Ethiopian Rift and Plateau Illuminated by Joint
Inversion of Surface Waves and Scattered Body Waves

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by

Jon Mitchell Petruska

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Curriculum Vitæ

Jon Mitchell Petruska

Education

- | | |
|------|---|
| 2022 | M.S. in Geophysics (Expected), University of California, Santa Barbara. |
| 2018 | M.S. in Geology, Southern Illinois University. |
| 2015 | B.A. in Geology, University of Colorado, Boulder. |

Publications

Distributed Extension across the Ethiopian Rift and Plateau Illuminated by Joint Inversion of Surface Waves and Scattered Body Waves, *Geochemistry, Geophysics, Geosystems*, <https://doi.org/10.1029/2021GC010179>.

Abstract

Distributed Extension across the Ethiopian Rift and Plateau Illuminated by Joint
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by

Jon Mitchell Petruska

The East African Rift System provides a rare location in which to observe a wide scope of rifting states. Well-defined active narrow rifting in the Main Ethiopian Rift (MER) transitions to incipient extension and eventually pre-rifted lithosphere through the northwestern flank of the Ethiopian Plateau (EP). Although the MER is well studied, the off-axis region has received less attention. We develop Rayleigh wave phase velocity maps, Ps receiver functions, and $H-\kappa$ stack surfaces, and jointly invert these data using a trans-dimensional, hierarchical Bayesian inversion algorithm to create shear velocity profiles across the MER and EP. Shear velocity profiles characterize the subsurface structure as it changes with depth and identify the thickness of the crust. Reduced velocities may indicate elevated mantle temperatures and/or the presence of partial melt, and thinned crust may indicate lithospheric degradation and crustal thinning processes, respectively, all of which are associated with rifting. All shear velocities observed are slower than the PREM global average, which we interpret as a reflection of the elevated temperatures that persist from plume impingement. In the EP, we find a shallow mantle slow shear velocity lineament parallel to the MER axis, amidst otherwise faster shear velocities. The crust is shallow in the MER, but also in the northwestern EP flank, with thicker crust found throughout the plateau caused by crustal underplating and flood basalt emplacement. Shear velocities more reduced than the already low regional average, in concert with surficial volcanic features, geodetic observations, and slow P- and S-wave

anomalies, support off-axis extension in the Ethiopian plateau, requiring reevaluation of the localization of continental breakup in the narrow MER.

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Chapter 1

Introduction

1.1 Permissions and Attributions

1. The content of chapter 2 and appendix A are the result of my research with my advisor Zach Eilon, and has previously appeared in “Distributed Extension across the Ethiopian Rift and Plateau Illuminated by Joint Inversion of Surface Waves and Scattered Body Waves, *Geochemistry, Geophysics, Geosystems*, <https://doi.org/10.1029/2021GC010179>”.
2. The contents of appendix B consist of preemptory work that are relevant to the research project as a whole, but were not included in journal submission.

Chapter 2

Introduction and Background

2.1 Introduction and Motivation

The Main Ethiopian Rift (MER), the northern segment of the East African Rift System (EARS), represents an intermediate state between unbroken continental crust and incipient sea-floor spreading. Numerous studies have detailed the dynamics and evolution of this unique natural laboratory for extensional processes (1; 2; 3; 4; 5; 6; 7; 8; 9; 10; 11; 12; 13), and they agree on several key points. There exists widely distributed melt beneath the MER and nearby portions of the Ethiopian Plateau (EP). Fractional melt estimations vary from 0.5% to 2.7% (6; 11; 13; 14), and the presence of both melt and elevated temperatures are required to achieve low observed velocities and anisotropy (e.g., *Hammond et al., 2014*) within the rift. Crust in the MER is thinner than that of the Somalian Plateau (15; 16; 3; 17; 6; 18; 19; 20; 21) and the EP. Many studies also agree that a mantle-sourced impetus in the form of a plume was necessary to initiate rifting the strong, thick Ethiopian lithosphere (e.g., *Hayward & Ebinger, 1996; Ebinger & Sleep, 1998; Corti, 2009*), which consequently caused widespread thermal perturbations, associated magmatism, crustal underplating, and uplift across Ethiopia.

The scope of previous geophysical studies has been primarily constrained to the MER axis, Afar, and the near-axis portions of the EP. However, geodetic data suggests that $\sim 20\%$ - 30% of present-day extension does not occur within the magmatic segments of the MER (22) and may be distributed across the EP (23). While seismicity supports continuing extension along the border faults (24; 25; 26), westward distribution of seismicity (27; 28; 29), volcanism (30), small-scale GPS movements (31; 32), and low shear velocities (13; 14) in the EP support a degree of off-rift extension, which begs the questions: where does this off-axis extension occur, and how is extension accommodated at depth?

In order to answer these questions, we interrogate the shear velocity structure of the lower crust and upper mantle in the EP and northwestern Ethiopian Plateau flank (NWEF) using a recently deployed broadband seismic array in conjunction with legacy data (fig. 2.1a). We develop Rayleigh wave phase velocity maps, P_s receiver functions, and H- κ stack surfaces from teleseisms, and jointly invert these complementary data types using a trans-dimensional hierarchical Bayesian inversion algorithm (33). We solve for 1-D shear velocity profiles beneath the EP, NWEF, and within the MER, and use these to support the existence of off-axis extension. We interpret shear velocity structure in the context of previous studies (e.g., *Dugda et al., 2007*; *Gallacher et al., 2016*; *Weeraratne et al., 2003*) and thermal structure associated with rifting.

2.2 Geological and Tectonic Background

Ethiopia has a complex tectonic and volcanic history that has shaped the evolution of the MER and EP (fig. 2.1b). Formed within the Neoproterozoic orogenic Mozambique Belt (34; 35; 36; 37), Ethiopia preserves the Proterozoic-aged Himalayan style collision zone (38; 39) between East and West Gondwana, which scarred the basement with numer-

ous sutures (40; 41; 42), and Precambrian aged fractures (43). Some of these structures were reactivated in the Paleozoic-Mesozoic, creating ultimately aborted rifts (44; 45; 35), which left lithospheric weaknesses that have the potential to be exploited by magmatic or extensional forces.

Within the MER, NW-SE trending strike-slip faults have reactivated to allow extension (46; 35; 47; 43), demonstrating that existing fault structures control some degree of extension within the MER. NW-SE faults in the Ogaden basin and Blue Nile basin (fig. 2.1b), which transect the EP and NWEF, reactivated from the early to late Jurassic and interspersed lithospheric structures that may accommodate localized extension. Interrelation between Precambrian structures, active normal faulting, and magmatism outside the MER (*cf. Hautot et al., 2006*) is evidenced by Paleozoic-Mesozoic and Late Miocene-Quaternary dike swarms to the southwest and west of Lake Tana (45; 49; 50). Today, earthquake distribution is focused within the MER and Afar, often near magmatic segments, (51; 27; 52; 53), and along border faults separating the MER from adjacent plateaus (24; 25; 54). Sparse earthquakes are dispersed throughout the EP (29; 55), with increasing concentration nearing the MER.

Thermal modification from impingement of the Afar plume and associated Tertiary volcanics have altered the lithospheric structure of the MER from that of the Mozambique Belt, which may serve as a comparison point to the pre-rift lithospheric structure of Ethiopia (6). Plume impingement may have also chemically altered the lithosphere (56; 57; 58; 59; 60). Plume-related magmatism, exploiting pre-existing weaknesses and fabrics in the lithosphere, produced a ~1000 km diameter (61; 62; 63), 2 km thick region of flood basalts that cap much of the EP (fig. 2.1b). Flood basalt emplacement began ~45 Ma (64), but was most prevalent between 31 - 29 Ma. Within the EP, shield volcanoes developed at 22 Ma and 10 - 7 Ma (30) (fig. 2.1b). Between 10 - 12 Ma, volcanoes appeared both within the MER and near Lake Tana (65; 66; 30). Present day fumaroles

and thermal springs are active in both regions (fig. 2.1b). Quaternary volcanism is present in the form of Strombolian volcanoes and spatter cones south of Lake Tana (30), and along the Yerrer-Tullu Wellel Volcanotectonic lineament (YTVL), where clusters of volcanoes and fractures running E-W from the MER have erupted coevally with northern MER volcanoes (67; 68).

The MER is roughly outlined by Miocene aged border faults, and is trisected into Northern (NMER), Central (CMER), and Southern (SMER) sections (fig. 2.1b). Rift initiation was asynchronous, with SMER volcanism beginning in 20 - 18 Ma (69; 70; 71) and rifting following soon after at ~ 18 Ma (69). The NMER then initiated ~ 10 - 11 Ma (68; 72), with accompanied volcanism. The CMER initiated last and exhibits a less evolved stage of rifting than the SMER and NMER, but its exact age is contested: between 5 and 9.8 Ma based on K/Ar dating (73; 74). Surface expressions of contemporary extension are largely localized to magmatic centers within the MER (22; 72), having shifted from strike-slip faults perpendicular to the rift axis that accommodated Quaternary extension (46). Geochemical data show that the mantle beneath the MER and the nearby EP is hotter than the global average by 100° - 170°C (7; 10), and slow seismic velocities resolved here (1; 75; 4; 6; 7; 8; 9; 10; 11; 12; 13) reflect this elevated temperature, and the consequent presence of melt.

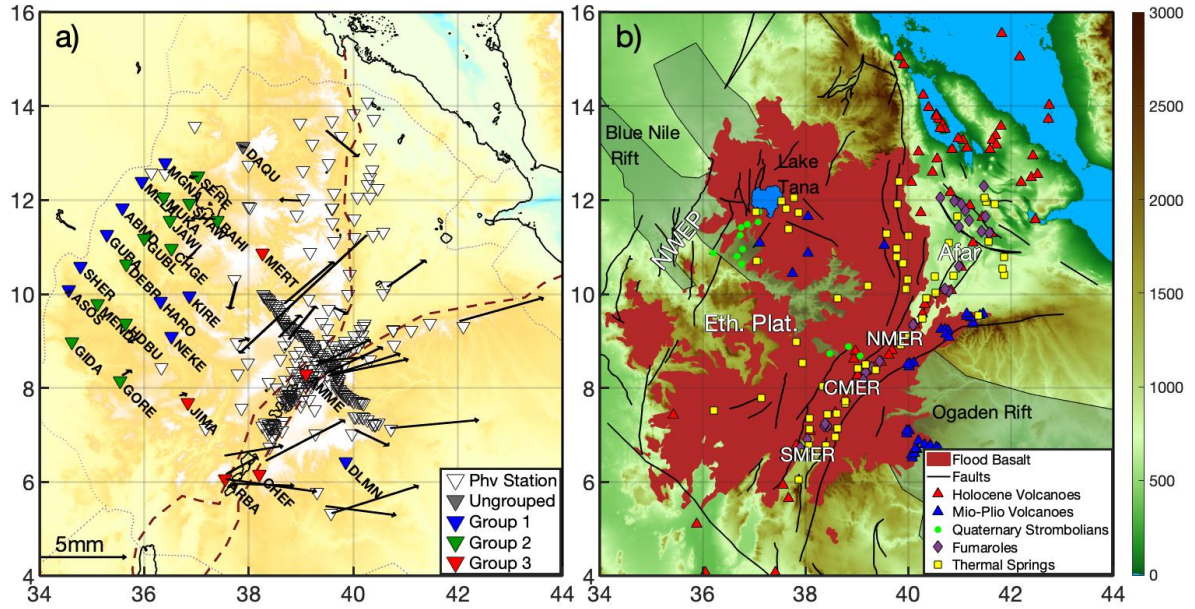


Figure 2.1: a) A map of all seismic stations used to develop phase velocity maps (triangles), atop an exaggerated DEM for symbol clarity. The x-axis is latitude, and the y-axis is longitude. Stations used to develop receiver functions, and that are inverted to develop shear velocity profiles are plotted in blue, green, and red, depending on their shear velocity group (see section 3.6). Station DAQU, which was inverted but not grouped, is plotted in dark grey. Stations plotted in white were used only for phase velocity measurements. Black arrows are GPS data from (23). b) A map of relevant geology in the study area, colored by elevation (in m) from (76), showing flood basalt extent from (77), geologically and currently active major faults digitized from (78), and a variety of volcanic and hydrothermal features (27). The bounds of the Ogaden and Blue Nile Rifts are shown in light green.

Chapter 3

Data and Methods

3.1 Phase Velocity Maps

In order to interrogate the velocity structure of the study region, we used the Automated Surface Wave phase velocity Measuring Systems (ASWMS) (79) to develop phase velocity maps at 25, 32, 40, 50, 60, 80, and 100 s using Rayleigh waves from teleseismic earthquakes. This period range provides sensitivity to lower crustal and mantle wavespeeds (fig. A.17). We used 5741 high quality earthquakes occurring between 2000-03-10 and 2017-10-03, of $Mw \geq 5.5$, a maximum event depth of 250 km, and spanning a distance of 30° - 160° from the center of our study area (12°N 39.5°E). We processed vertical seismograms for all stations between 4°N and 20°N latitude, and 34°E to 45°E longitude (fig. 2.1a). Our final dataset includes 415 stations within 17 networks (see Data Availability Statement).

We briefly explain the ASWMS methodology and quality controls used. For the full procedure, see (79). For each earthquake, vertical cross-correlograms, $C(t)$, were

computed in the time domain for each possible station pair according to:

$$C(t) = S_1 \star W_S S_2 \quad (3.1)$$

where S_1 and S_2 are the de-trended, windowed, vertical seismograms of the two stations with instrument response removed, and $\star$ denotes cross-correlation. W_S is a 300 s windowing function, with a 75 s Hanning taper. $C(t)$ contains lag and coherency information about the Rayleigh wave energy traveling between the two stations. The dominant energy of the waveform was isolated using another windowing function, $W_c(t)$, and then filtered using a series of Gaussian narrow band filters. By fitting the filtered, windowed cross-correlogram function, $F_i \star (W_c C(t))$ using a 5-parameter Gaussian wavelet we extracted frequency-dependent phase delays, t_p , for this station pair (79).

Station pairs with inter-station distance (in km) less than 5, or greater than $3.55 \times$ the period (in seconds), were discarded to reduce the possibility of cycle skipping or spurious measurements. Station pairs with cross-coherency ≤ 0.6 were discarded to prevent measurement on poorly correlated cross-correlograms. t_p were then inverted using the Eikonal equation to develop phase velocity surfaces at each period for each event (80). Regularization for this step was implemented as in (79), with consequent resolution as demonstrated by synthetic recovery tests (Section 3.2, fig. A.2). Finally, these phase velocity surfaces were stacked (on a per-pixel basis; see below) over all events, thereby piecing together regions of temporally and spatially isolated station coverage.

ASWMS was initially designed for the large, evenly spaced US Transportable Array (79). Use of this method for smaller arrays is feasible (81; 82; 83) but requires careful quality control. As such, events for which more station pairs were discarded than not were disincluded from the stacks. We also culled individual phase delay measurements that had apparent velocities below 3.1 km/s or above 4.8 km/s, as velocities outside these

bounds are unrealistic for our period range (84; 85). Testing demonstrated that expanding these limits by 0.2 km/s did not affect the final result. Pixels were included in the final stack (fig. 4.1) if the number of rays crossing those pixels from all event maps exceeded 66 (providing on average one ray entering each pixel per km of circumference). We also excluded pixels with phase velocities more than three standard deviations from the stack mean. For the stacking, each event’s phase velocity map was weighted proportionally to the number of inter-station rays measured for that event. Maps were smoothed using a linear function with a lengthscale of $0.4\times$ the wavelength.

3.2 Resolution

To assess the resolution of our phase velocity tomography, we conducted checkerboard (fig. 3.1), (fig. A.2), and semblance (figs. A.3, A.8) tests following (86). We construct input checkerboard maps using 2-D sinusoids that vary in amplitude between 3.8 - 4.2 km/s. We varied the checker length scale linearly with period to approximately match the Rayleigh wave wavelengths: the sizes range from 0.9° (25 s) to 3.6° (100 s). Using the same rays at each period as in the real data, we forward modeled inter-station delay times, adding ray-length-dependent noise by randomly perturbing the true ray-path averaged velocity, with standard deviation of 0.05 km/s. This level of noise was found to replicate the fraction of quality control measurements discarded in ASWMS for the real data. The checkerboard tests show velocity anomalies are recovered well at all periods (fig. 3.1, S2), with a small amount of NE-SW smearing in Afar, a region which is not the focus of the present study. We used these checkerboard tests to calculate the semblance at each period (87) (figs. A.3, A.8), and mask regions with <0.7 semblance in our results (fig. 4.1). To assess depth sensitivity of phase velocity at each period to shear velocity structure, we compute V_s sensitivity kernels using our final (i.e., most applicable for the

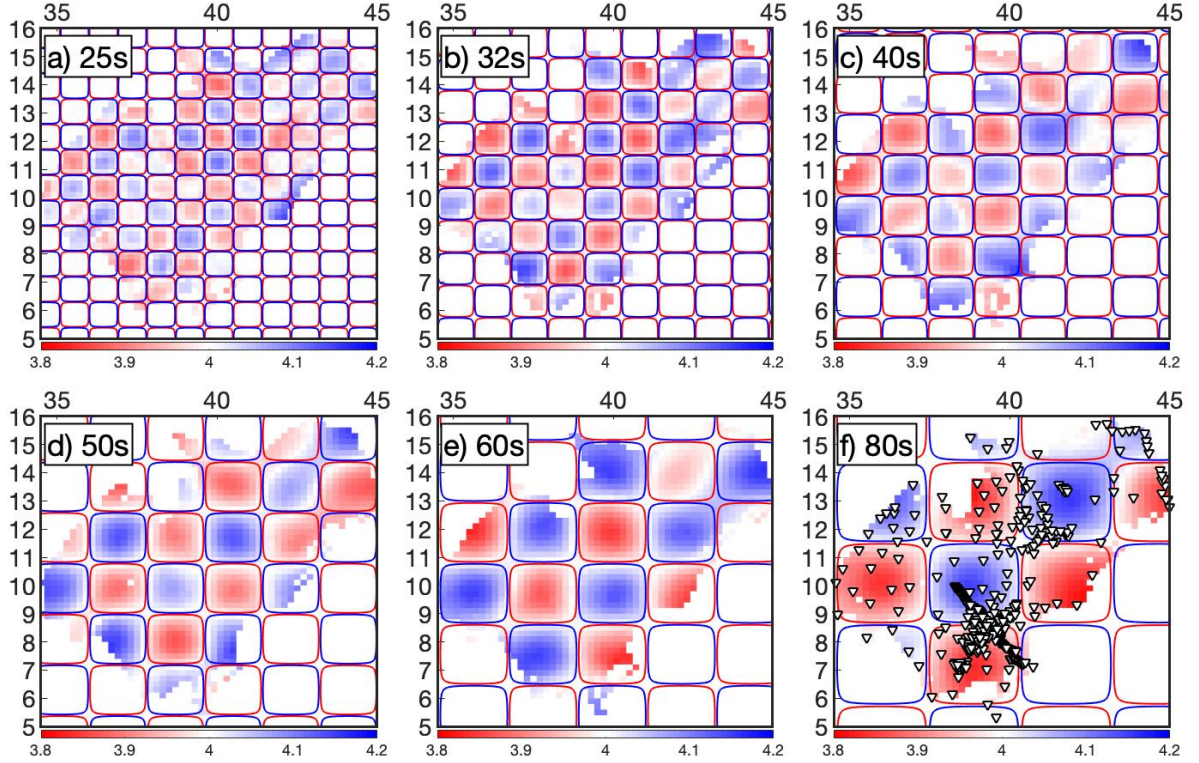


Figure 3.1: Checkerboard tests at 25, 32, 40, 50, 60, and 80 seconds, with the x-axis as longitude, and the y-axis as latitude. 100 s is available in appendix A (fig. A.2), and displays similar resolution. Each period has the input checkerboard displayed as dark red (3.8 km/s) or blue (4.2 km/s) contours, atop the checkerboard surface, with the colorbar below set the same at each period. At 80s, all stations used in checkerboard tests are plotted in inverted white triangles with black edges.

area) shear velocity models from the joint inversion (section 3.5). Our highest overall sensitivity to V_s is between 25 and 200 km (fig. A.17, zipped supplement).

In order to assess the veracity of observed low phase velocities near Lake Tana, we calculate a smear/spike test. A circle roughly the size of the Lake Tana velocity anomaly was set as 0.2 km/s higher than the surrounding area, and this structure was input into the ASWMS checkerboard algorithm at each period. Spike tests (fig. 3.2) reveal that the input structure is well recovered at all periods, but substantial smearing occurs at periods <60 s. Additionally, we find that a slightly fast velocity anomaly ($\sim 4.05 - 4.1$ km/s)

appears within the Afar region, which occurs with and without the addition of noise. We conclude that this is due to covariance within the inversion itself. However, this feature is not cause for concern, as we do not interpret the Afar region in this research.

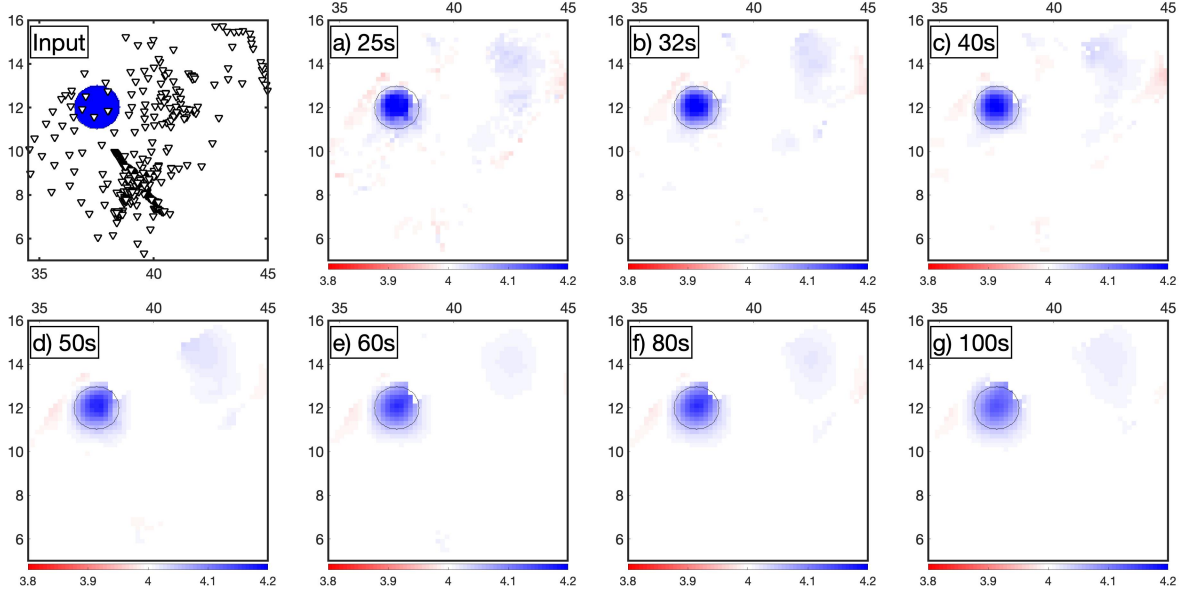


Figure 3.2: Spike tests using a static velocity anomaly, with the x-axis as longitude, and the y-axis as latitude. a) The input model, with each station used in the development of phase velocity stacked maps. The anomaly is 4.2 km/s, with all surrounding area set at 4.0 km/s. Spike tests are shown at 25 (a), 32 (b), 40 (c), 50 (d), 60 (e), 80 (f) and 100 s (g). All use the same velocity bounds between 3.8 - 4.2 km/s, shown on the colorbar beneath each plot.

3.3 Receiver function quality control

Receiver functions (RFs) were calculated at several recently deployed stations on the EP and rift flank as well as a few long-standing stations within the rift (fig. 2.1a, indicated in blue, green, and red). We used $M_w \geq 5.5$ earthquakes 30° to 90° from the approximate center of the study region (12°N , 39.5°E) between 2000-03-10 and 2017-10-03. We extracted seismogram traces spanning ± 100 s from the predicted P -wave

arrival. Traces were de-meaned and bandpass filtered from 0.01 Hz - 0.25 Hz using a 2 pole Butterworth filter. We culled events with a signal-to-noise (SNR) below 1.5, where signal and noise were computed as the maxima of envelope functions based on 8 s and 26 s smoothing of the whole trace, respectively (88). Waveforms were rotated into Z,R,T orientation and then migrated to the median ray parameter of the earthquake. We used the iterative time domain deconvolution technique (89) to calculate receiver functions.

3.3.1 Receiver Function Quality Control

On the basis of an *a priori* assumption that first-order crustal structure is one-dimensional, we assume that highly similar RFs reflect the true 1-D crustal impedance structure. Other more disparate and complex RFs are either caused by noise or reflect more complicated 2-D structure which cannot be recovered by this study. While strongly anisotropic crust or dipping layers may produce receiver functions with bimodal phase arrival times or strong back-azimuthal variations (e.g., *Savage 1998; Hammond, 2014*), we did not encounter this in our data.

We used a multi-stage process to select a family of receiver functions with similar waveforms, relying partly on the Density Based Spatial Clustering of Applications with Noise (DBSCAN) algorithm (90). DBSCAN is a path-specific hierarchical algorithm that uses two parameters to define clusters. The first is a variable distance metric, ϵ , which defines the maximum tolerable difference between two RF traces within the same cluster. The second is the minimum number of data to form a cluster, which we set at a constant 30% of the number of RFs for a given earthquake, a value was found to adequately differentiate families of good RFs from noise. In practice, we found that varying this value between 25% - 75% made little difference (fig. A.9).

Our quality control procedure comprised six steps: (i): We calculated the L_1 norm of

the differences between all receiver functions for a given station (fig. 3.3a), and used those values as distances subject to ϵ . (ii) We used DBSCAN with a high ϵ (lenient) culling step on the RFs windowed narrowly around the expected P arrival, from $T = 0$ to $T = 1.5$ s. This discriminated a cluster of RFs with “normal” P -wave arrivals from those with abnormally late or early arrivals, which were discarded (fig. 3.3b). (iii) We culled receiver functions with amplitudes less than $0.5\times$ the mean amplitude at $T = 0$. (iv) RFs that passed these culling steps were again processed using DBSCAN, this time with a time window from $T = 2$ s to $T = 15$ s, which included the Ps and $PpPs$ arrival phases for most receiver functions. We excluded the $PpSs + PsPs$ phases in this clustering step as they generally had lower amplitude arrivals. Within this step, we changed ϵ until a cluster of RFs were found that had a clear Ps arrival pulse, and the number of clustered traces did not change by increasing ϵ (fig. A.10). (v) After a cluster of RFs were found, we culled traces with crustal multiple phase ($PpSs + PsPs$) amplitudes greater than 30% of the trace’s P arrival amplitude, which is greater than expected for such phases (e.g., *Juliá, 2007*). (vi) Finally we visually assessed each station’s RFs, and culled RFs with an unclear Ps arrival (fig. 3.3c). The stacked receiver function (fig. 3.3d) used in the joint inversion (Section 3.5) was the linear mean of the ‘cleaned’ receiver functions that passed the above quality controls.

3.4 H- κ Stacks

H- κ stacking provides additional constraints on crustal thickness and Vp/Vs ratio by grid-searching a range of Moho and κ values, in order to calculate the likelihood of each Moho- κ pair based on the expected arrival times of seismic phases. H- κ pairs that well-predict the arrival times of seismic phases recorded by receiver functions are indicated by a higher amplitude on the H- κ grid surface. Receiver functions at 28 stations

that passed the “cleaning” process were used in a standard H- κ stacking procedure (91) to estimate crustal thickness and Vp/Vs ratio. We used a Vp of 6.3 km/s, and phase weights of 0.7, 0.2, and 0.1 for the Ps , $PpPs$, and $PpSs + PsPs$ phases, respectively. The resultant H- κ stack for each station was used to constrain crustal thickness during joint inversion (Section 3.5, fig. 3.4c). H- κ stacks for each station are provided in the zipped supplement.

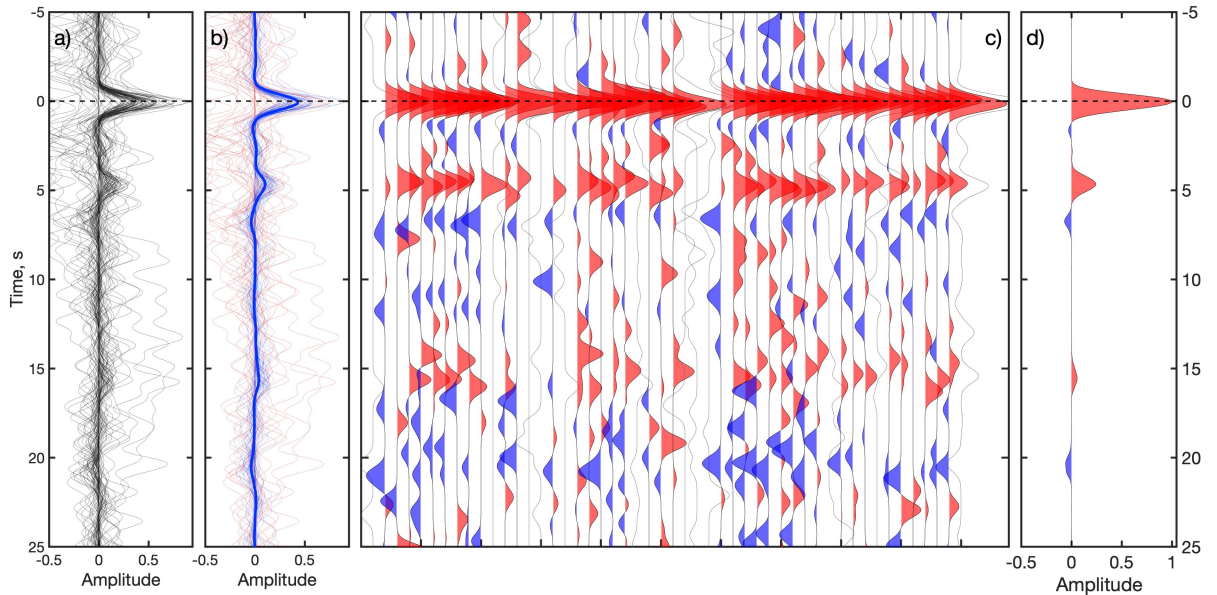


Figure 3.3: A graphical representation of the steps in our quality control process at station KIRE. a) The 104 raw receiver functions. b) The results of stages (i) and (ii) of the quality controls. Thin red lines are receiver functions culled by DBSCAN, thin blue lines are receiver functions passing quality controls, and the thick blue line is the mean of the clustered receiver functions. c) The manual visualization and cull step (vi) after clustering. Traces with colored red and blue fills, signifying positive and negative phases, were used for the final stack, others were manually removed. d) The final stacked receiver function from the 38 receiver functions that passed quality control, normalized to the parent phase amplitude.

(21) found that many H- κ derived crustal thickness and Vp/Vs ratio estimates were unreliable within the EP. They found that stations within flood basalts produced unreliable H- κ stacks with high standard deviation and broad minima, due to low RF SNR and

broad, low amplitude arrival phases. They attributed this to a gradational Moho, complex crust, or sediments and basalts. We performed a modified version of (21)’s random input perturbation procedure on several stations in the region using receiver functions from before the quality control process and those that passed the quality control process (fig. A.11). We found that a distinct high energy maximum was able to be well resolved after quality control for most stations (26 of 28), and be moderately resolved for the rest. We note that strong anisotropy may impact the H- κ process (*cf. Hammond, 2014*), or might cause the quality control algorithm to reject back-azimuthal swaths. However, while our earthquake arrivals are not uniformly distributed, we do not find evidence of systemic back azimuthal patterns or automatic rejections.

3.5 Joint Inversion Algorithm

We used a trans-dimensional, hierarchical, Bayesian inversion algorithm (33) to create 1-D shear velocity profiles of lower crustal and mantle structure. This algorithm uses Bayesian statistics and Monte Carlo Markov Chains to explore the model space, developing ensembles of shear velocity models that closely predict observed Rayleigh wave phase velocities and receiver functions. To ensure we adequately explored the parameter space for each station, we ran 6 parallel Markov Chains for 21,000 iterations, approximately $16\times$ the number of iterations required for a given chain to achieve stationarity (the “burn-in”). Proposed model updates were made by drawing from local Gaussian distributions for single parameters in each iteration. Model perturbations were accepted or rejected using a Metropolis-Hastings acceptance criterion (92; 93), and were confined by *a priori* distributions determined by generous geologically plausible ranges. An empirical prior was developed which confirms that this algorithm does not systemically introduce bias (fig. A.13). We adapted the method of (33) to incorporate H- κ surfaces as an additional

data type. To some extent, this step simply increases the power of the RF data within the joint inversion. However by implicitly migrating crustal multiples in a more self-consistent manner than in a synthetic RF stack, it sharpens the inversion's constraints on crustal Vp/Vs ratio, for essentially zero increased computational expense.

We assumed that errors between different data types are uncorrelated (94; 95; 96; 97; 98; 99), so the overall model likelihood function is computed as the product of likelihoods for each individual data type:

$$p(\mathbf{d} | \mathbf{m}) = \prod_i p(\mathbf{d}_i | \mathbf{m}) = \prod_i \frac{1}{(\sigma_i \sqrt{2\pi})^{n_i}} \exp\left(-\frac{\Phi_i}{2\sigma_i^2}\right) \quad (3.2)$$

where σ_i is the data error for data type $\mathbf{d}_i$, n_i are the degrees of freedom, $\mathbf{m}$ is the proposed model, and Φ_i are misfit functions for body waves, surface waves, and H- κ surfaces, respectively, as follows:

$$\Phi_{\text{BW}}(\mathbf{m}) = \|\mathbf{z}_p(t, \mathbf{m}) * \mathbf{R}(t) - \mathbf{r}_p(t, \mathbf{m}) * \mathbf{Z}(t)\|^2 \quad (3.3)$$

$$\Phi_{\text{SW}}(\mathbf{m}) = \|\mathbf{C}(f) - \mathbf{c}_p(f, \mathbf{m})\|^2 \quad (3.4)$$

$$\Phi_{\text{H}\kappa}(\mathbf{m}) = \frac{\eta_{\text{max}}}{\eta(H_p, \kappa_p)} \quad (3.5)$$

where $\mathbf{R}(\mathbf{t})$ and $\mathbf{Z}(\mathbf{t})$ are the radial and vertical observed time series of stacked receiver functions, $\mathbf{r}_p$ and $\mathbf{z}_p$ are corresponding predicted time series, and $*$ is the convolution operator. Note that this “cross-convolution” body wave misfit equation does not require synthetics to match the true source time function (100; 101). $\mathbf{C}$ and $\mathbf{c}_p$ are the phase velocity observations and predictions at frequency f . η is the amplitude of the H- κ

surface, and H_p , κ_p , are the proposed model Moho thickness and V_p/V_s ratio.

Body wave data were forward modeled using a propagator matrix code (102; 103; 104) which creates synthetic Z,R,T waveforms from a 1-D layered input velocity model. We used a 1-s Gaussian synthetic source time function and a ray parameter averaged from the RFs in the station stack. Only the first 10 seconds of RFs were used, which contain the P and Ps arrival phases. Surface wave phase velocities were forward modeled using MINEOS from CIG (Computational Infrastructure for Geodynamics) (105). MINEOS is computationally expensive, so data prediction relied on perturbation kernels calculated infrequently (and updated per (33) using MINEOS whenever the current accepted model departed substantially from the most recent kernel basis model). Because MINEOS requires a whole Earth model, our models linearly graded into the PREM global model (106) between 200 km and 300 km, and basal knots were prevented below 160 km to ensure no unrealistic sharp velocity gradients were involved in grading into the global velocity average.

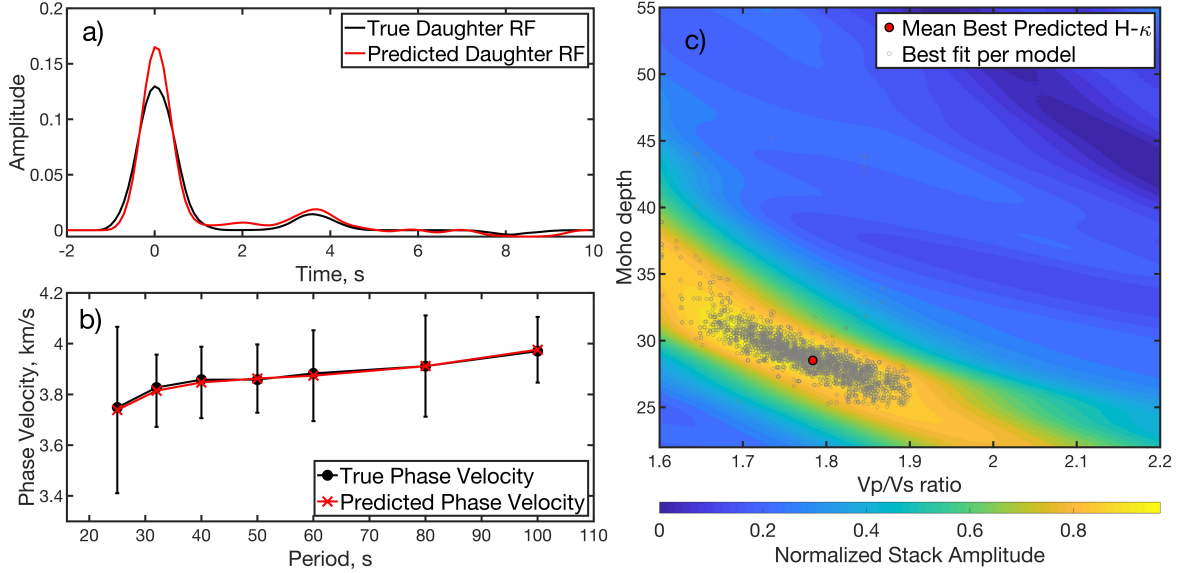


Figure 3.4: An example of the joint inversion data types at station GUBA, including final data fit. a) Stacked P-wave radial receiver function (black). The predicted receiver function computed from the mean posterior shear velocity profile following inversion is shown in red. b) Rayleigh wave dispersion curve at this station (black) and standard deviation at each period, together with MINEOS-predicted phase velocities from the mean posterior shear velocity profile (red). c) H- κ amplitude surface (η), with the corresponding values from the entire posterior model ensemble (light grey dots), and the mean of this posterior (red dot).

3.6 Shear Velocity Aggregate Groups

On the basis of the observed shear velocity profiles (Section 4.3), we determined that stations could be assimilated into three characteristic groups that well span the range of rifting states found in our dataset. We tested clustering the stations into subgroups to identify trends and simplify interpretation. To form these clusters, we started with N random groups, and for 1000 iterations, we randomly moved a single station to a different random group. We then calculated the summed absolute misfit between individual profiles and their group mean $V_S(z)$, and added the total misfit across all the groups. If this misfit decreased from the previous iteration, indicating overall more similar intra-group

velocities, we accepted the new group assignment. We tested this scheme for $N = 2$ to 5 groups and found that $N = 3$ achieved the best balance of minimizing intra-group variability, while maintaining parsimony and geographic coherency. Our preferred groups of stations are as follows: Group 1 encompasses 10 stations in the NWEF, Central EP, and Somalian Plateau (fig. 2.1a, blue). Group 2 comprises 12 stations the central-NWEF (fig. 2.1a, green). Group 3 includes 5 stations in the MER, and the MER-adjacent EP (fig. 2.1a, in red). Station AMME (8.3031°N , 39.0934°E) was manually placed into group 3 because of its location in the MER instead of group 2, to which the shear velocity structure more closely matches. For this analysis, we ignored station DAQU (13.1448°N , 37.8976°E) which had a shear velocity structure that does not closely match any group, despite the joint inversion yielding a good data fit (fig. 2.1a, in grey).

Chapter 4

Results

4.1 Phase Velocity Maps

Phase velocities in the NMER and SMER follow the surficial border faults of the MER at short periods (25-32 s), and are slower than much of the EP and the average phase velocity from 25 - 80 s (fig. 4.1a). The CMER appears up to 0.1 km/s faster than the SMER and NMER from 32 - 60 s, but remains slower than the EP by 0.1 km/s and Somalian Plateau by 0.2 km/s. While we note that the error threshold in the CMER is 0.14 km/s, regions we compare it to show lower error (figs. A.7, A.8). We can, however, resolve that the CMER remains substantially slower than the EP by 0.05 - 0.2 km/s, and Somalian Plateau by 0.1 - 0.3 km/s between 32 - 60 s. The region of slow velocity extends further west from the MER into the EP with increasing period. This does not occur east of the rift, within the Somalian Plateau, as observed previously (e.g., *Bastow et al., 2005; 2008*). By 80 s, the MER, the Somalian Plateau, and eastern EP homogenize to phase velocities of 3.75 - 3.8 km/s, which remain reduced compared to the westernmost EP.

Slow phase velocities extend in a lineament east-southeast from Lake Tana at periods of 25 - 60 s (fig. 4.1a - e). This lineament roughly follows the trend of inferred late

Paleozoic-Mesozoic aged rifts (45), Quaternary volcanoes, and spatter cones (30). Phase velocities here are ~ 0.1 km/s faster than the MER at 25 s (~ 3.5 km/s, vs. 3.4 km/s). In the north, this structure links to slow phase velocities at the western boundary of Afar from 25 - 32 s. This structure grades into less reduced, but still slow, phase velocities to the southwest of Lake Tana between 25 - 50 s. This slow phase velocity lineament does not broaden with period, and homogenizes to the fast background NWEF phase velocity by 80 seconds (fig. 4.1f).

A pronounced ~ 200 km E-W, ~ 160 km N-S, fast phase velocity region in the central EP at 32 s separates the slower phase velocities in the MER and EP lineament (fig. 4.1b). Velocities here are $\sim 4.0 \pm 0.14$ km/s, $\sim 8\%$ faster than the $\sim 3.7 \pm 0.15$ km/s area average at 32 s. This appears to result from a shallow feature: the phase velocities in this location are not noticeably faster than the regional average at periods greater than ~ 50 s, although at longer periods most of the EP and flank are ~ 0.1 km/s faster than the study region average. Fast phase velocities also exist in the Somalian Plateau from 25 - 50 s ($\sim 3.7 - 3.9$ km/s). At periods > 60 s, the phase velocities on either side of the MER are roughly symmetrical at $\sim 3.7 \pm 0.17$ km/s, and the rift-proximate part of the Somalian plateau that we image is clearly slower than the rift-distal NWEF (3.95 ± 0.15 km/s).

4.2 Receiver Functions

The number of total receiver functions at each station varied between 9 - 165, and averaged 65. The fraction of receiver functions culled by our automatic quality controls averaged 51%, and ranged between 26 - 82% of recorded traces. On average, manual culling removed a further $\sim 16\%$ of traces. The number of RFs that passed quality controls to develop the final stacked RF varied from 2 - 73, with a mean of 23. The

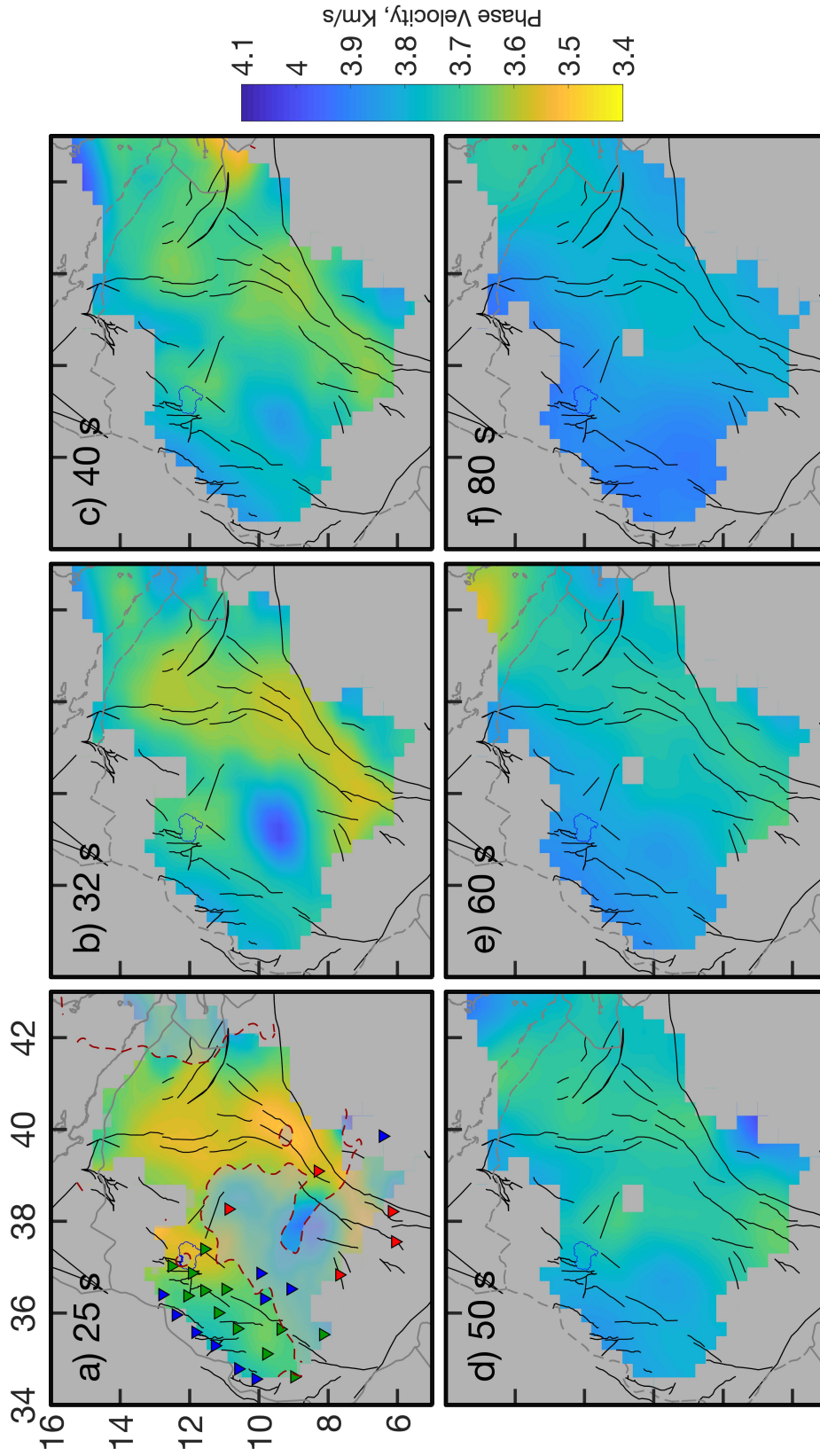


Figure 4.1: Phase velocities at 25, 32, 40, 50, 60 and 80 s from stacking of 5741 individual earthquake phase velocity maps, in km/s, with the x-axis as longitude, and the y-axis as latitude. Warmer colors indicate slower phase velocities. Lake Tana is outlined in blue, and faults from figure 2.1b) are plotted at each period. Stations inverted for shear velocity are plotted at 25 s, corresponding to their shear velocity aggregate group (Section 3.6). Regions with semblance <0.7 are plotted with high transparency, and red dashed lines denote the distinction. Note that periods greater than 25 s all have >0.7 semblance.

quality control process summarized in fig. 3.3 may be found for each inverted station in the zipped supplementary file.

All stacked receiver functions include a distinct Ps phase arrival at ~ 5 s (fig. 3.3d), and 25 show a clear $PpPs$ phase arrival (fig. 3.3d, ~ 16 s). 21 stations displayed a $PpSs + PsPs$ arrival phase (fig. 3.3d, ~ 21 s), which was often broader and lower amplitude than preceding phases, potentially caused by stacking across moveout, or more local issues outlined in (21), including non-simple crustal structure, gradational Mohos, or flood basalts overlaying sediments. We do note, however, that no stations lacking the $PpSs + PsPs$ phase arrivals failed the test criteria posed in (21), supporting our argument that (notwithstanding tradeoffs) crustal thickness may be determined here using H- κ stacking, even without these later crustal multiples.

A negative phase following the Ps arrival is observed at 16 stations (*e.g.*, fig. 3.3d, ~ 6 s). All group 1 station RFs record this phase, but it does not arrive at a consistent time. It is absent from all group 3 stations except AMME, where it immediately follows the Ps phase, and has a large negative amplitude. More rigorous investigation of these negative velocity gradients, which may reflect layering within or at the base of the lithosphere (107), is beyond the scope of this research. Because of the time window used for RFs in joint inversion, these phases must also be accounted for in the body wave forward modelling (fig. 3.4).

4.3 Joint inversion

We inverted for shear velocity models at 28 station locations within the EP, NWEF, and MER, which are most relevant to interrogating the possibility of off-axis extension (fig. 2.1a). Stations in the CMER and SMER (fig. 2.1a, in red) serve as points of comparison between our results and previous studies (*e.g.*, *Dugda et al., 2007, Gallacher*

et al., Weeraratne *et al.*, 2003). Overall, our inversions fit the data well (fig. 3.4). With misfit defined as in equations 3.3-3.5, we find a median RMS misfit for RFs of 0.054 parent-normalized amplitudinal units, for phase velocities a median RMS misfit of 0.0453 km/s, and for the HK surfaces, on average our posterior models are at or above 94.9% of the H- κ energy surface maximum. The closer fit to smooth phase velocity curves compared to the RFs is typical to these types of inversions (108).

4.3.1 Crustal Thickness

We report the median and 1σ (actually the $\pm 17^{th}$ percentile of the posterior ensemble about the median) crustal thicknesses at inverted stations (fig. 4.2a). Crustal thickness values are generally well constrained, with the a mean 1σ uncertainty of $1.68 \text{ km} \pm 0.67 \text{ km}$, a minimum of 0.84 km, and a maximum of 3.65 km. Crustal thicknesses are similar to those found from the H- κ stacking method (fig. A.12), but notable divergences appear at two stations in the central EP (Stations NEKE, HARO), indicating that phase velocities are not straightforwardly compatible with the receiver function constraints here.

The crust is thickest northeast of the central EP, and near Lake Tana, varying from 35 - 40 km. This is also the region with some of the highest Vp/Vs ratios: ~ 1.89 . Stations in the CMER and Somalian plateau also exhibit thick crust of $\sim 37.5 \text{ km}$ and $\sim 36.5 \text{ km}$, respectively. Crust at stations in the SMER is thinner, at $\sim 29 - 31 \text{ km}$. Trending northwest of the SMER axis to the central EP are three stations of similar thickness, $\sim 31 \text{ km}$, amongst $\sim 34 - 36 \text{ km}$ thick crust.

Crustal thickness is bimodal along the slow phase velocity lineament, split between $\sim 29 - 30 \text{ km}$ and $\sim 33 - 34 \text{ km}$. Crust in the NWEP is between 28.5 - 31 km. We find that our crustal thicknesses are more similar to those of (6) than (3) and (15), but are not strongly inconsistent with the latter studies. A list of crustal thicknesses with associated

standard deviation may be found below.

T1. Moho Depth Table

Station	Moho (km)	Std (km)	Lat °	Lon°
ABMD	29.25	1.52	11.832	35.583
AMME	37.80	1.41	8.303	39.093
ARBA	28.06	1.76	6.067	37.556
ASOS	33.06	2.10	10.091	34.564
BAHI	35.30	1.70	11.574	37.393
CHEF	31.29	1.44	6.161	38.210
CHGE	34.04	2.86	10.961	36.518
DAQU	32.79	1.24	13.145	37.898
DEBR	27.98	1.52	10.635	35.662
DLMN	36.86	3.22	6.424	39.856
GIDA	34.02	2.42	8.980	34.610
GORE	34.34	1.87	8.150	35.533
GUBA	28.52	1.32	11.271	35.289
GUBL	32.58	1.20	11.179	36.005
HARO	32.00	1.28	9.847	36.310
HDBU	34.55	0.84	9.376	35.651
JAWI	31.48	1.28	11.571	36.491
JIMA	30.88	3.65	7.684	36.831
KIRE	37.23	1.63	9.957	36.869
MELA	28.37	1.44	12.396	35.960
MEND	31.81	0.96	9.787	35.111
MERT	36.59	1.71	10.876	38.266
MGNA	28.80	1.77	12.786	36.404
MUKA	31.89	1.02	12.064	36.374
NEKE	31.22	1.03	9.089	36.523
SERE	38.96	1.89	12.513	37.028
SHAW	33.99	2.13	11.930	36.870
SHER	31.28	1.01	10.594	34.786

4.3.2 Shear Velocities

We obtain a median shear velocity profile at each station from the posterior model ensemble. We focus on interpreting velocities from 30 and 140 km depth for tectonophys-

ical relevance (fig. 4.2b-e), but our phase velocity sensitivity extends to ≥ 165 km (fig. A.17). We note that at 30 km the observed shear velocities represent crustal structure at most stations (fig. 4.2a). The 1σ shear velocity uncertainties between 60 and 120 km are approximately ± 0.05 km/s across all inverted stations. 1-D shear velocity profiles at each inverted station, with 1σ and 2σ uncertainties, may be found in the appendix (fig. A.16) and zipped supplement. We compare stations' shear velocity profiles to PREM (106), and also to the average shear velocity of the inverted stations at depth, since we find that the entire region is slower than the global average. For the grouped velocity profiles, we calculated median and 32 - 68% percentile velocity ranges from the combined posterior distributions concatenated across all stations within the groups.

Shear velocities closely mirror the trends seen in phase velocities (fig. 4.1). Slow shear velocities exist in the CMER, SMER, and near-rift EP stations, but also near Lake Tana, and southeast of it along the EP. Fast velocities are resolved within the central EP, Somalian Plateau, and NWEF.

For group 1, the median V_s profile (see section 3.6, figs. 4.2a, 2.1a) is faster than the regional average at all depths. 1σ bounds of group 1 shear velocity span 3.0 - 4.3% (0.13 - 0.19 km/s) slower than PREM from 60 - 120 km (fig. 5.1a, d). This depth range is later used as representative of each group due to its consistency between stations, as measured by similarity of shear velocity profiles within each group (fig. 5.1).

Four of six northwestern-most NWEF stations are faster than the regional average at 50 km by 1.2 - 4.8% (4.3 - 4.45 km/s) (fig. 4.2b). At 70 km depth and below, all six northwestern stations are 0.5 - 5% faster (4.27 - 4.46 km/s) than regional average (fig. 4.2d, e). Stations within the central EP manifest heterogeneous mean shear velocities between 30 and 50 km, but this region is 0 - 4% faster (4.25 - 4.42 km/s) than the regional average deeper than 50 km. The Somalian Plateau is faster than the regional average between 50 - 100 km by 1 - 3.5% (4.3 - 4.4 km/s).

Group 2 shear velocities, which include Lake Tana and EP stations to the southwest of it, display slower velocities than those found in group 1. Shear velocities are 4.6 - 6.2% (0.21 - 0.28 km/s) slower than PREM from 60 - 120 km (fig. 5.1b, d). Unlike group 1, group 2 velocity profiles seem to have a thin or absent lid, and deeper than 60 km get faster monotonically with depth. At depths greater than ~ 100 km, they are statistically indistinguishable from group 1.

At 50-70 km, shear velocities in the southwestern EP and near Lake Tana are moderately slower than the regional average by 0 - 2% (4.17 - 4.25 km/s) (fig. 4.2c, d). By 110 km, only the southwestern EP shows reduced shear velocities slower than the regional average by 0 - 1% (4.2 - 4.25 km/s) (fig. 4.2e, f), while other stations throughout the EP are homogeneously faster than the regional average by 0.5 - 1.5% (4.27 - 4.31 km/s).

Group 3 shear velocities, which include the MER and the near-rift EP, are the most reduced compared to PREM; 7.7 - 8.7% (0.34 - 0.39 km/s) slower from 60 - 120 km (fig. 5.1c, d). Between 50 and 110 km depth, these stations are slower than the regional average by 2 - 6% (4.00 - 4.17 km/s), with SMER stations displaying the slowest velocities. Station MERT displays an increase in shear velocity between ~ 40 - 50 km, and then decreases between ~ 50 - 60 km. This may be evidence of a thin fast lithospheric lid, which is not displayed by other stations in group 3.

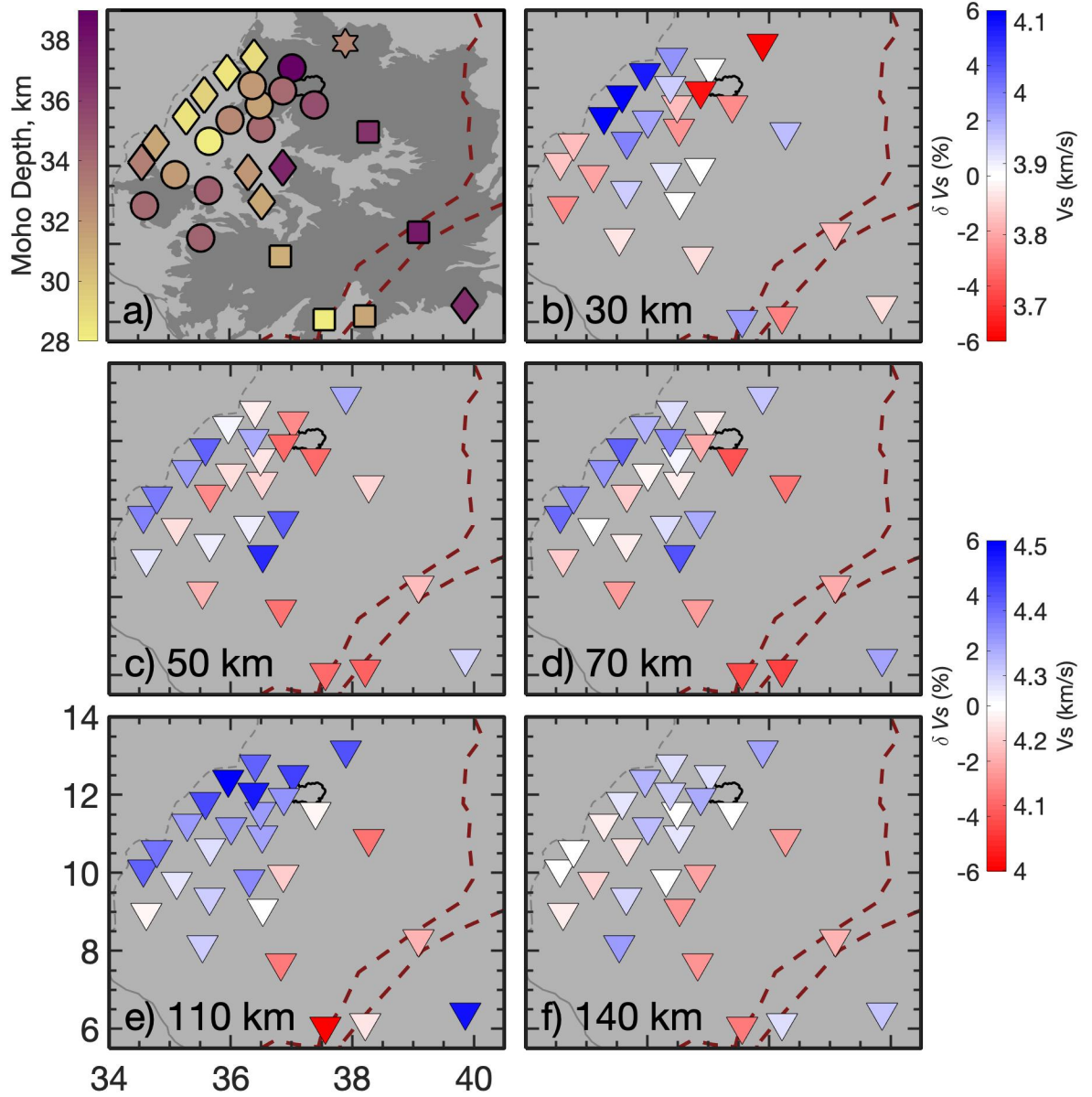


Figure 4.2: a) Post-inversion crustal thickness. Yellower colors indicate thinner crust, redder colors indicate thicker crust. Symbol type indicates velocity group, where group 1 = diamond, group 2 = circle, group 3 = square. Station DAQU is not grouped, and marked as a star. Flood basalt bounds are plotted in dark grey, from (77). b) - e) Median shear velocity from posterior ensembles of inverted stations at 30, 50, 70, 110, and 140 km depth. Color signifies both shear velocity and deviation from the mean velocity at each depth slice $\pm 6\%$. Velocities at 50, 70, 110, and 140 km share the same color scaling, seen to the right of subplots d, f. Thick dark red: rough bounds of the MER rift valley. Lake Tana is outlined in black.

Chapter 5

Discussion

5.1 Regional Shear Velocity Trends

We find mantle shear velocities in this region are slower than PREM and much slower than a global average (109) for nominally Proterozoic (110) continent. We observe this for all stations and velocity aggregate groups, even the “fast” group 1 most distal from the rift (fig. 5.1d). As many others have shown before, plume impingement and continental rifting have combined to produce one of the slowest regions of upper mantle on the planet.

In the Central EP and NWEF (group 1), we find a faster uppermost mantle from ~ 40 - 80 km depth (fig. 5.1a, d). Some individual stations, and our median group 1 V_s profile, hint at a negative velocity gradient below this depth, implying a gradual lithosphere-asthenosphere boundary (LAB). However, our posterior ensemble contains enough range that we cannot identify a statistically significant LAB. Despite this, at some of the stations in this region, RFs also record a negative pulse, as seen in others’ work (12), implying steep negative velocity gradients. Although hardly representative of the entire Somalian Plateau, our single station east of the rift also exhibits the above features. We interpret this as evidence for a lack of substantial extension in the central

EP and, in agreement with previous velocity models (4; 13), a sharp transition to unextended lithosphere immediately east of the rift.

A faster uppermost mantle is also present in the Mozambique Belt (111). We argue that the Mozambique Belt serves as an informative comparative structure (6) from which to assess velocity differences resultant from plume impingement and associated volcanism, despite its non-continuity along the EARS. The similarity in shear velocity structure, but uniformly slower velocities of group 1 compared to the Mozambique Belt, suggest holistic lithospheric velocity perturbations far off-rift (fig. 5.1a, d). At ~ 45 km depth, the group 1 profile is ~ 0.25 km/s slower than the Mozambique Belt. Velocities in the two regions gradually become more similar with depth, and are barely distinguishable by ~ 165 km. The fact that these regions seem to differ particularly in shallow mantle structure demonstrates the role ongoing tectonics play.

Middling shear velocities in group 2 are distinct from group 1 and 3, with stations in this region lacking a fast velocity lid, but also failing to display a negative concave velocity structure between 40 - 140 km (fig. 5.1b, d). Shear velocities are up to $\sim 3.5\%$, or 0.13 km/s slower than group 1, but faster than group 3 by a maximum of $\sim 3.8\%$, or 0.15 km/s. We interpret this region as manifesting an intermediate rifting state between two end-members (the far flank and the MER). We do note that shear velocities within group 2 are more heterogeneous than groups 1 and 3, perhaps a reflection of complex crustal structure. (3) observed a lack of a fast velocity lid in the MER and Afar (fig. 5.1b), and concluded that advanced thermal erosion of the lithosphere had already taken place within the rift (*cf. Rychert et al., 2012*). We similarly propose that the minimal fast velocity lid within the EP slow velocity lineament is indicative of extensional lithospheric perturbation surprisingly far from the rift axis (see section 5.3).

Group 3 evinces particularly slow shear velocities in the upper mantle. Extensive velocity structure perturbation here is consistent with the tectonic and thermal effects

of rifting in the MER and proximal EP. For reference, we compare group 3 to the mean global shear velocity structure of 0 - 25 Ma oceanic crust from the compilation of (109) and find $V_S(z)$ is slower than this reference by 3.6 - 4.3% (fig. 5.1c, d). The fact that these values are even slower than the young oceans likely reflects unusually high mantle potential temperatures and/or less mature melt percolation network beneath still-extant continental crust, as well as perhaps a high-biasing of the oceanic regional average by cooler off-ridge structure. Taken at face value, and attributing velocity reduction entirely to partial melt, the $\sim 4\%$ (0.16 km/s) offset reflects an excess of between 0.5% and 4% *in situ* melt compared to the young oceans depending on the chosen V_S - % melt relationship (112; 113; 114). The similarity in velocity profile shape evidences that the maturation of the rifted asthenosphere significantly precedes crustal breakup here (6; 115).

Group 3 has the best (albeit limited) geographic overlap with previous studies, with which we find good agreement. For instance, the profile of group 3 is nearly indistinguishable from station ARBA in the work of (3) (fig. 5.1c), except for a more negative velocity gradient seen between 35 - 60 km depth. Compared to the MER shear velocity aggregate from (11), we find dissimilarity between 30 and 60 km, after which both profiles are nearly indistinguishable. (11) sampled mostly the NMER, CMER, and Afar, which capture a more extensively rifted, and thus slower, upper mantle structure.

5.2 Quantifying the Contributions of Melt and Temperature

Pervasive low velocities throughout this region point to systematically elevated temperature (3), a more silicic mantle composition, grain size reduction, the presence of melt and other fluids, or a combination of these factors (e.g., *Faul & Jackson, 2005*). Compo-

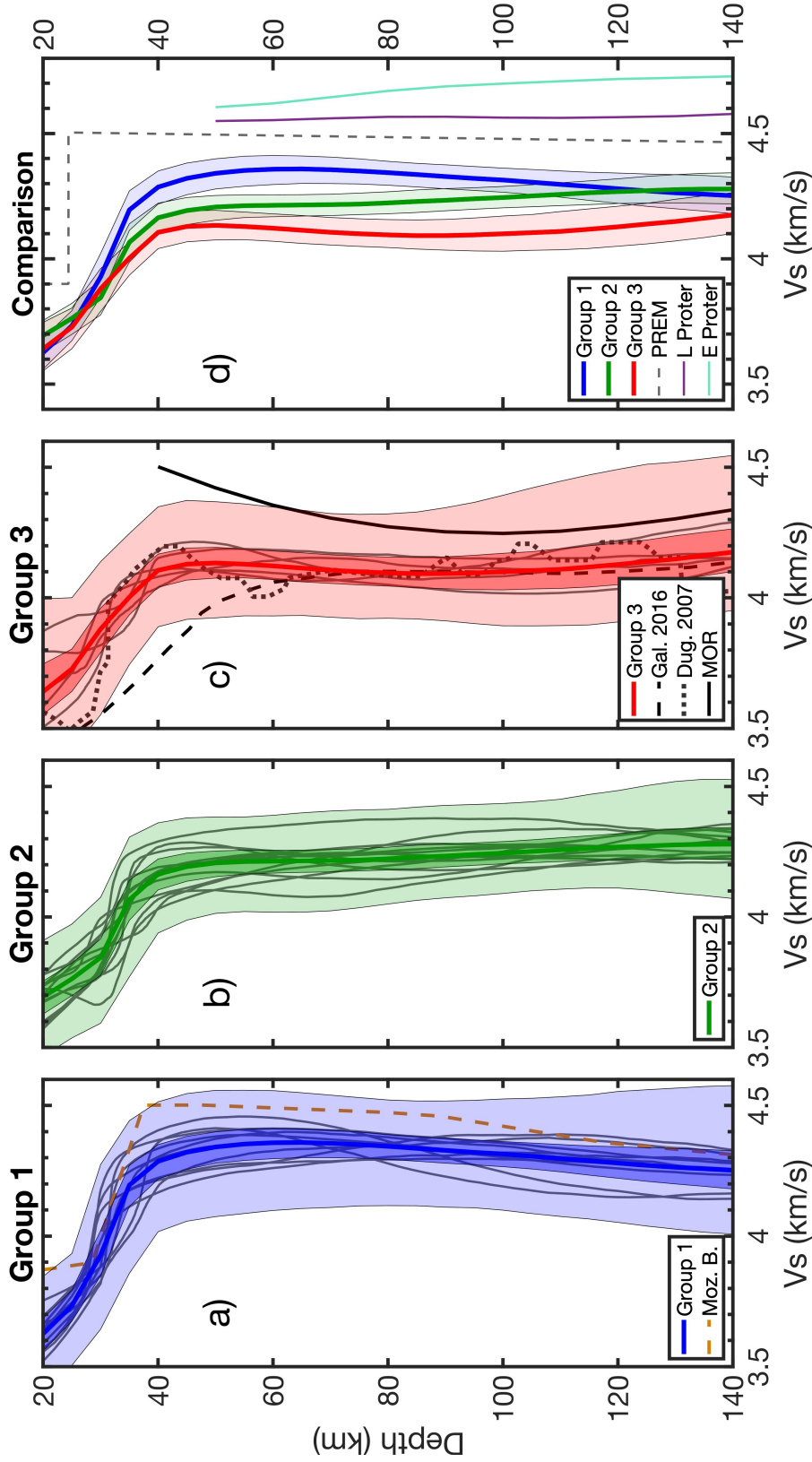


Figure 5.1: Grouped shear velocity profiles for all stations jointly inverted. a, b, c) show groups 1, 2, and 3, respectively, where light grey lines are individual median V_s profiles of the aggregated stations. The median group V_s profile is shown in thick colored lines, while dark and light shaded patches correspond to 1σ and 2σ bounds developed by combining model posteriors for all stations in the group. a) also shows shear velocity profiles for the Mozambique Belt (dashed orange) (111). c) also shows the shear velocity profile of station ARBA (3) (dotted black) and the MER aggregate velocity of (11) (dashed black). d) compares the median $\pm 1\sigma$ profiles of all groups together with reference models to PREM (106), Late Proterozoic crust (purple), and early Proterozoic crust (teal) (109).

sitional variance between pyrolite, piclogite, and harzburgite can account for only $\sim 1\%$ δV_s (116), and grain size reduction from 1 cm to 1 mm at 1200 °C reduces velocities a maximum of 0.1 km/s (117), or $\sim 2\%$ δV_s (6). Our data therefore likely constrain the combination of elevated mantle temperatures and/or *in situ* melt fraction. (3) explained slow MER shear velocities by invoking ~ 250 °C temperature increase caused compared to the Mozambique Belt. Other tomographic studies posit even hotter temperatures, 160 - 550 °C above normal mantle (13). We attempt to quantitatively connect our shear velocity observations to mantle conditions in two ways. First, by generating a suite of geotherms and forward-modelling shear velocity profiles, seeking a best fit to the average $V_s(z)$ of each group. Second, we take a simpler approach, positing that independent effects of excess temperature (dT) and *in situ* melt fraction (ϕ) can explain slowness compared to PREM (δV_S), at some representative depth.

5.2.1 Forward-Modeled Geotherms fit to Observed V_s

In method 1, candidate geotherms are constructed by grid searching over a range of mantle potential temperatures (T_p), surface heat flows (q_0), and radiogenic crustal heat production values (A_0) (figs. A.21, A.22, A.23, table A.14). We use the formulation of (118) to compute $V_s(T, z, d)$ at 1 Hz. We assume a grain size, d , of 1 mm, obtain anharmonic shear moduli from the tool of (119) assuming a pyrolitic composition, and calculate self-consistent pressure profiles. Continental geotherms cannot fully predict observed shear velocities between 30 and 60 km (figs. A.18, A.19, A.13) using existing scaling relationships (120), so we choose to minimize predicted misfit between 60 and 120 km, the range where our phase velocities are most sensitive, and the majority of melt is expected to reside.

We augment our forward calculation to account for the velocity reduction effects

of partial melt. We use super-solidus temperature excess to predict a melt fraction, assuming 0.0995% melt is generated per degree Celsius above the dry peridotite solidus (fig. 5.2a), which was calculated from a pMELTS algorithm curve (121) using a LOSIMG composition peridotite (122), at 3 GPa, with no melt escape (a necessary simplification). We translate this melt fraction into a velocity reduction beyond the melt-free case using the relationship from (113), whereby 1% partial melt slows shear velocities by 3.3% (this value is for penny shaped inclusions, likely more applicable here based on the presence of anisotropy and high differential stress from rifting (123)). We account for melt percolation effects by convolving the instantaneous melt curve with an exponential function to smooth melt upwards with decay lengthscale of 20 km. We also apply a compaction function downwards at the Moho with a lengthscale of 5 km to mimic the effect of neutral buoyancy of the mafic melt in the crust while ensuring total melt percentage is preserved. This *ad hoc* procedure acts to distribute the effect of any melt across a broader depth range, in keeping with our lack of observed sharp low velocity layers. Our treatment of melt production and percolation is admittedly simplistic, but sufficient for this exercise. A second caveat is that we are combining a melt-free anelastic scaling law with a nominally anharmonic constitutive relationship for the effect of melt on velocity. However, given the absence of a more self-consistent approach and the other uncertainties in this forward modelling approach, we argue that this is a good approximation to balance the effects of temperature and melt on velocity. We estimate uncertainty in melt fraction from the suite of forward models that fit the velocity profiles to better than 95% of the best fit model.

5.2.2 A Simple Melt - Temperature Tradeoff

In method 2, we assume that $\delta V_S = \frac{\partial V_S}{\partial T} \delta T + \frac{\partial V_S}{\partial \phi} \delta \phi$, using derivatives of -0.0012 km/s/K (3; 124; 125) and -0.047 km/s/% (113), respectively. Note that we prefer this model to (114), as discussed above. Consistent with method 1, we use the average V_S between 60 - 120 km for each group (4.33 ± 0.053 km/s for group 1, 4.25 ± 0.048 km/s for group 2, and 4.11 ± 0.063 km/s for group 3) to yield contours of relative slowness for each group in reference to PREM (4.47 km/s Voigt average) (fig. 5.2b). We assume that PREM represents a melt-free mantle with a potential temperature of 1300 °C, and account for a 0.4 K/km adiabat when extending to 90 km depth. Naively, this approach is limited to establishing the linear trade-off between thermal and melt perturbations. However, these parameters are themselves of course linked. We therefore use the pMELTS calculation described above to compute the peridotite melting curve (fig. 5.2b) at 3 GPa (~ 90 km depth), assuming batch melting. Theoretically, the intersection between this curve and the groups' slowness contours constrains mantle conditions. Realistically, due to melt percolation, our estimates of melt fraction are likely maxima, and estimates of temperatures may be minima.

5.2.3 Physical Interpretation of Low Velocities

Using the first method, group 1 is best fit by a geotherm with parameters: 45 km Moho, $Q_0 = 75$ mW/m², $T_p = 1425^\circ\text{C}$, $A_0 = 0.99\text{e}^{-7}$ W/m³. This estimate agrees with independent xenolith-based (7) and geodynamically modelled (10) estimates of local mantle temperatures in the range 1350 - 1450 °C within the MER and Afar. This geotherm does not intersect the solidus of dry peridotite (fig. 5.2a). Similarly, the second method suggests an upper mantle temperature $\sim 100^\circ\text{C}$ above average, that does not intersect the pMELTS melting curve (fig. 5.2b). Our seismic velocities do not seem

to require *in situ* melt here, even when considering error.

This is in agreement with an absence of volcanism in the NWEF. While some thermal springs exist near the central EP, no recent volcanism exists in any location aggregated into group 1. Group 1 stations sit atop the coolest mantle in the region, apparently $\sim 1450^\circ$. That said, ~ 30 Ma xenolith geothermobarometry (126) (fig. 5.2a) and potential temperature estimates of 1410°C from primitive magmas 10 - 40 Ma (7) imply that in past mantle temperatures below this region were lower than those inferred today. Even in the distal flank of the rift, there seems to be a warming trend over time.

Group 2 is best fit by a geotherm with parameters: 30 km Moho, $Q_0 = 75 \text{ mW/m}^2$, $T_p = 1475^\circ\text{C}$, $A_0 = 1.32\text{e}^{-6} \text{ W/m}^3$. This is hotter than the $1350 - 1450^\circ\text{C}$ ambient mantle temperatures estimated for the MER and Afar (7; 10). Group 2 velocity averages are low enough to require the presence of upper mantle partial melt. A maximum of $0.46\% \pm 0.08\%$ partial melt is found at 70 km depth (fig. 5.2a) using method 1. A mantle temperature of $\sim 1500^\circ\text{C}$, and $0.1^{+0.3}_{-0.1} \%$ melt (fig. 5.2b) is predicted using method 2. This 0.1-0.46% partial melt component is slightly lower than the 1.1% melt component invoked by (13) to the east of Lake Tana, possibly because here we group the slower northern and less slow southern stations in the off-axis lineament.

Xenolith data (fig. 5.2a) at Injibara, just south of Lake Tana, contain signatures of hydrous enrichment (66; 126). Xenoliths indicate the mantle here is primarily depleted peridotite (66; 7) which may require hydration for substantial melt absent of very high temperatures (127). Quaternary volcanoes and thermal springs extend southwest from Lake Tana as far as 11°N , following surficial border faults. These features, previously linked to shallow velocity reduction (128; 13), are mirrored by the low velocity lineament we observe in the group 2 footprint (fig. 4.2). Our data support the presence of particularly elevated uppermost mantle temperature, as well as hydration and melt, extending south of Lake Tana.

Group 3 is best fit by a geotherm with parameters: 30 km Moho, $Q_0 = 75 \text{ mW/m}^2$, $T_p = 1525 \text{ }^\circ\text{C}$, $A_0 = 1.65\text{e}^{-6} \text{ W/m}^3$. This is 75 - 175 $^\circ\text{C}$ hotter than some others have proposed (7; 10). This region, close to the magmatic rift, is melt-rich (129; 130; 131). We obtain estimates of up to $1\% \pm 0.18\%$ partial melt at 80 km (fig. 5.2a) using method 1, and a temperature of $\sim 1515^\circ\text{C}$ with $0.95 \pm 0.4\%$ melt is predicted using method 2 (fig. 5.2b). These methods closely agree on a partial melt percentage $\sim 1\%$, which aligns with recent estimates within the MER between 0.5 - 2.7% (6; 11; 13; 14). Fully disentangling seismic velocities, melt fraction, and pore orientation/anisotropy is not fully possible using our methodology. However, recent studies (14) have found radial anisotropy as high as $6.5 \pm 0.5\%$ in the shallow crust (5-15 km) of the near-MER EP, supporting melt storage in the form of sills, in agreement with geochemical studies in the area (132). Strong azimuthal anisotropy, primarily in the rift, also complicates interpretation (e.g., *Hammond et al., 2014; Kendall et al., 2005*). However, good backazimuthal distribution (fig. A.1) likely means that we average out this effect. While volcanoes are confined mostly to the narrow rift valley (fig. 2.1b), our phase velocities are slow in a relatively broad swath extending 100 km beyond (particularly to the west of) the border faults even at 32 - 40 s. This may stem partly from resolution issues, but likely also reflects the nature of melt channelization and focusing within the rift (*c.f Holtzman & Kendall, 2010*).

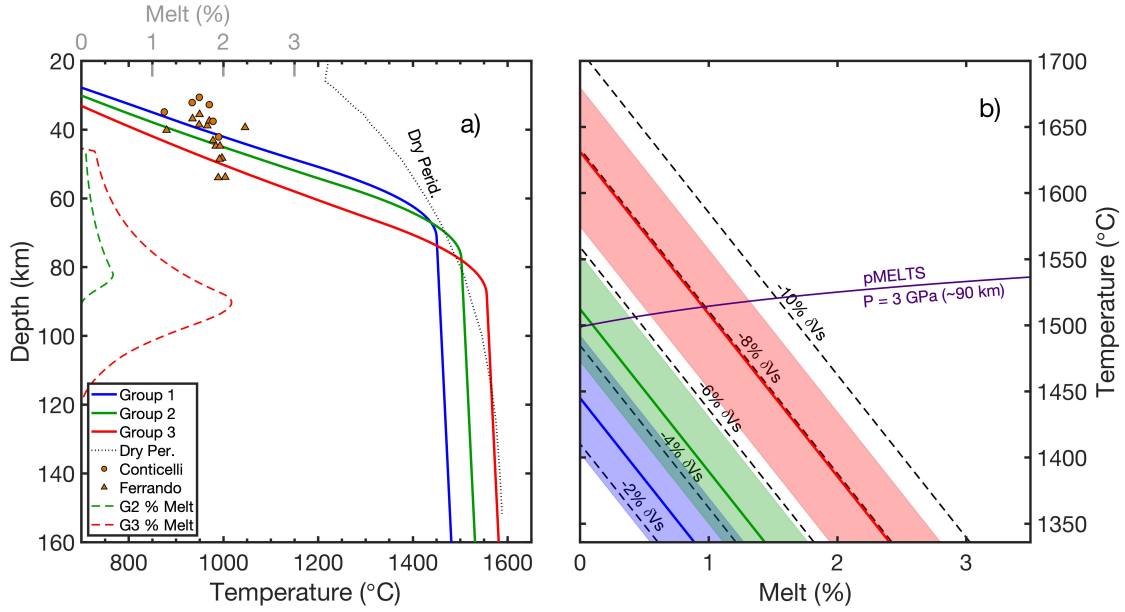


Figure 5.2: a) Best fit geotherms for group 1 (blue), group 2 (green), and group 3 (red) using method 1 to fit shear velocity between 60 - 120 km depth (Section 5.2), with temperatures indicated by the lower x-axis. The dry peridotite solidus from (127) is plotted in black dots. Xenolith pressure - temperature equilibration conditions from (66) and (126) are shown in orange circles and triangles, respectively. Partial melt percent (upper x-axis) calculated using the super-solidus forward modeling method is shown for groups 2 and 3 (dashed, colored as above) to the left. b) The slowness domain, showing linear tradeoff between melt and mantle temperature excess, contoured in increments of $-2\% \delta V_s$. Mean $\pm 1\sigma$ slowness compared to PREM between depths of 60 - 120 km is shown for each velocity group (colors as above). The melting curve of a LOSIMG composition peridotite (122) at 3 Gpa is plotted in solid purple.

5.3 Crustal Thickness Variations

The Moho depths we observe in the center of the Ethiopian Plateau are qualitatively consistent with crustal thickening following the impingement of the Afar Plume ~ 30 Ma (16; 15). This event is thought to have thickened the crust in two ways: up 2 km of flood basalt emplacement at the surface (61; 62; 63), accompanied by up to 10 km of underplating, yielding a total crustal thickness up to ~ 45 km. This is best observed

from wide-angle seismic reflection/refraction data (16), but that study only imaged the deepest crust ~ 100 km on either side of the rift valley. Some degree of thickening by igneous intrusion is also well recorded by gravity surveys (19), and increased V_p/V_s ratios signifying a more mafic crustal composition (21). Xenolith data collected along the YTVL also support underplated material beneath the near-rift EP (67; 133; 132), and melt emplacement well outside the rift axis is supported by abundant basaltic dikes (45; 50) west of Lake Tana, where magmatism exploited zones of pre-existing weakness.

We do not observe crust as thick as 45 km in our field region. The average crustal thickness of stations in our inversion within the EP is 34 km, and the thickest crust we see is 39 km at SERE, just north of Lake Tana. Ignoring erosion, 12 km of thickening would suggest a pre-plume thickness of 22 km in the plateau, which we argue is unrealistically thin for continental crust. In order to estimate an appropriate baseline, we propose that stations that lie outside the bounds of flood basalt emplacement (fig. 2.1b) in the NW of our field region best exemplify pre-rift/plume lithospheric structure and crustal thickness: ~ 30 - 32 km (fig. 4.2). One caveat is that four of six of these stations lie within the erosional basin of the Blue Nile Gorge, which has incised a mean of 0.37 km of rock within the basin since 29 Ma (134). As another regional comparison, ~ 35 km crust is observed further west in Sudan (135; 136), which similarly lacks flood basalts or underplating.

With this baseline of ~ 31 km, the crustal thicknesses of 31-39 km we see in the EP to Lake Tana imply igneous addition on the order of 0 - 8 km, distributed non-uniformly across the region. This disparity in estimated underplating compared to previous studies may simply be geographical - there was more igneous thickening in the region of the MER than on the flank. Alternatively, it might be methodological; our receiver functions might be sensing the substantial velocity boundary between pre-existing crust and igneous underplate, and not the (perhaps more gradational) underplate-mantle bound-

ary (e.g., *Stuart et al., 2006*) to which wide-angle reflection/refraction is more sensitive. Nonetheless, since our method attempts to fit the full receiver function waveform, and since we do not see an abundance of double-peaked *Pxs* phases, we estimate that the crustal thickness patterns we see are real.

Two observations stand out in the EP slow velocity lineament: Firstly the thickest crust in our field region is around Lake Tana and coincides with some of the lowest off-rift shear/phase velocities. Secondly, we see a modest trend of thinning to the southwest (fig. 4.2a). This tapering has several possible explanations. It may be that more erosion has occurred in the south-central plateau than previously thought, or alternatively that the plume emplaced more igneous material to the north, creating gradients in flood basalt thickness. Finally, it may be that the pre-plume, pre-extension thickness increased from north to south, and we are imaging a step in the flood basalt province; N-S trending sutures and dike swarms are more concentrated near Lake Tana (fig. 2.1b) perhaps connoting more extensional strain in that area in the geologic past that has altered crustal thickness. This region is further discussed in Section 5.4.

Although we only include three stations from within the rift axis in this study, they show evidence for thin crust within the SMER (<31 km), and thicker crust (~38 km) within the CMER, consistent with previous work (6; 15; 3). Although simple models of pure shear rifting would predict the thinnest crust in the region within the well defined rift valley (137), thicker crust in the CMER supports complex, diachronous rift evolution of the MER (72; 74; 5).

5.4 Off - Axis Extension

The northern portion of the slow velocity lineament in the EP coincides with Quaternary volcanoes, spatter cones, fumaroles, and thermal springs (fig. 2.1b) (67; 30; 27).

These features are concentrated near Lake Tana, where we find the slowest velocities, and where calculations (Section 5.2) seem to require 0.1 - 0.46% partial melt in the upper mantle to fit group 2 V_s observations. This region is also observed to have regionally pervasive slow P and S-wave arrivals (1; 4; 27; 60) and small amounts ($<1\text{mm/yr}$) of contemporary surface extension as determined by GPS measurements (23). We find V_p/V_s ratios at stations closest to Lake Tana among the highest in our region, in excess of 1.8 (fig. A.12), indicative of melt (18).

Surficial volcanic features suggest that the northern portion of the slow velocity lineament is more mature than the southern, which is also supported by a larger magnitude GPS velocity measurements near Lake Tana (fig. 2.1a). Geologic evidence suggests that Lake Tana was previously a particular locus for extension. Pre-existing N-S trending sutures in this area (38; 138) are believed to have guided plume-fed dike swarms (45; 50) and may remain active zones of weakness. Multi-phase graben faulting (49, and references therein), strike-slip motion within the God Serpent dike swarm, shear structures within their margins (43), and magnetic imbrications (139; 140) all indicate that Lake Tana has experienced localized extension episodically since the Oligocene. Ti systematics indicate that the region around Lake Tana may have been initial, or an important secondary, locus of plume-sourced melt infiltration 30 Ma (49; 50).

Without continued extension, the magmatic activity and - as we have shown - low velocities southwest of Lake Tana would have ceased since the plume arrival. Our results, together with a host of previous work, suggests ongoing localized, shallow extension >450 km from the rift axis along a rift-parallel lineament extending south west from Lake Tana. Particularly reduced shear velocities, and volcanism, near Lake Tana suggest that the northern portion of the slow velocity lineament is more mature than the south. That said, our imaging of velocity perturbation at depth extending south of the $\sim 10.6^\circ\text{N}$ southernmost limit of volcanism indicates that deeper lithospheric perturbation can lead

surficial extension. This extension may be driven from the bottom-up.

P- and S-wave delay time observations and tomography of the Ethiopian plateau have previously hinted at a slow upper mantle extending well beyond the MER valley (1; 4; 27; 11). Absolute P-wave delay times of up to 1 second imply δV_p of as more than -2% off axis, with these low velocities asymmetrically observed to the north west of the rift (60). These results agree qualitatively with our data, which reveal upper mantle velocities 4.3% (0.19 km/s) slower than PREM in the group 2 region. Clearly, there is substantial thermal perturbation and -our calculations imply- *in situ* melt beneath large swaths of the Ethiopian plateau, with particular apparent concentration in the slow velocity lineament.

Geodetic data suggest that while the majority of extension in the northern EARS is localized within the MER and Afar, up to $\sim 20\%$, or 1 - 2 mm/yr, of extension may be distributed across the EP (23). (6) have interpreted crustal thickness maps as implying broad distribution of strain beyond the MER. Small magnitude GPS measurements in the EP of < 1 mm/yr are of similar magnitude to those in the near-MER EP (fig. 2.1a) (23), suggesting extensional accommodation is present in both locales. Such a combination of diffuse and narrow extension has also been proposed to occur in the southern EARS (141).

Finally, it is interesting that notably thick crust near Lake Tana (~ 36 - 39 km), and at other Group 2 stations, coincides with some of the slowest upper mantle velocities where we have argued extension is ongoing (Figure 4.2). At face value, this is paradoxical. We propose two potential explanations in the context of off-axis extension. One possibility is that > 10 km of magmatically driven thickening (16) associated with plume impingement has taken place, following which high ($> 40\%$) strain pure shear extension in the plateau - evidenced by slow shear velocities - has re-thinned the crust by up to 9 km to present day values (~ 30 - 39 km). Alternatively, extension off axis has reduced velocities due

to lithospheric degradation (142) and melt percolation, but is still too immature to have initiated much crustal thinning. In this case, our results would seem to be inconsistent with >5km of magmatic thickening. It may also be that underplating of the Ethiopian plateau tapers substantially away from the - now rifted - center. The more parsimonious explanation would seem to be the latter; melt and thermal effects of bottom-driven incipient rifting well precede crustal breakup, at least within the Ethiopian rift flank.

Chapter 6

Conclusions

We have investigated the northwestern EP flank of the East African Rift and Ethiopian Plateau using Rayleigh wave phase velocity measurements and P s receiver functions to interrogate the deep crustal and shallow mantle structure. We jointly inverted these data for shear velocity profiles using a trans-dimensional, hierarchical Bayesian inversion algorithm, incorporating H- κ receiver function stacks to better constrain crustal thickness.

Throughout the study region we found slow phase and shear velocities, reflecting mantle temperatures significantly elevated compared to a stable continent. We showed that high temperatures and *in situ* mantle melt content extend beyond the rift. The central Ethiopian plateau has a fast lid above a gradual negative velocity gradient. Within and near the SMER, we found maximally thinned crust and extremely low upper mantle velocities - the effects of rifting are clear. Similarly thin crust within the NWEF flank, on the other hand, is paired with fast upper mantle velocities. These flank stations appear to be the best local representatives of pre-rift and pre-plume lithosphere, comparatively unaffected by either process.

In the NWEF, we observed a lineament sub-parallel to the rift axis with slow shallow upper mantle velocities. This lineament coincides with GPS-observed divergence,

pervasively delayed teleseismic arrivals, and Quaternary aged volcanism, fumaroles, and thermal springs near Lake Tana. These observations support the presence of localized off-axis extension, hundreds of kilometers from the MER. Plume-derived underplating and flood basalts cause ambiguity in determining the extensional maturity of the slow velocity lineament. If igneous addition has offset crustal thinning this region may already be heavily extended. Alternatively, there may be insufficient off-axis extension to have allowed much crustal thinning at all, but melt percolation and lithospheric degradation have already begun to substantially reduce shear velocity. In either case, it is apparent that some amount of that off-rift extension occurs well away from the MER, and that continental breakup in Ethiopia is more complex than simple narrow rifting.

Data Availability Statement All data used are available from the Iris Data Management Center (<https://service.iris.edu/>). IRIS Data Services are funded through the Seismological Facilities for the Advancement of Geoscience (SAGE) Award of the National Science Foundation under Cooperative Support Agreement EAR-1851048 and EAR-1261681. All models and derived datasets associated with this manuscript are available in Dryad repository ([doi:10.25349/D9RK6H](https://doi.org/10.25349/D9RK6H)). Networks used to develop Rayleigh wave phase velocity maps and receiver functions, in alphabetical order, are: AFAR0911 2H (https://doi.org/10.7914/SN/2H_2009), NABRO 4H (https://doi.org/10.7914/SN/4H_2011), ESP 5H (https://doi.org/10.7914/SN/5H_2011), AfricaArray AF (<https://doi.org/10.7914/SN/AF>), GEOSCOPE G (<https://doi.org/10.18715/GEOSCOPE>).

G), GSN IU (<https://doi.org/10.7914/SN/IU>), XI (https://doi.org/10.7914/SN/XI_2000), EAGLE II XJ (https://www.fdsn.org/networks/detail/XJ_2002/), ARGOS XM (https://doi.org/10.7914/SN/XM_2012), YOCMAL XW (https://doi.org/10.7914/SN/XW_2009), Riftvolc Y6 (https://doi.org/10.7914/SN/Y6_2016), EAGLE YJ (https://www.fdsn.org/networks/detail/YJ_2001), Ethiopia YY (https://doi.org/10.7914/SN/YY_2013), Boina YZ (https://www.fdsn.org/networks/detail/YZ_2005/), OldAFAR ZE (https://doi.org/10.7914/SN/ZE_2007), AFAR ZF (http://www.fdsn.org/networks/detail/ZF_2007/), ADDSA ZK (https://doi.org/10.7914/SN/ZK_2009).

Appendix A

Manuscript Appendices

1. Figure & Text A.2, Earthquake Distribution
2. Figure & Text A.3, Checkerboard Test - 100s
3. Figure & Text A.4, Semblance Tests
4. Figure, Text & Table A.5, Ray Paths & Hitmap Density
5. Figure & Text A.6, Standard Deviation Plots
6. Figure & Text A.7, DBSCAN Parameter Selection
7. Figure & Text A.8, H- κ Reliability Test
8. Figure & Text A.9, Receiver Function Derived Crustal Thickness and V_p/V_s Ratios
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11. Figure & Text A.12, Sensitivity Kernels
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13. Figure & Text A.14, Geotherm Parameter Gridsearch

A.1 Introduction

Below we present several figures to support observations and conclusions we develop in the main document. All data are developed from the same two data sets presented in the main document, and figures presented show interim steps that may clarify decisions made during interpretation, or error analysis and testing.

Figure A.1 shows the epicenter locations and azimuthal distribution of events used to develop phase velocity maps. Figure A.2 shows the 100 s checkerboard test, which was cut from the manuscript for figure clarity. Figure A.3 shows the semblance between the input and output checkerboard models. Figures A.4 and A.5 show the inter-station ray paths used in developing our phase velocity maps, and table A.5 reports the number of raypaths at each period. Figure A.6 shows hitmaps in terms of ray density per-pixel. Figure A.7 reports the standard deviation of phase velocity measurements at each period. Figure A.8 shows the semblance, hitmap, and standard deviation at 100 s. Figures A.9 and A.10 show how parameterization choices for minpts and ϵ affect the result of DBSCAN. Figure A.11 shows variability of the H- κ surface by varying the input parameters, before and after the DBSCAN cleaning process. Figure A.12 shows the crustal thickness and Vp/Vs ratio found by H- κ stacking. Figure A.13 shows the empirical prior developed to display lack of bias within the THBI algorithm. Figure A.14, A.15, and A.16 show data fit during inversion, the fit of crustal thickness and Vp/Vs found via inversion, and the final shear velocity model with Moho depth distribution, respectively. Figure A.17 displays a sensitivity kernel developed using the median shear velocity model. Figures A.18, A.19, and A.13 show the inherent dissimilarity between forward modeled shear velocities from a continental geotherm and observed shear velocities. Figures A.21, A.22,

and A.23 show gridsearch surfaces used to determine the best fit geotherm to each shear velocity aggregate group, and table A.14 shows the parameters used to develop these geotherms. Zipped files contain figures A.14, A.15, A.16, and A.17 for each inverted station.

A.2 Earthquake Distribution

Epicenter and azimuthal distribution of the 5741 earthquakes used to develop our phase velocity maps (fig. A.1) show that we have good azimuthal distribution of events, which supports that our phase velocity maps are not azimuthally biased.

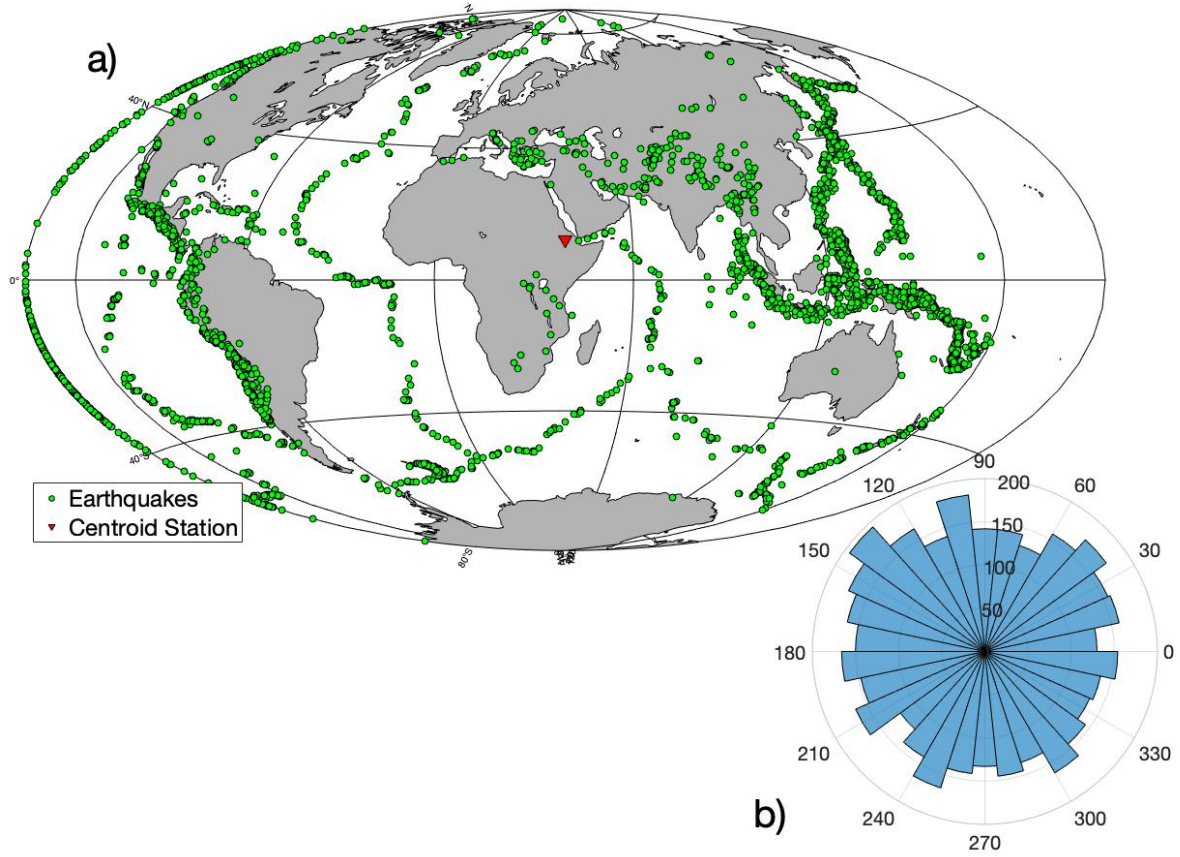


Figure A.1: The epicenter distribution of earthquakes (green dots) used in developing phase velocity maps in relation to the centroid of our research area (red inverted triangle). b) The same events, plotted in 30 ° azimuthal bins, to elucidate azimuthal distribution.

A.3 Checkerboard Test - 100s

We provide the checkerboard result for 100s (fig. A.2), as it was not included in the main manuscript. The process used to develop this checkerboard is the same as that defined in the methodology.

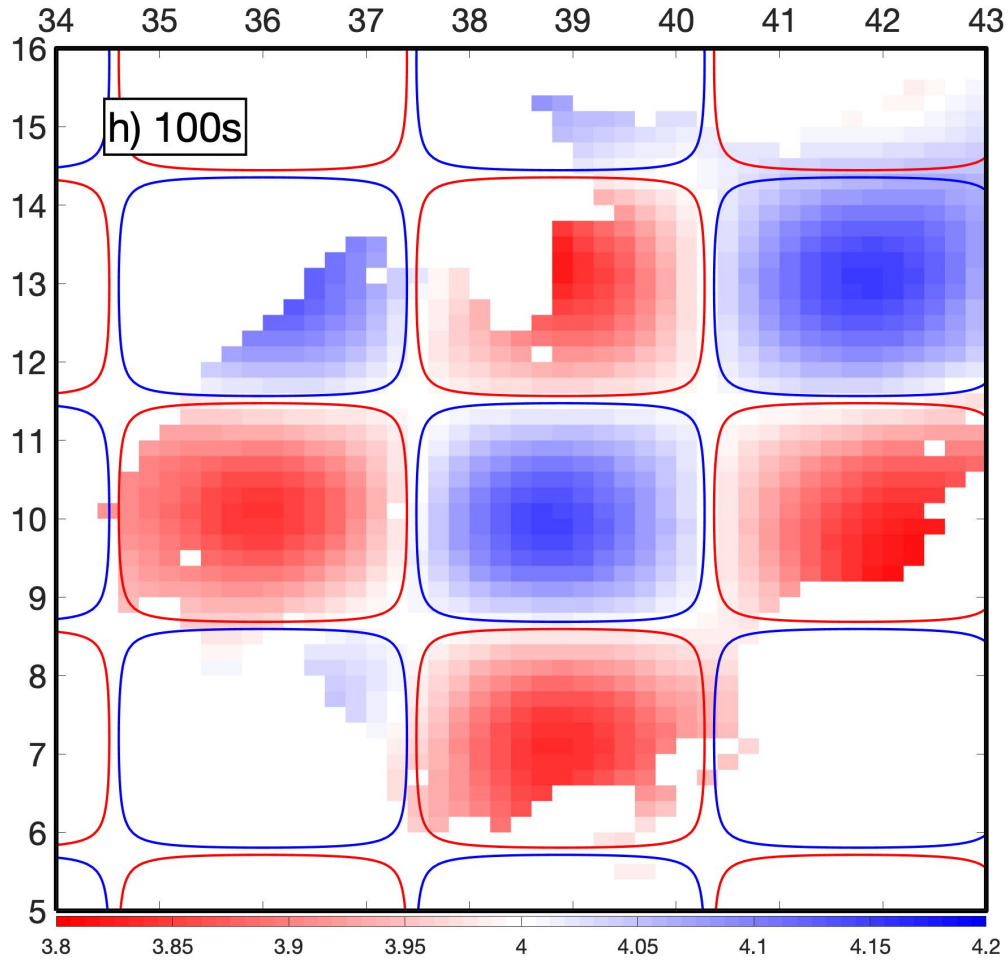


Figure A.2: The checkerboard result at 100 s, with velocity contours displayed in red (3.8 km/s) and blue (4.2 km/s). Checker size is $\sim 3.6 \times 3.6$ km.

A.4 Semblance Tests

Using checkerboard tests, we calculate the semblance between the resolved checkerboard output and the ideal checkerboard (87), with the radius equal to $2 \times$ the checker size, in km (fig A.3). A caveat to this scaling is that shorter periods will inherently have

lower semblance because the resolution requirements are higher than those at longer periods. We find that semblance is high enough to support our observations are robust, particularly in the NWEP and CMER.

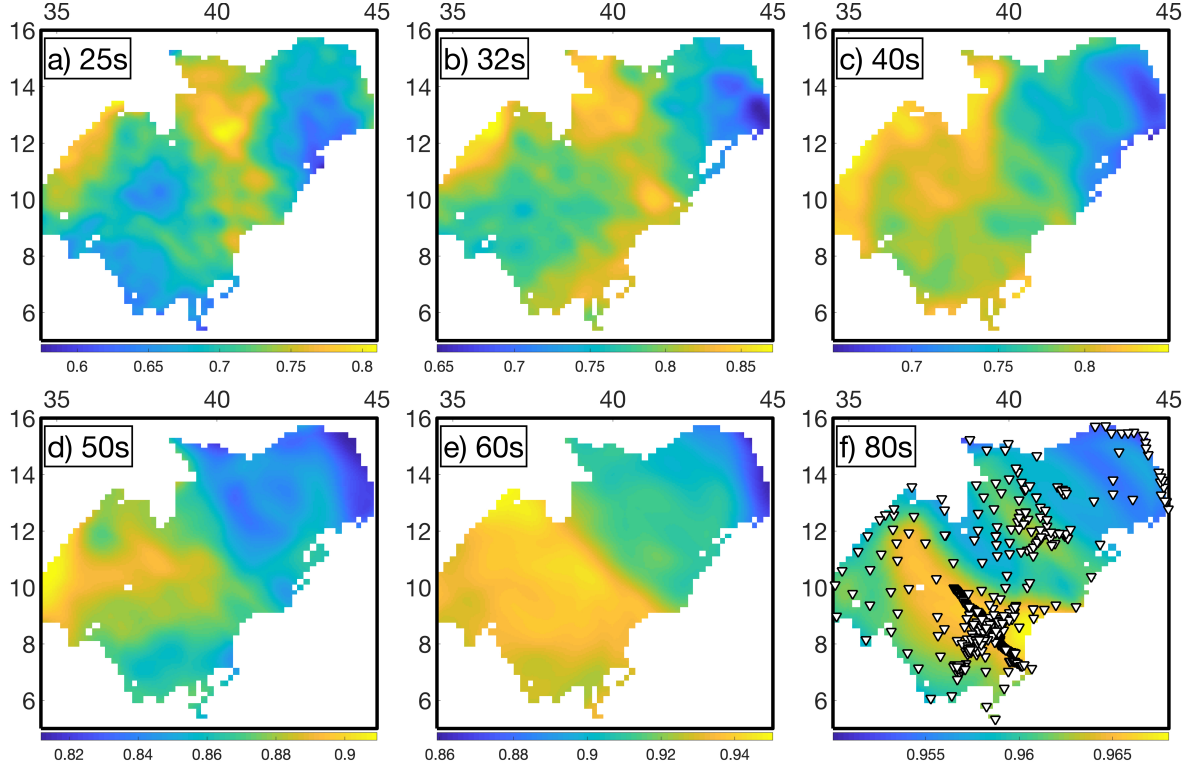


Figure A.3: Semblance plots for 25 - 80s, as a companion to fig. 2 in the manuscript. Semblance radius was set as $2\times$ the checker size in fig. 2, at each period.

A.5 Ray Paths & Hitmap Density

In order to demonstrate reliability of our phase velocity maps, we show the raypaths between station pairs that were used in the final phase velocity stack (fig. A.4, A.5). Only station pairs that passed quality controls laid out in methods section 3.1 are plotted at each period. We believe the raypaths display good coverage throughout the region,

which supports our phase velocity results. We display only two examples of the raypath maps (the highest and lowest number of raypaths), but the the number of raypaths corresponding to each period is shown below. A list of raypaths per periods is available in table A.5. Hitmaps are also developed (fig. A.6) where each pixels within the map at each period denotes the number of intersecting rays passing through. Densities are the highest within the CMER, SMER, NMER, Afar, and the NWEF. We note that the regions of higher density are the same between all periods.

Raypath Plots

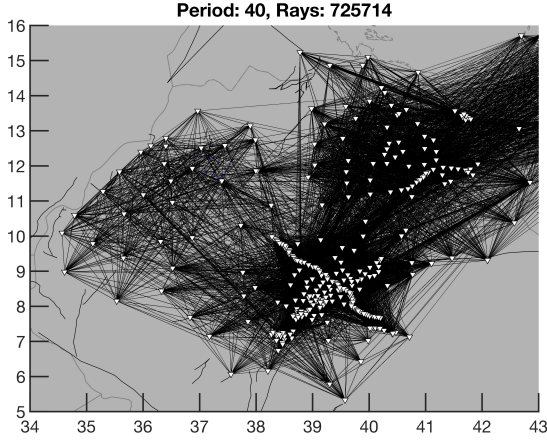


Figure A.4: Interstation raypaths at 40s. Black lines indicate an inter-station pair that was used in phase velocity map creation, and white inverted triangles denote station locations.

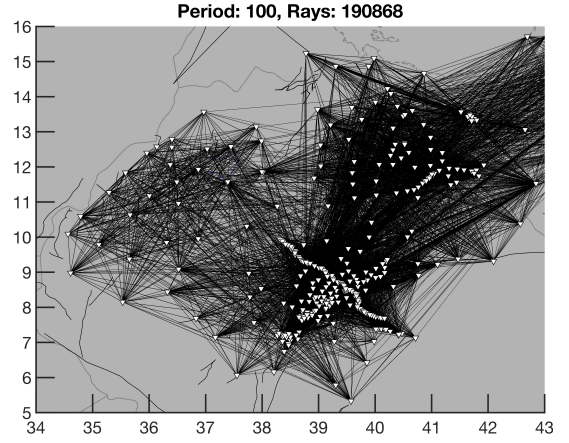


Figure A.5: The same as A.4, but showing interstation raypaths at 100s.

T2 Raypaths per Period

Period	Raypaths
25	475,110
32	345,300
40	725,714
50	453,983
60	481,227
80	472,921
100	190,868

Hitmaps

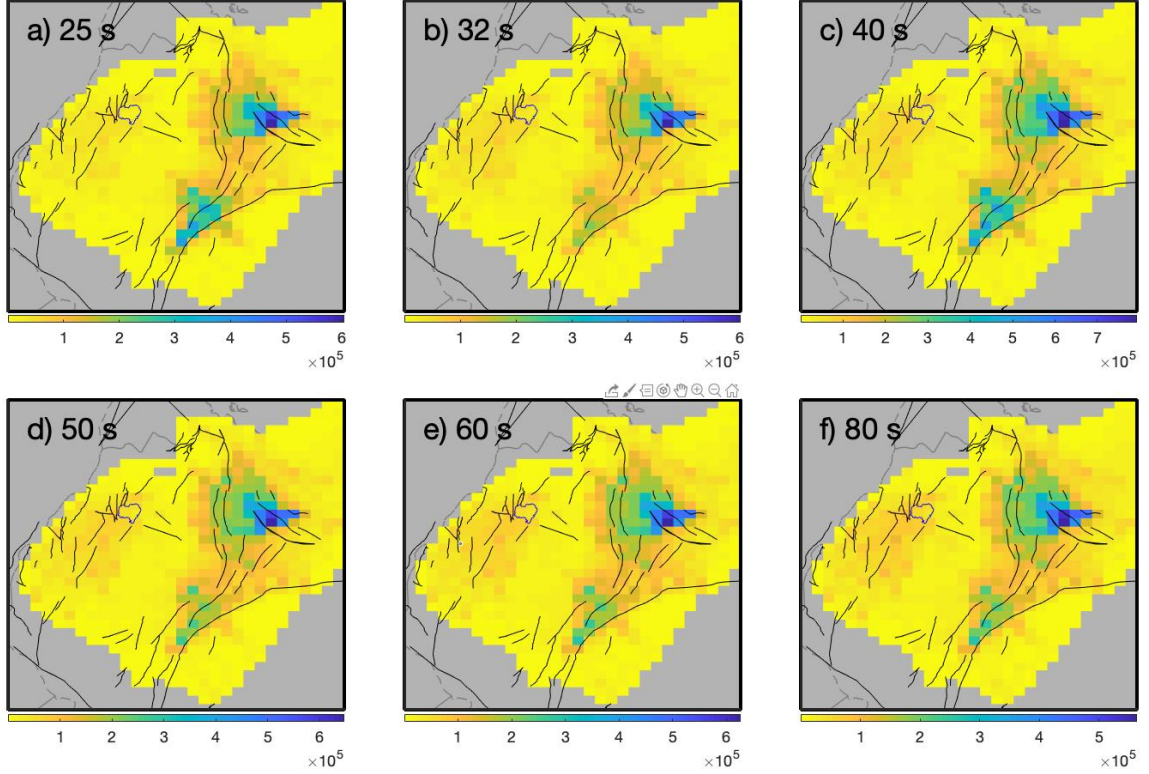


Figure A.6: Event hitmaps from 25 - 80s, in the same fashion as the checkerboard and semblance plots. Pixel color denotes the number of raypaths that passed through it. Pixel size is static at each period, and is $0.3^\circ \times 0.3^\circ$. The outline of Lake Tana is plotted in blue, and surficial faults are plotted in black for orientation.

A.6 Standard Deviation Plots

In addition to the above tests, we also report that standard deviation of our phase velocity maps at each period (fig. A.7). Regions of low standard deviation are most apparent within the NWEF and Afar. The NWEF has < 0.1 km/s at most periods, which support that our observations are robust. A caveat to these standard deviation measurements is that the THBI algorithm inverts for the standard deviation hyperparameter for each data type, for each model, and will fit phase velocities depending on

this hyperparameter. Therefore, THBI is not intrinsically sensitive to initial standard deviation measurements, and can invert high- and low- std data equally well. Furthermore, reducing the number of events that are included in the final stack does not strongly change observed phase velocities such that stacking phase velocities within three standard deviations, and phase velocities within one standard deviation show a 0.85% change in shear velocity, but a reduction in standard deviation up to 0.07 km/s.

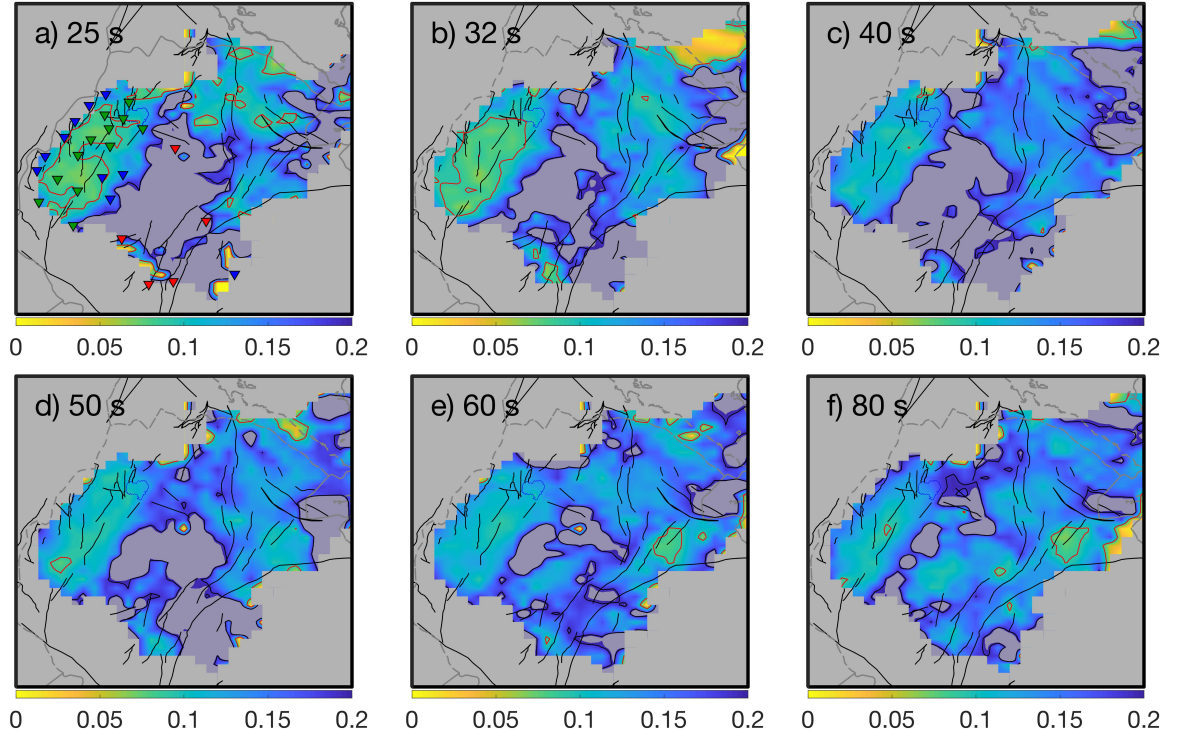


Figure A.7: Standard deviation maps from 25 - 80s. Red contours indicate regions of < 0.1 km/s std, and black contours indicate < 0.15 km/s std. Regions greater than 0.2 km/s are plotted with high transparency. Lake Tana is plotted in blue, and surficial faults are plotted in thin black lines.

100s Semblance, Hitmap, and Standard Deviation Plots As per the above tests, we also include 100s, which was left out of the previous figures for figure clarity.

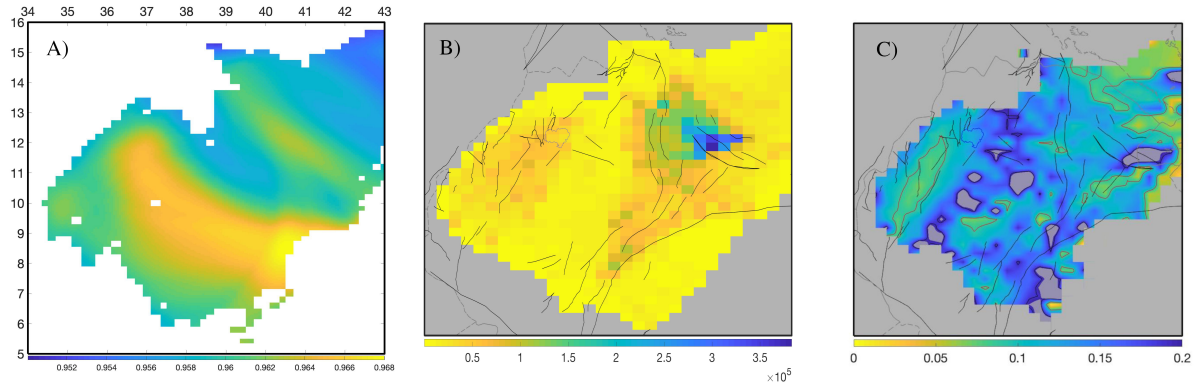


Figure A.8: a) The semblance plot at 100s, as in figure A.3. b) The hitmap at 100s, as in figure A.6. c) The standard deviation map, as in figure A.7.

A.7 DBSCAN Parameter Selection

Parameterization will change the outcome of DBSCAN, despite path-specificity. Therefore, determining robust parameterization for min points and ϵ was necessary. Min points was chosen as a static value of 30% of the total number of RFs for each cull step. Fig. A.9 reveals that for station GIDA, requiring between 10 - 75% of the total RFs to be within the cluster has little effect on the total number of RFs in the cluster. Varying minpts is not particularly useful for discriminating good traces between 20 - 50%, while varying 50 - 75% begins to exclude traces, and requiring a minimum number of RFs that encompasses 75% of the population of traces or higher drops acceptance of a cluster to zero.

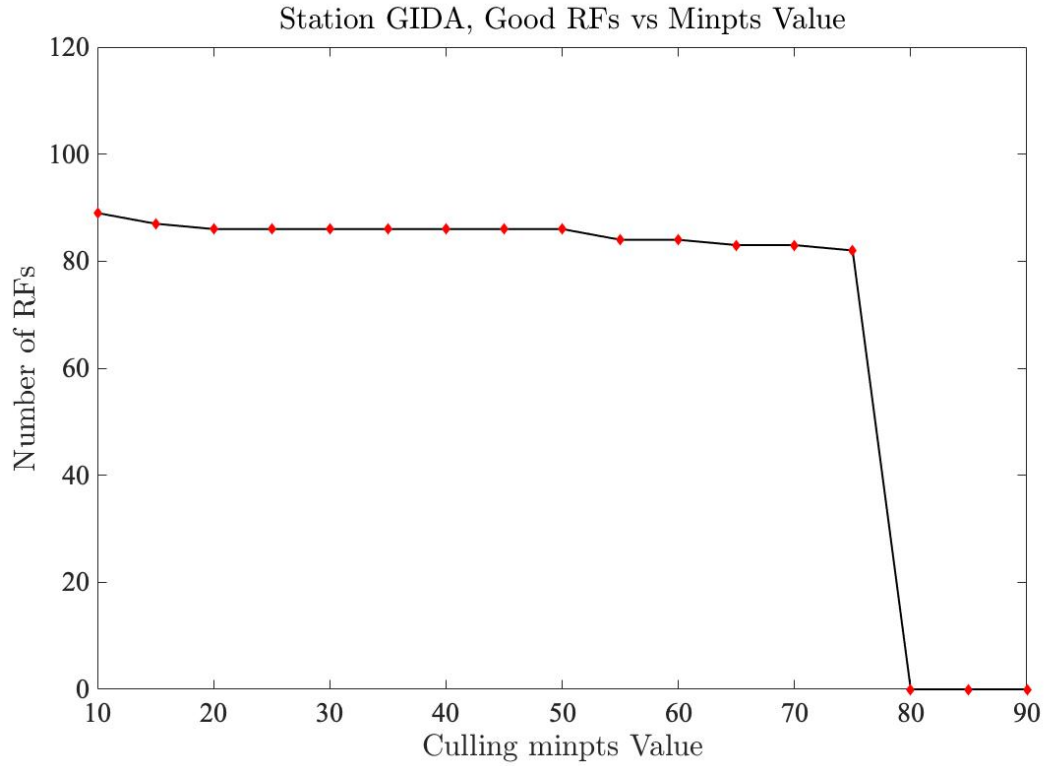


Figure A.9: Red diamonds signify the discretization for minpts percentage. ϵ was held static, with ϵ_1 set at 20, and ϵ_2 set as 28, the value used during DBSCAN processing for this station.

Choice of ϵ had a much greater effect on the result of DBSCAN. In order to maximize the number of traces, we increased ϵ until the number of RFs in the cluster plateaued. We believe this plateau represents the best point of tradeoff between the number of RFs, and the similarity of the RFs within the cluster. It should be noted, however, that even with a liberal ϵ a population of noise is still differentiated.

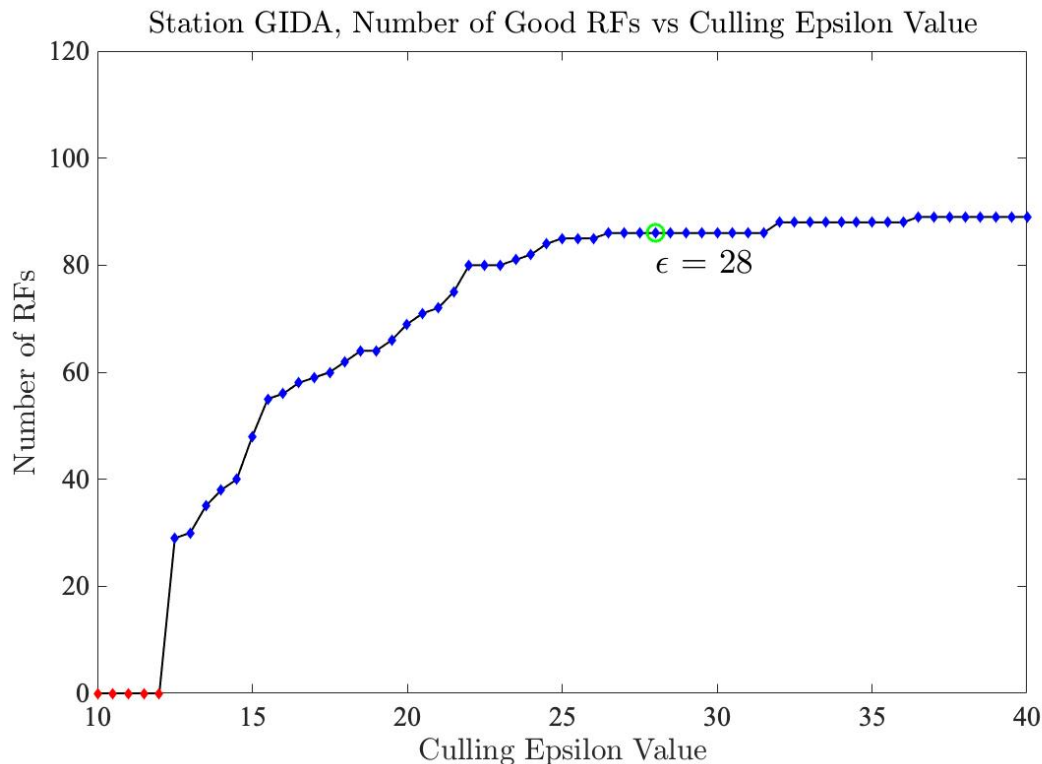


Figure A.10: Red diamonds signify that no cluster was found, and blue diamonds signify that a cluster of good RFs was distinguished. The number of traces incorporated by an ϵ value increases rapidly, and then plateaus. The plateau is the point at which an epsilon was defined as stable. Notably, the number of RFs between $\epsilon_1 = 24$ and $\epsilon_2 = 28$ is 3 (86 vs. 89 RFs). The chosen ϵ for station GIDA was 28, which is circled and labeled. The total number of traces passed into this culling step was 116.

A.8 H- κ Reliability Test

(21) found that many H- κ derived estimates were unreliable within the EPL. Using an automatic method to perturb input parameters for H- κ stacking, they found that stations within flood basalts produced unreliable H- κ stacks with high standard deviation, low SNR, and broad low amplitude arrival phases. Such H- κ stacks failed to resolve a high energy maximum for the Moho and κ due to a gradational Moho, complex crust, or

sediments and basalts.

To test the reliability our H- κ stack results, we modified (21)’s random input perturbation procedure on several stations in the region using receiver functions from before the DBSCAN process (raw) and those that passed the DBSCAN process (cleaned). We randomly perturbed the Vp from [6.2: 0.1: 6.8], the maximum filter frequency from [0.1: 0.1: 2.0] Hz before calculating the receiver function, the weight of the Ps phase from [0.4: 0.05: 0.9], the $PpPs$ phase from [0.1: 0.05: 0.6], and the $PpSs + PsPs$ phase from [0.0: 0.05: 0.5] (with total phase weights required to sum to 1.0). These random perturbations were run 5000 times for the raw and cleaned receiver functions. A robust result was expected to display a distinct high energy maximum in the H- κ surface, closely clustered mean model solutions between different input parameters, a similar mean and mode H- κ for all 5000 models, and an H and κ that were not capped at the edges of the grid search surface. An example of this process is shown in fig. A.11. Raw H- κ stack surfaces (fig. A.11a) generally fail to find a high energy maximum - the summation of all H- κ surfaces finds five high energy maxima in which input parameter variability solved to different maxima despite using the same data. The mean H- κ value of the 5000 iterations (40.81, 1.81) is dissimilar to the absolute maximum of the surface (39.9, 1.6). The surface maximum is also capped by the bounds of the κ values. Contrarily, the cleaned H- κ stack surface (fig. A.11b) finds a well defined maximum in the H- κ stack surface (37.4, 1.68), which is similar to the mean (37.52, 1.67) of the 5000 best fit H- κ iterations. We note that the difference between raw and cleaned H- κ stack surfaces are not consistently different across all stations, but quality controls still improved the H- κ energy surface.

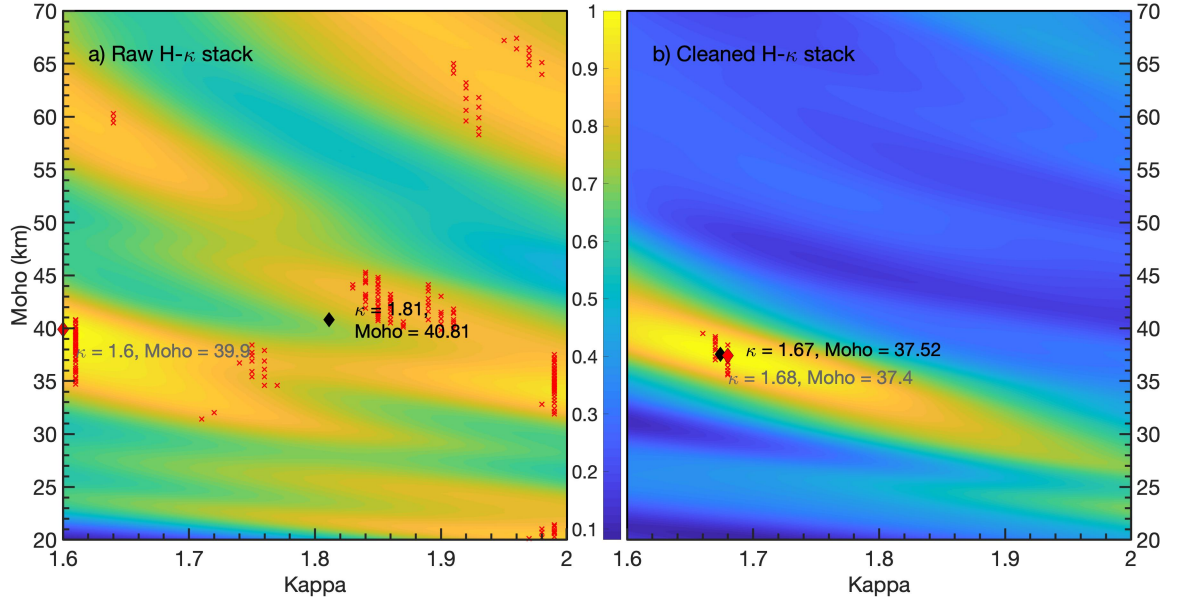


Figure A.11: The results of the cleaned vs raw H- κ stacking procedure. a) shows the results of 5000 H- κ model iterations with randomly perturbed parameters. The red crosses are the best fit H- κ value for a single model iteration. The black diamond is the mean of those models, with corresponding κ and Moho values labelled in black. The red diamond (at the left edge of the surface) is the maximum of the error surface, with the corresponding κ and Moho values labelled in grey. b) shows the results of 5000 H- κ model iterations with randomly perturbed parameters after the DBSCAN process. The black diamond is the mean of those models, with the corresponding κ and Moho values labelled in black. The red diamond is the maximum of the error surface, with the corresponding κ and Moho values labelled in grey. The summed H- κ surfaces are shown in the background for both the raw and cleaned stacks, color coded by the same normalized amplitude scale.

A.9 Receiver Function Derived Crustal Thickness and V_p/V_s Ratios

Crustal thickness and V_p/V_s ratios are taken as the maximum of the normalized H- κ surface at each station. A V_p of 6.3 is used, and phase weights of 0.7, 0.2, 0.1 are used in the H- κ stacking process. Crustal thickness results are similar to those of the inverted

results, and trends discussed in the inverted crustal thickness section are relevant here as well. Our V_p/V_s ratios are similar to those of (18) in the western Ethiopian Plateau (1.7 - 1.9). There is broader uncertainty with the V_p/V_s ratio than the crustal thickness, even for well constrained H- κ stack surfaces, as can be seen below in the inversion data fit subsection. Crustal thickness at each station, and the standard deviation are reported in T2.

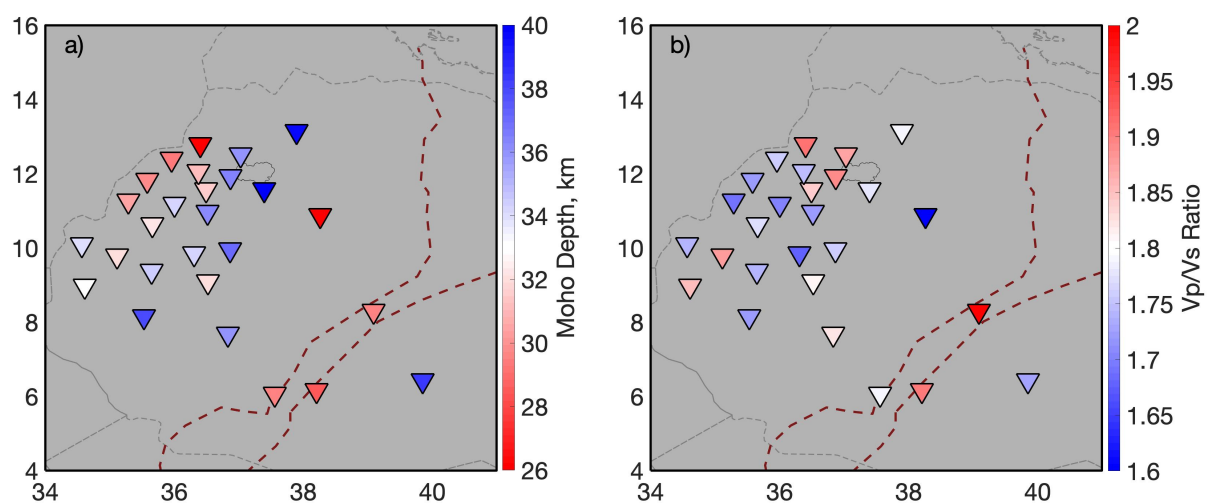


Figure A.12: a) The crustal thicknesses for each station taken as the maximum energy region in the H- κ surface. Warmer colors indicate thinner crust. b) The V_p/V_s ratio from the same H- κ surface, at each station. Warmer colors indicate a higher V_p/V_s ratio.

A.10 Inversion Prior

In order to show that our joint inversion approach contains no inherent bias, we develop a six chain, 21,000 iteration empirical prior. A prior is the result of running the joint inversion with the likelihood set to 1, such that all models are accepted, with proposed models conforming to the *a priori* bounds imposed during joint inversion with data. Distributions show that the algorithm itself does not impose a strong influence on

the posterior (33). So long as the same model parameters and data types are used, the same posterior model may be compared to many inverted shear velocity structures (fig. A.13).

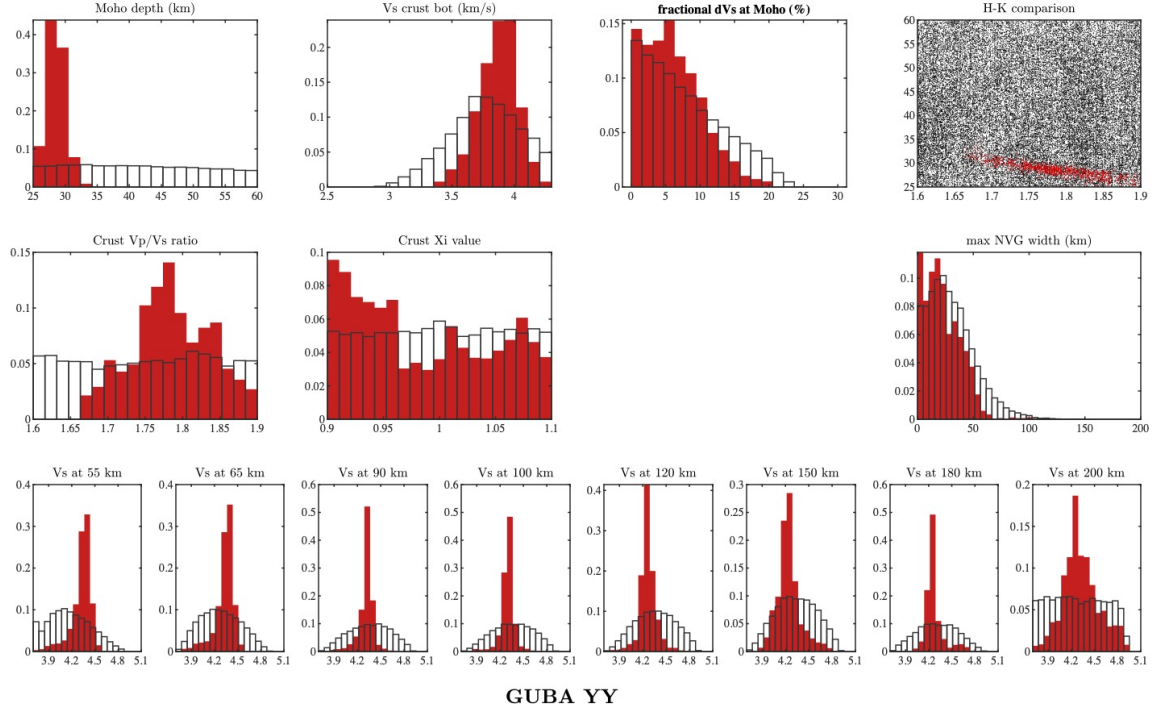


Figure A.13: The prior developed for station GUBA (hollow black bars, dots), compared to the posterior (solid red bars, dots). Distributions for various parameters show the inherent bias, if any, in the THBI algorithm. Note that not every distribution is uniform, which is dictated by how the algorithm parameterizes velocity at depth using cubic splines. The effects of constraints on candidate models (for instance, requiring a positive velocity jump at the moho) filter thorough to non-uniform empirical prior distributions for fractional δV s at the Moho and the maximum negative velocity gradient width (33).

A.11 Inversion Data Fit, H- κ stacks,, and Median Velocity Model

For each station we jointly inverted, we have included three images similar to figure 4 in the manuscript. The naming scheme is as follows, where STATION NAME refers to the inverted station. `sf#_STATION NAME_data_pred.jpg` : The fit of the mean surface wave, receiver function, and H- κ model. `sf#_STATION NAME_HK_fit.jpg` : The same as the previous image, but with every H- κ solution from shear velocity models plotted in the gridsearch area. `sf#_STATION NAME_vs_prof.jpg` : The inverted shear velocity profile. The red line indicates the mean model, the grey patch is the 1σ , and white patch bounded in light grey is the 2σ . The posterior distribution of Moho solutions is plotted as a red histogram, and the best fit Moho depth and standard deviation are denoted by the horizontal dashed black line. An example of the three images is shown below for clarification.

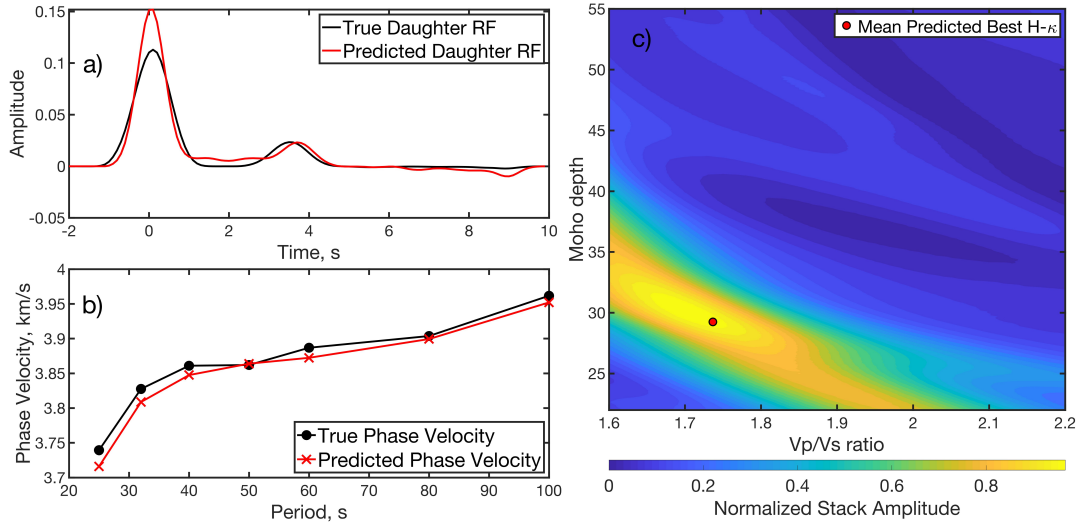


Figure A.14: a) The predicted and observed receiver function data for station ABMD. The observed receiver function is shown in black, the mean of the ensemble of well-fit models is shown in red. b) The observed surface waves (black) and predicted (red) surface waves. c) The H- κ surface after normalization. Warmer colors indicate higher amplitudes. The H- κ result of the mean model is plotted in red.

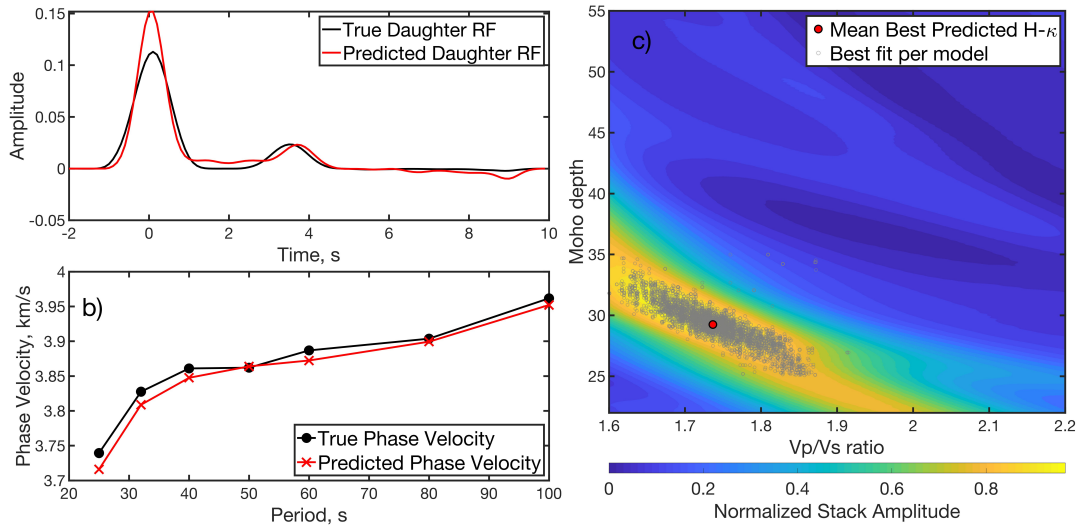


Figure A.15: The same figure as fig. A.14, but c) now contains the H- κ of all fit models developed during inversion. Light grey circles denote a single model.

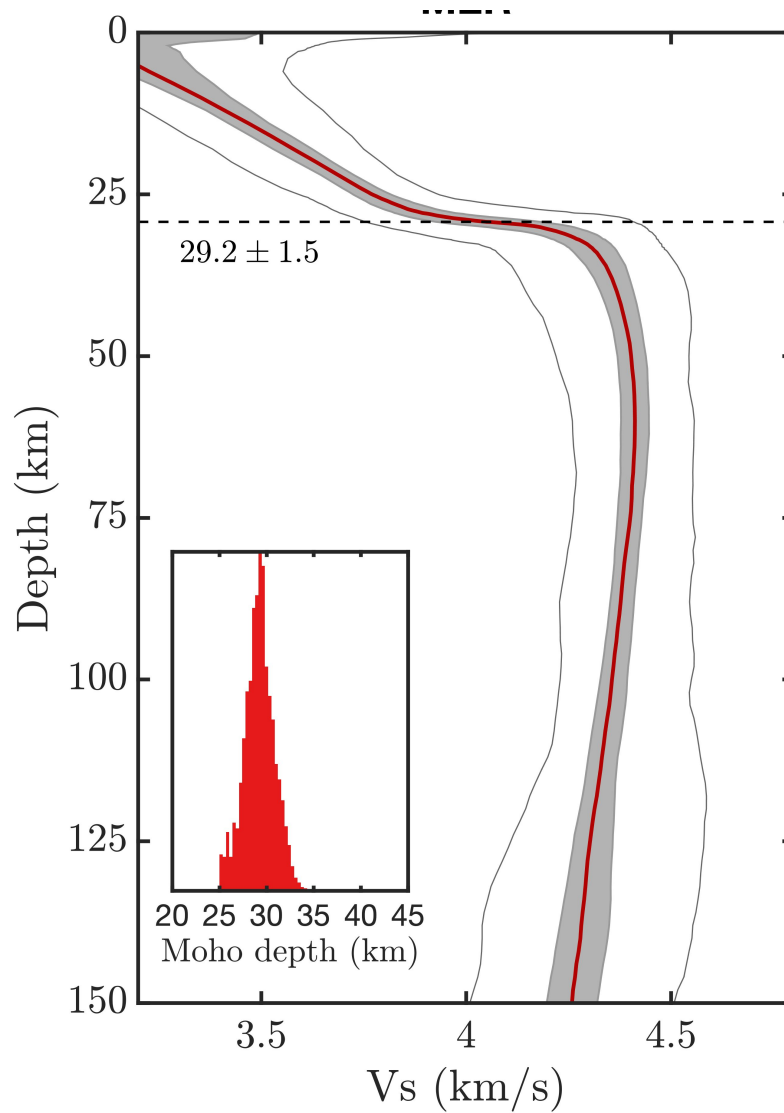


Figure A.16: The shear velocity model of station ABMD. The mean velocity model is plotted in red, the 1σ patched in grey, and the 2σ patched in white bounded by light grey. The histogram of Moho depths is shown in the bottom left. The maximum of the Moho distribution and distribution standard deviation are labelled and displayed as a dashed horizontal black line.

A.12 Sensitivity Kernels

In order to demonstrate our depth sensitivity, we developed sensitivity kernels during inversion (fig. A.17). In a similar fashion of finding the mean shear velocity model from

posterior distributions, we similarly find the 50th percentile for S and P anisotropy of well fit models, and forward model the sensitivity kernel using MINEOS (105). The resultant sensitivity kernel demonstrates our sensitivity to vertical and horizontal V_s combined. While only one sensitivity kernel is provided in this supplement, we have provided sensitivity kernels for each inverted station in the zipped supplement, named as: sf#_STATION NAME. Sensitivity kernels are similar at all stations, with peak V_s sensitivity varying from ~ 25 - 125 km, depending on period, but resolvable to below 200 km. Because 165 km is the greatest depth we discuss shear velocity observations, we believe our results and conclusion are valid based on our sensitivity kernels.

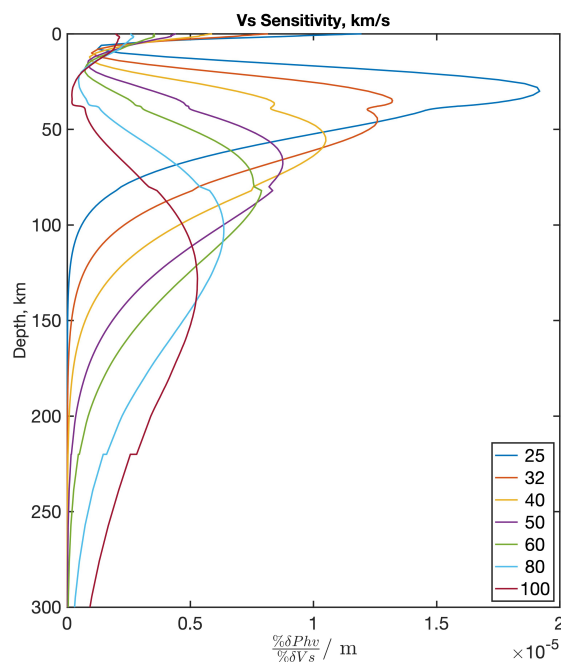


Figure A.17: Sensitivity kernels developed using the mean velocity model at station AMME. V_{sv} and V_{sh} are combined to show the V_s sensitivity in km/s at depth. Periods are indicated by color. The x-axis is in units of 1/m, and represents a percent change in phase velocity per percent change in V_s , per meter.

A.13 Geotherm Gridsearch Considerations

It is difficult to forward model shear velocities from a continental geotherm that can closely predict shear velocities using existing scaling relationships (120). (143; 144) discuss the difficulties in matching velocity models from geophysical observations and compositional, thermal, and anisotropic effects in the Earth. Calculating a shear velocity profile from geotherms cannot, as far as we currently understand it, perfectly predict shear velocity structures. Using different forward modeling methods, with the same input geotherm, can produce shear velocities at the same depth as different as ~ 0.4 km/s. To calculate shear velocities, we use calculations from (118; 145; 124; 146) and an anharmonic calculation to show variance between these methods (figs. A.18, A.19, A.13).

Deviation between observed and modeled V_s are most apparent between 30 and 60 km, where the geotherms change slope to an adiabatic gradient. Determining how to better forward model shear velocities from a geotherm is beyond the scope of this research, so we instead use a geotherm gridsearch, minimizing misfit between predicted and observed V_s between 60 and 120 km, the depths our phase velocity data are most sensitive, to determine which geotherm best predicts our observed shear velocity group aggregates.

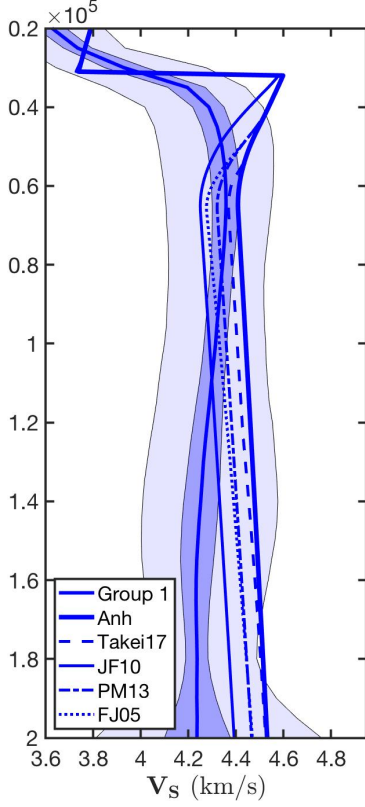


Figure A.18: The range of modeled V_s by varied methods, of the geotherm best fit to group 1. The group 1 velocity aggregate is plotted in solid dark blue, surrounded by its 1σ (dark blue patch) and 2σ (light blue patch). The anharmonic calculation is also plotted in solid dark blue, but is always faster than the velocity profile. Calculated V_s from (118) are in dashed blue, (145) in thin solid blue, (146) in blue dash-hatches, and (124) in blue dots.

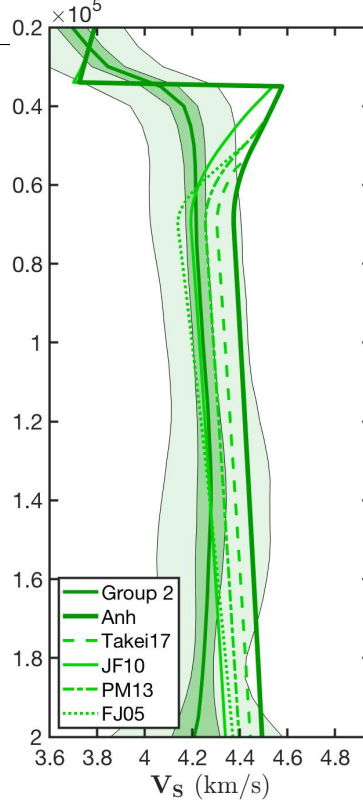


Figure A.19: The same as fig A.18, but for group 2. The group 2 velocity aggregate is plotted in solid dark green, surrounded by its 1σ (dark green patch) and 2σ (light green patch).

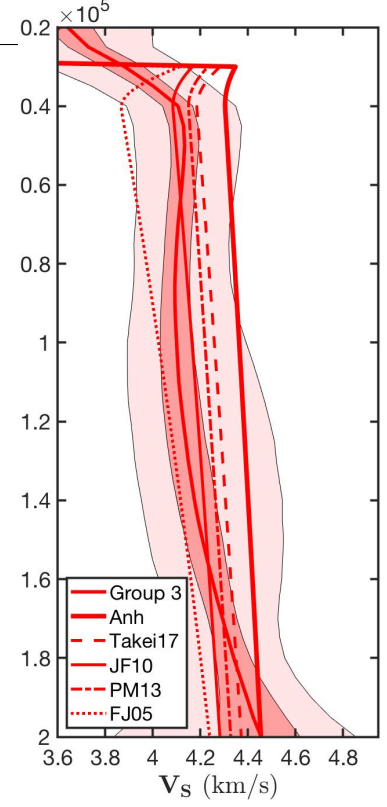


Figure A.20: The same as fig A.18, but for group 3. The group 3 velocity aggregate is plotted in solid dark red, surrounded by its 1σ (dark red patch) and 2σ (light red patch).

A.14 Geotherm Parameter Gridsearch

Because of the inherent difficulty in perfectly matching shear velocity observations forward modeled from a geotherm, we gridsearch crustal thickness, mantle potential temperature (T_m), surface heat flow (Q_0), and radiogenic heat production (A_0), which are shown in table A.14. This determines a range of parameters that produce well-fit shear velocity models by minimizing the misfit between the forward modeled geotherm and the observed shear velocity of each group (figs. A.21, A.22, A.23). Because of the notable spread in shear velocity estimates seen above, we chose to use a single method during gridsearching, (118), as this method generally provided better fits to the observed shear velocities. Note that we do not show the results of crustal thickness, as we found our geotherms insensitive to it.

Range of Parameters Used to Forward-Model Geotherms

T_p ($^{\circ}$ C)	q_0 (mW/m ²)	A_0 (W/m ³)
1350 - 1600	0.05 - 0.15	0 - 1.98 e ⁻⁶

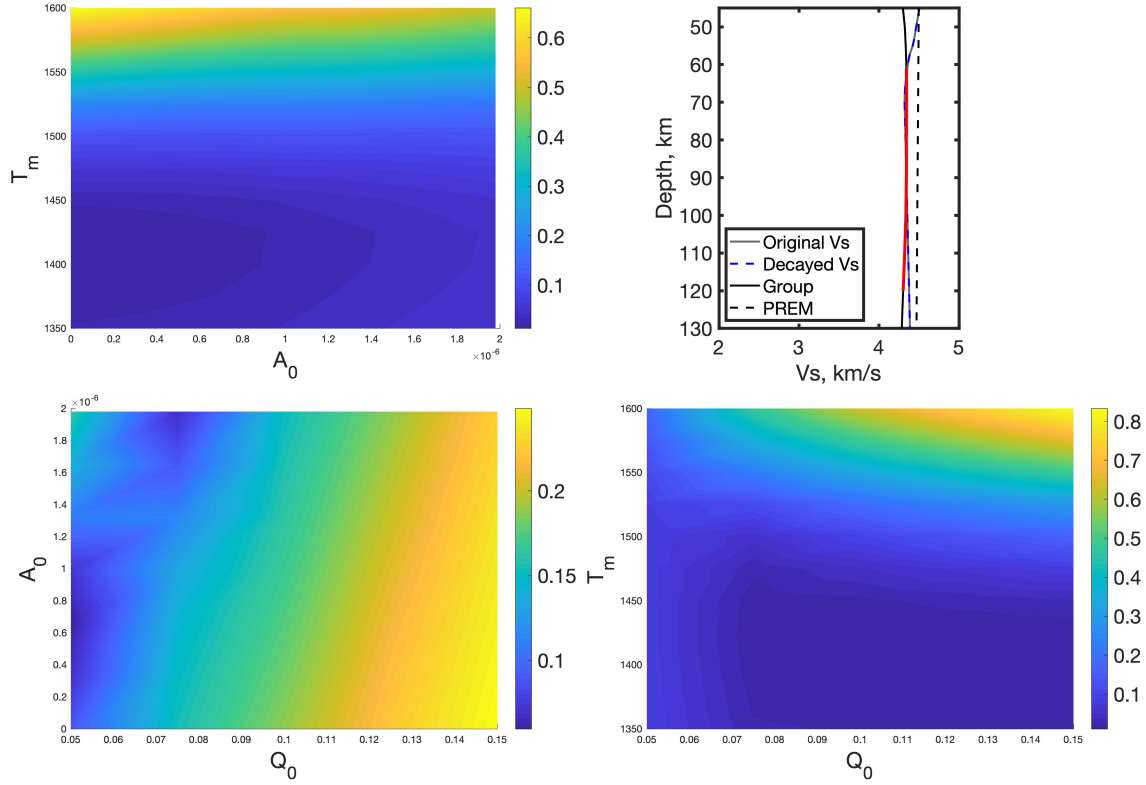


Figure A.21: Misfit surface for Group 1, comparing mantle potential temperatures (T_m), surface heat flow (Q_0) and crustal radiogenic heat production (A_0). Cooler colors indicate lower misfit, warmer colors indicate higher misfit. The shear velocity profile of group 1 (black) is shown in the upper right corner, with the best fit velocity profile before percolation and compaction (grey), and after (dashed blue). PREM is dashed in black, and the depth range used to minimize misfit is highlighted in red on the group V_s profile.

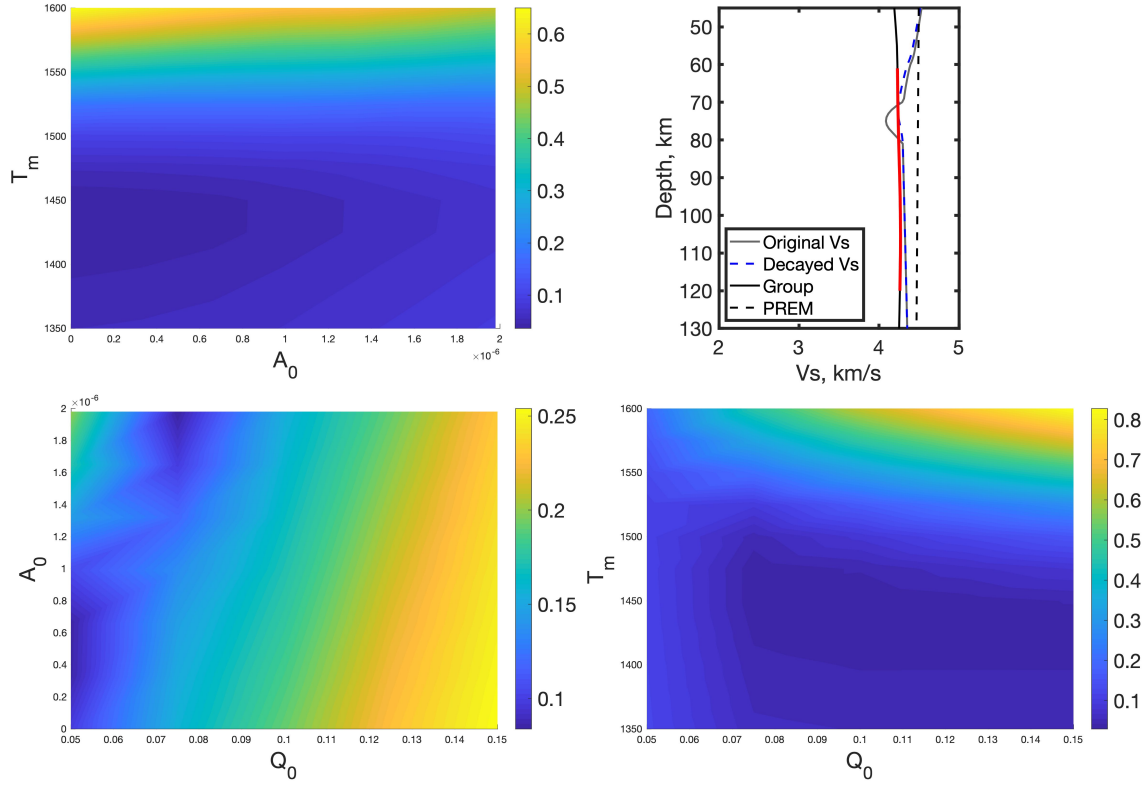


Figure A.22: Misfit surface for Group 2 comparing mantle potential temperatures (T_m), surface heat flow (Q_0) and crustal radiogenic heat production (A_0). Cooler colors indicate lower misfit, warmer colors indicate higher misfit. The shear velocity profile of group 2 (black) is shown in the upper right corner, with the best fit velocity profile before percolation and compaction (grey), and after (dashed blue). PREM is dashed in black, and the depth range used to minimize misfit is highlighted in red on the group V_s profile.

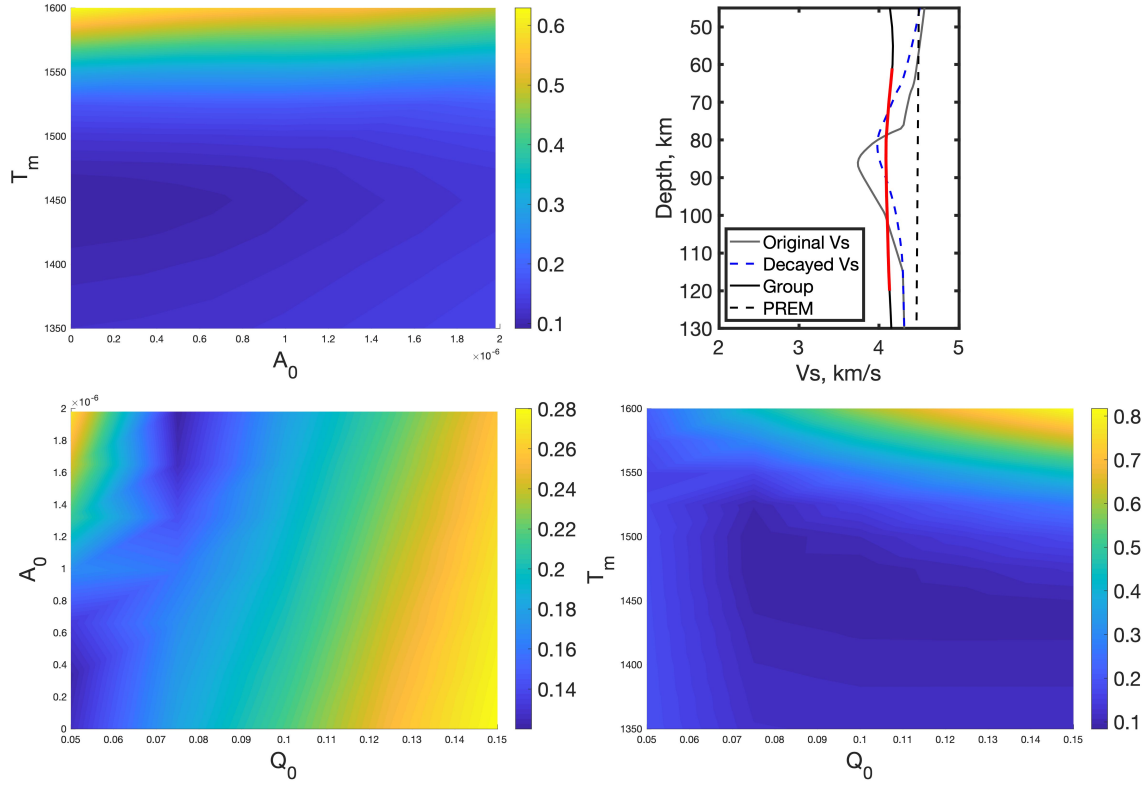


Figure A.23: Misfit surface for Group 3 comparing mantle potential temperatures (T_m), surface heat flow (Q_0) and crustal radiogenic heat production (A_0). Cooler colors indicate lower misfit, warmer colors indicate higher misfit. The shear velocity profile of group 3 (black) is shown in the upper right corner, with the best fit velocity profile before percolation and compaction (grey), and after (dashed blue). PREM is dashed in black, and the depth range used to minimize misfit is highlighted in red on the group V_s profile.

Appendix B

Supplementary Appendices

1. Figure & Text B.2, Dispersion Curve Transects
2. Figure & Text B.3, Phase Velocity V_s Inversions
3. Figure & Text B.4, Group Velocity Maps
4. Figure & Text B.5, Helmholtz Tomography
5. Figure & Text B.6, $\delta V_s/V_s$

B.1 Introduction

Below we present several figures that elucidate aspects of the study region, methodology, and data. These figures were not included in the scientific manuscript due to size constraints, but remain integral in categorizing completed research. Figures B.1 and B.2 display dispersion curve transects across the MER and EP. Figure B.3 shows the shear velocities found by inverting only phase velocities with THBI. Figures B.4 and B.5 show Eikonal tomography stacks and standard deviation using group velocities from the same dataset used to develop phase velocities in section 3.1. Figure B.6 shows the effect of

applying the Helmholtz correction to develop structural phase velocity maps. Figure B.7 shows $\delta V_s/V_s$ of station locations inverted using the phase velocity only inversion.

B.2 Dispersion Curve Transects

Phase velocity maps commonly involve some degree of smoothing in order to account for the pixelized nature of observed velocities at discrete latitude/longitude nodes. This smoothing process inherently causes normalization of observations, and thus introduces bias. In order to more accurately determine phase velocities at lat/lons, non-smoothed dispersion curves were used (fig. B.2). This approach allowed for precise determination of phase velocities, and accurate observation of trends found in the dataset, such as the near-linear increase in phase velocities between 60 and 100s between the rift axis and NWEF.

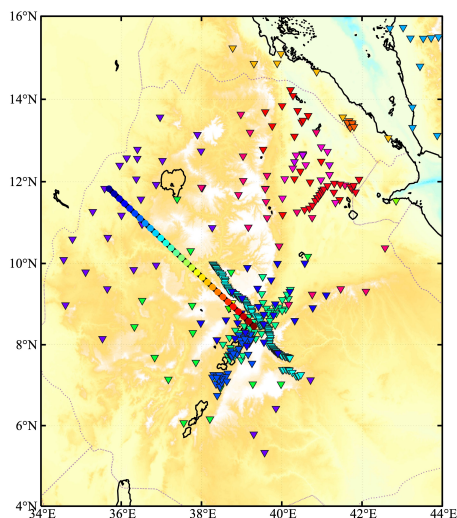


Figure B.1: A transect from the MER into the NWEF flank. Dotted color along the transect indicates the corresponding dispersion curve, in fig B.2. Stations used to develop phase velocity maps seen in fig 4.1, and are color coded by network.

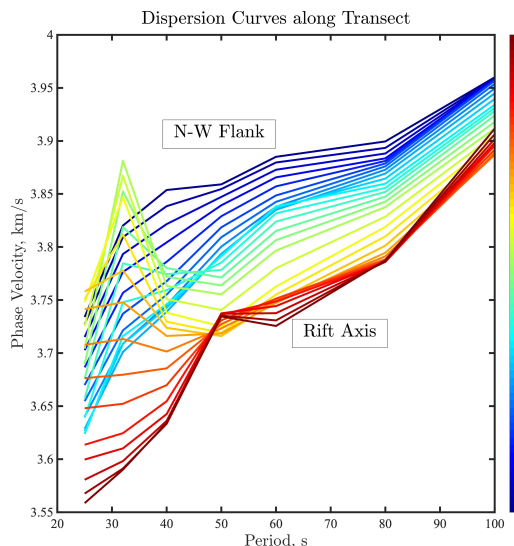


Figure B.2: Dispersion curves plotted traversing from the MER to NWEF. More northerly stations are indicated by darker blue colors, while dark red colors indicate locations near the MER rift proper.

B.3 Phase Velocity V_s Inversions

The main benefits of THBI is the ability to combine different data types to develop a parsimonious Earth velocity model that well-fits each data type. However, it is also able to forward model shear velocity structure using a single data type expeditiously. In early phases of my research, I had developed phase velocity maps, but not receiver functions. It was feasible to develop shear velocity models using only phase velocities, albeit with a known lack of constraint on lateral velocity layers, at individual stations locations. Such an approach is commonplace in the form of ambient noise tomography (13; 14; 128). Observations in figure B.3 reveal similar features to those observed after full joint inversion, namely: distributed slow velocities at mantle depths in the EP, a

slow velocity lineament near Lake Tana, substantial velocity reductions within the MER, with the CMER showing less reduction than the NMER and SMER.

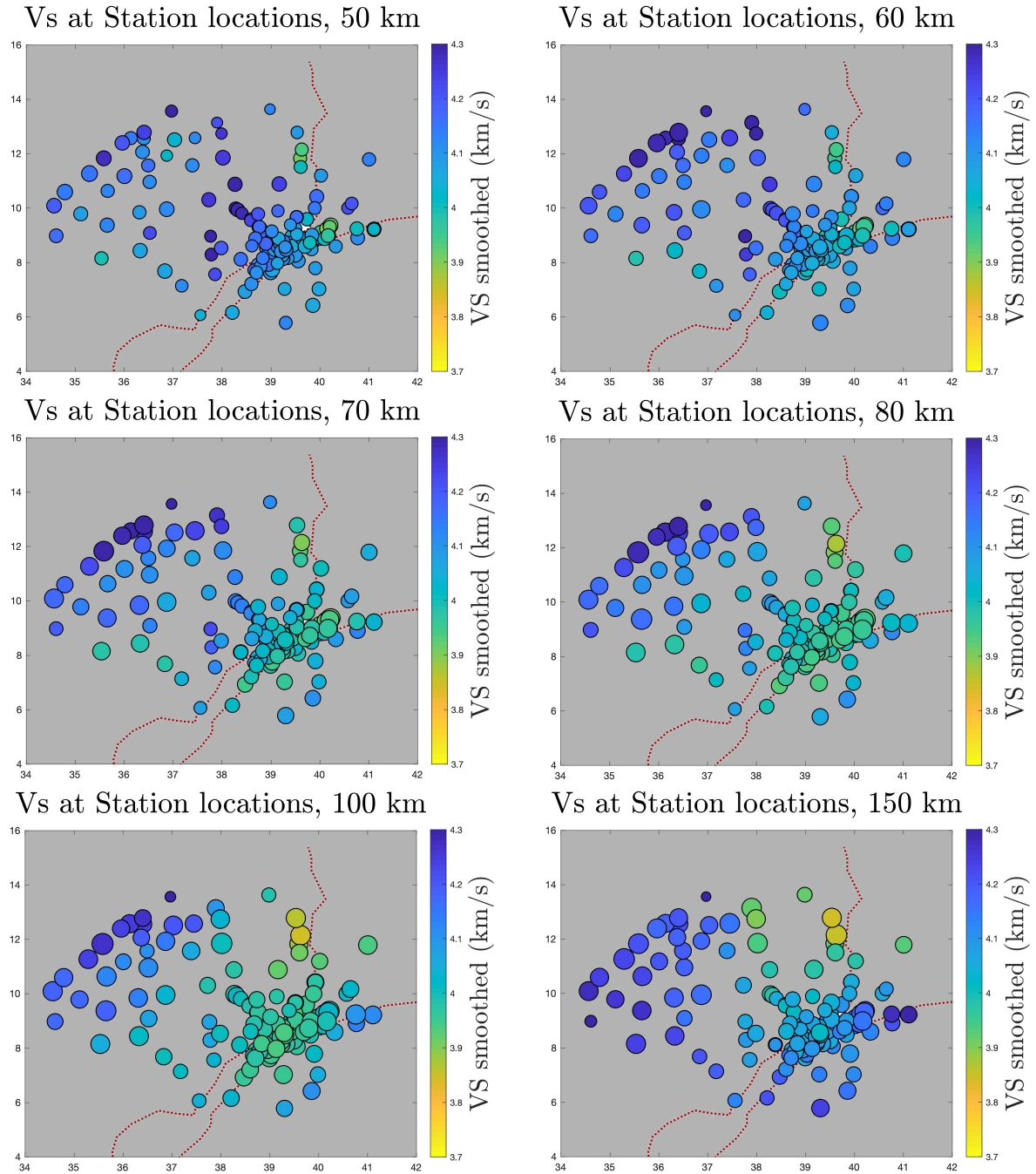


Figure B.3: Shear velocity at depth developed using only phase velocity data. Location size is scaled by $1/\sigma$ at depth, and warmer colors indicate slower shear velocities. The Miocene aged MER border faults are dashed in light red.

B.4 Group Velocity Maps

During narrow band filtering of the cross-correlation function in ASWMS (79), a five parameter wavelet (a Gaussian envelope multiplied by a cosine function) is fit such that:

$$F_i * W_c C(t) \approx A Ga[\sigma(t - t_g)] \cos[\omega(t - t_p)] \quad (\text{B.1})$$

in order to obtain a frequency dependent inter-station phase delays t_p (147). $F_i * W_c C(t)$ is the narrow-band filtered cross-correlogram, Ga is the Gaussian envelope function, A is a positive scaling value, σ is the half-bandwidth, ω the center frequency of the waveform (79). These delays are inverted using the Eikonal equation to develop phase velocity maps. t_g is also fit during this step, but not used subsequently. I modified ASWMS to invert for group velocity maps, using the same methodology as phase velocity maps (fig. B.4).

Group and phase velocities have similar, but not identical, depth sensitivity, and are directly applicable to THBI (33) as an additional data source. Incorporating this methodology provides a straightforward approach that increases the amount of observations from the same dataset by 20 - 51%, based on the metric of ray density. Group velocity maps display many of the same features found in the phase velocity maps. Slow group velocity observations in the MER, NWEF, and Afar are all directly compatible. Furthermore, group velocities identify a small region in the SMER that is less reduced than the surrounding area, also identified using ANT by Chambers et al. 2019, which was not observable in our phase velocity maps. Furthermore, group velocities appear to be biased towards slower group velocities at shorter periods, much more so than phase velocities. This data type is not without problems, however. This smaller dataset loses large swaths of resolvable velocities, particularly at shorter periods.

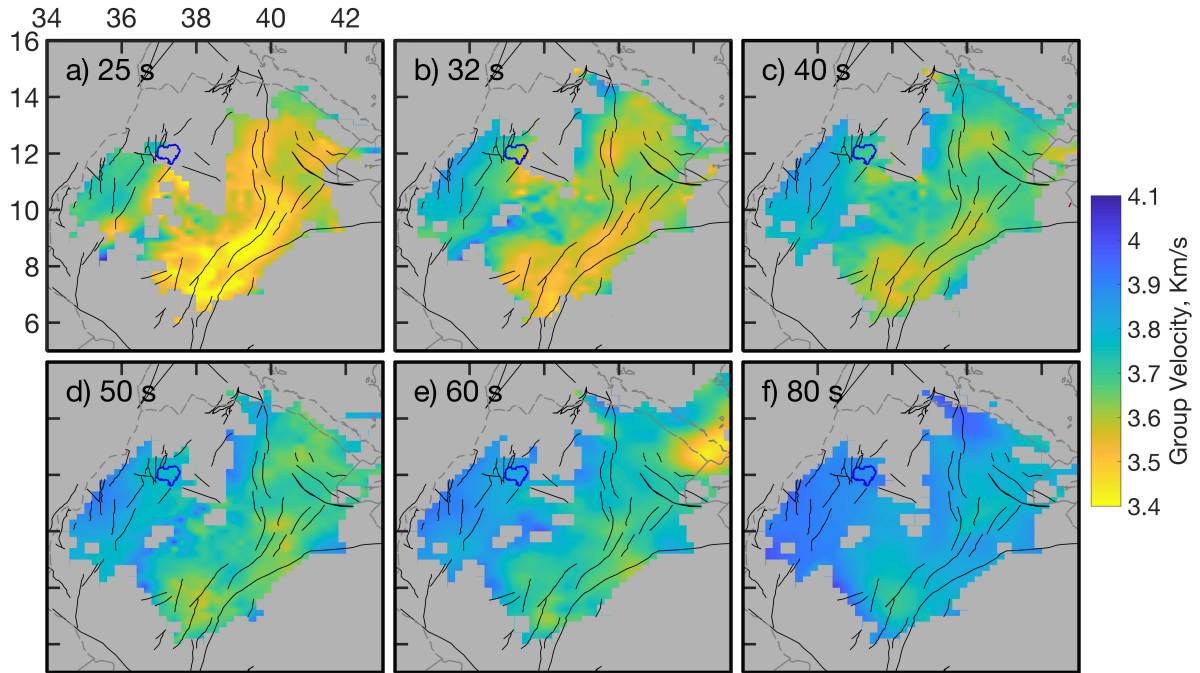


Figure B.4: Group velocity maps, at the same periods as fig 4.1. Semblance contours are not plotted, as this dataset did not undergo resolution testing. Warmer colors indicate slower group velocities, and all periods share the same color scaling. Lake Tana is bounded in blue, and country bounds are dashed in light grey.

Standard deviation per period for group velocity maps shows similar trends to those seen in figs. A.7, A.8 c. Primarily, σ in the NWEP and Afar, and high σ in the MER and Central EP. Both figures show similar σ , despite the vast difference in the number of measurements of each dataset.

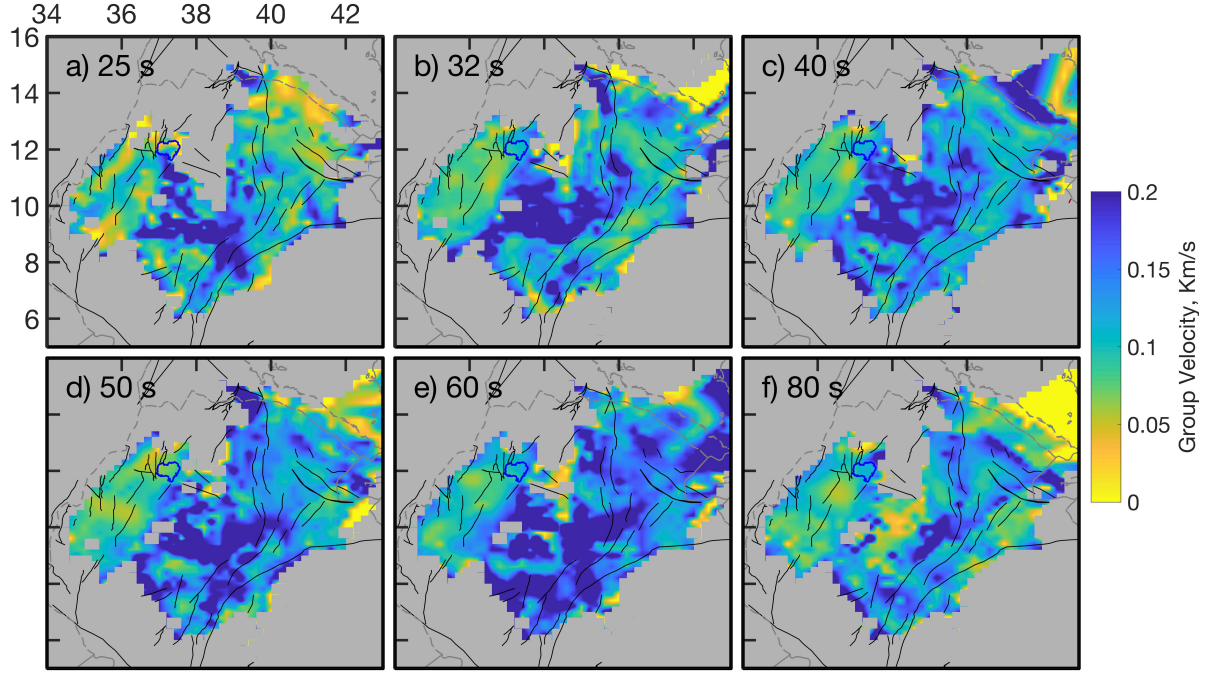


Figure B.5: Standard deviations for group velocity maps, between 25 and 80s, as per fig. A.7. The maximum standard deviation is capped at 0.2 km/s, as observations made for regions with this or a higher σ are not robust.

B.5 Helmholtz Tomography

The Helmholtz equation can be used to correct apparent phase velocity (Eikonal phase velocities) to structural phase velocity ((148; 79). This function is approximated by:

$$\frac{1}{c(\vec{r})} = \frac{1}{c'(\vec{r})} - \frac{\nabla^2 A(\vec{r})}{A(\vec{r})\omega^2} \quad (\text{B.2})$$

where $c(\vec{r})$ is the structural phase velocity, $A'(\vec{r})$ is the amplitude field of the measurement. While the Laplacian term acts as a correction for non-plane wave effects, such effects were minor in this data, or anomalous. For example, application of the Helmholtz correction consistently introduced extremely slow phase velocities in the southwest of

the MER (fig. B.6), despite such a feature not appearing in the eikonal stack. While it is not best practice to reject a dataset based on prior expectations, these observations contradicted both the literature, and my own Eikonal phase velocity measurements. Furthermore, incorporating the Helmholtz equation and stacking to develop a Helmholtz velocity map introduced a degree of computational expense. As per personal correspondence with Ge Jin, the creator of ASWMS, the Helmholtz correction should only slightly change observed phase velocities. Should that not be the case, the amplitudinal term, calculated during inter-station cross-correlation measurement, is likely spurious due to the lack of a waveform normalization step in ASWMS, and therefore the Eikonal equation should be used instead.

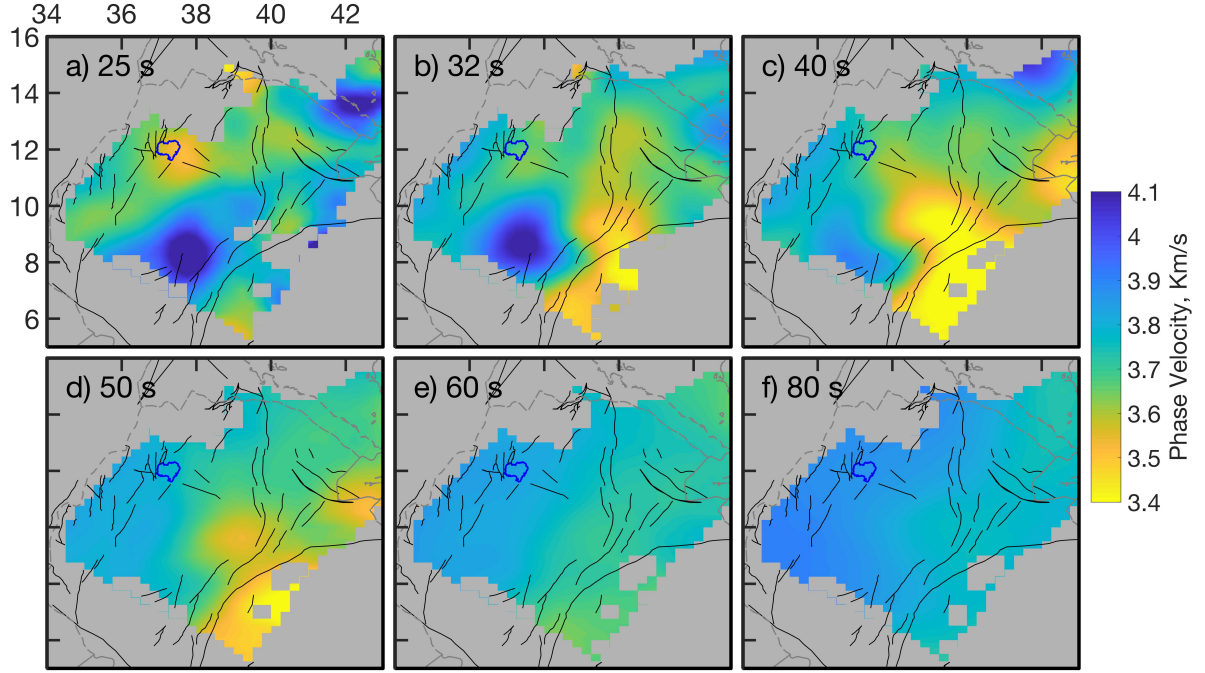


Figure B.6: Phase velocities after application of the Helmholtz correction, as per figure 4.1. While some features, such as the slow velocities near Lake Tana and the MER are consistent between datasets, notable deviations occur in the southwest from 32 (b) - 80s (f).

B.6 $\delta V_s/V_s$

After inverting phase velocities for shear velocity profiles at each station, I use $\delta V_s/V_s$ to identify localized trends in the data. $\delta V_s/V_s$ is a useful metric that disassociates shear velocities from global modes (*e.g.* PREM), and emphasizes dissimilarities within the region. Such an exercise reveals the degree to which the MER and Afar differ from the EP (fig. B.7). The average V_s of the region actually decreases between 50 - 100 km, before increasing as is normally observed with shear velocities at depth. From this, we can posit that melt in the MER and Afar resides above ~ 100 km. The lack of reduces

$\delta V_s/V_s$ at 50 km (fig. B.7a) indicates that crustal melt is either localized to the NMER and Afar, or that shallow crustal melt (≤ 50 km) exists in shallow vertical dikes (14), and not a direct channel to the melt body at depth.

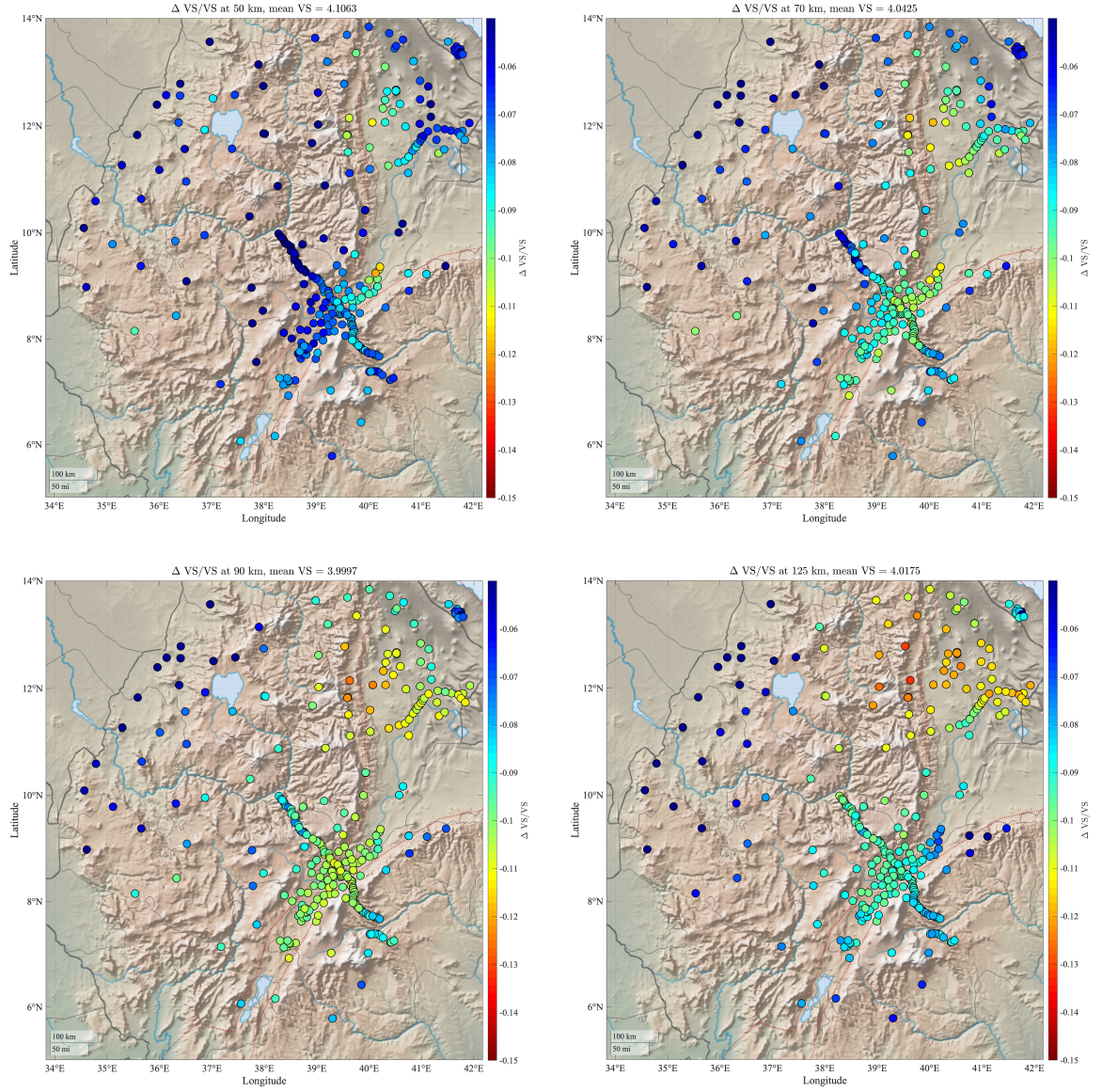


Figure B.7: $\delta V_s/V_s$ at 50, 70, 90 and 125 km depth. Color is fixed, and indicates the reduction in km/s from the average shear velocity across all inverted stations at each depth. Note that the colorbar caps at -0.05, in order to more accurately display areas of reduced $\delta V_s/V_s$

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