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Deciphering the Crustal Structure of the Lerma Valley (NW Argentina): A Multi-Method Seismic Investigation

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Abstract. We investigated the crustal structure beneath the Lerma Valley in northwestern Argentina using data from a local seismic network deployed between 2017 and 2018. This geologically complex transition zone between the Eastern Cordillera and the Sierras Subandinas is characterized by moderate to high seismicity (INPRES, 2024), yet remains largely understudied despite its strategic location within the Andean orogen (Jordan et al., 1983; Allmendinger et al., 1997). Its passive orogenic setting and evidence of inherited structures (Ramos, 2008; Mon and Salfity, 1995; Kley and Monaldi, 2002) make it a natural laboratory for exploring intraplate deformation and foreland basin evolution (Pérez et al., 2016; Tassara et al., 2018). We combined local and teleseismic receiver functions with ambient noise tomography (ANT), jointly inverting Rayleigh wave phase velocities to obtain 1D shear-wave velocity profiles. The results reveal a stratified crust with four main discontinuities at $\sim$ 53–43, 35–30, 10–8, and 1.5–1.2 km, corresponding to the Moho, mid- and lower-crustal boundaries, and the base of the sedimentary basin. A southward-dipping Moho is evident from CCP migration and T-component phase shifts. Velocity profiles also show a north–south contrast: lower velocities (1–2.5 km/s) in the south indicate thicker, less consolidated sediments, while the north exhibits more competent crust (up to 3.5 km/s). The final model comprises five layers, including three sedimentary and two crystalline crustal units. We also introduced a layer-dependent κ correction, revealing a trend from 1.65 at the Moho to 2 in upper layers. These results provide new geophysical constraints on the crustal architecture and tectonic evolution of this underexplored Andean region.

1 Introduction

The Lerma Valley, located in Northwestern Argentina, represents a geologically complex transition zone between the Eastern Cordillera and the Sierras Subandinas (Fig. 1). Characterized by moderate to high and disparate seismicity (INPRES, 2024) in comparison to its surrounding orogenic belts, this region exhibits unique tectonic features that remain largely understudied. Despite its strategic location within the Andean orogen, no detailed geophysical or seismological investigations have been carried out in the valley, leaving significant gaps in our understanding of crustal deformation processes in this area (Jordan et al., 1983; Allmendinger et al., 1997). The basins current structural configuration suggests a passive orogenic regime, where

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deformation is not strongly controlled by active tectonics but rather by inherited structures and long-term crustal reorganization (Ramos, 2008). This makes it a natural laboratory for investigating the dynamics of passive orogeny and foreland evolution in continental interiors.

Geological evidence indicates that the Lerma Valley has undergone a complex tectonic history marked by Paleozoic basement uplift, Cenozoic basin development, and Quaternary fault reactivation (Mon and Salfity, 1995; Kley and Monaldi, 2002). These features offer a valuable opportunity to analyze the interplay between ancient tectonic inheritance and modern stress fields. The lack of systematic geophysical data, including seismic imaging, ambient noise tomography, and receiver-function analysis, underscores the need for comprehensive studies aimed at understanding both the current rheologic and geodynamic behavior and its relation to broader Andean processes. The integration of multidisciplinary geophysical approaches in the Lerma Valley holds the potential to illuminate mechanisms of intraplate deformation and the evolution of passive orogens—topics that remain poorly constrained at a global scale (Pérez et al., 2016; Tassara et al., 2018).

In this context, improving our knowledge of the crustal structure of the Lerma Valley in northwestern Argentina has important implications for the understanding of the Andean crustal characteristics, ongoing orogenesis, and isostatic processes. Moreover, the Lerma Valley and adjacent areas in the Santa Bárbara System has a very active seismogenic history with several destructive events with $Mw > 5$. Recent events include the Mw 6.1 2010 Salta earthquake and the 1913 La Poma event (INPRES, 2024). In the Santa Bárbara System the Mw 5.8 2015 El Esteco, the 1825 Anta, and the 2015 el Galpón earthquakes testify to present-day seismogenic activity that reflects the stress transfer from the active continental margin to the orogenic hinterland. The destruction related to the 2015 El Galpón earthquake and the damage-buildings suffered by the 2010 Salta earthquake are testimony of potential high-acceleration zones in this region. Recent studies conducted in the vicinity of the Salta city (in the center of Lerma valley) have revealed the presence of unconsolidated sediments within the first 25 meters below the surface (Orosco et al., 2007, 2010). It has been demonstrated that these sediments are susceptible to water saturation after heavy rainfalls during the austral monsoon season these unconsolidated deposits have important implications for site effects and amplification phenomena.

The main geological structures of this area remain poorly characterized at depth, and they are very complex due to the existence of Cretaceous extensional faults that have been subjected to contractional inversion during Cenozoic Andean mountain building. A detailed characterization of the basin sediments is of paramount importance for further seismological and geotechnical applications and mitigation efforts. In addition, the deeper crustal structures are poorly known. For example, the boundaries for the upper, middle and lower crust were only studied for the northern and southern limit of the region covered by our study. The thickness of the crustal units was first established by Cahill et al.(1992) in his study of the seismicity of the Zapla ranges in the province of Jujuy, which provided a depth of $42km$ for the Moho. Thirty years later, Zeckra(2020) presented a model for the crust that placed the Moho at $46km$ to the southeast of our study region. However, deriving detailed velocity models was not the aim of neither of these previous studies, as the models were derived from inversions of the travel times of seismic phases of local crustal earthquakes for better location.

The limitations of traditional seismic methods that rely on active seismic sources include their limited spatial coverage and the associated implementation costs. In contrast, ambient noise tomography (ANT) uses records of the seismic ambient noise

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wavefield at different locations to passively probe subsurface structures. By cross-correlating such records between two seismic stations, it is possible to extract coherent signals that are, under certain assumptions, proportional to the Green's function between the pair of stations (Wapenaar, 2004; Stehly et al., 2006; Bensen et al., 2007). As complementary information to that provided by the ANT, receiver functions (RF) contain information related to the seismic discontinuities in the subsurface, which can in turn be used in the inversion of velocity models based on the dispersion curves calculated for the surface wave part of empirical Green's functions (Julia et al., 2000).

The goal of this study is to develop a detailed velocity model that includes the lower, middle, and upper crust. This model will be constrained by receiver function results and phase velocity dispersion curves (obtained from ANT); when jointly inverted these will provide a local S-wave velocity model that accounts for the discontinuities at different scales. These discontinuities will then be compared to those proposed by previous studies for the uppermost units of the upper crust.

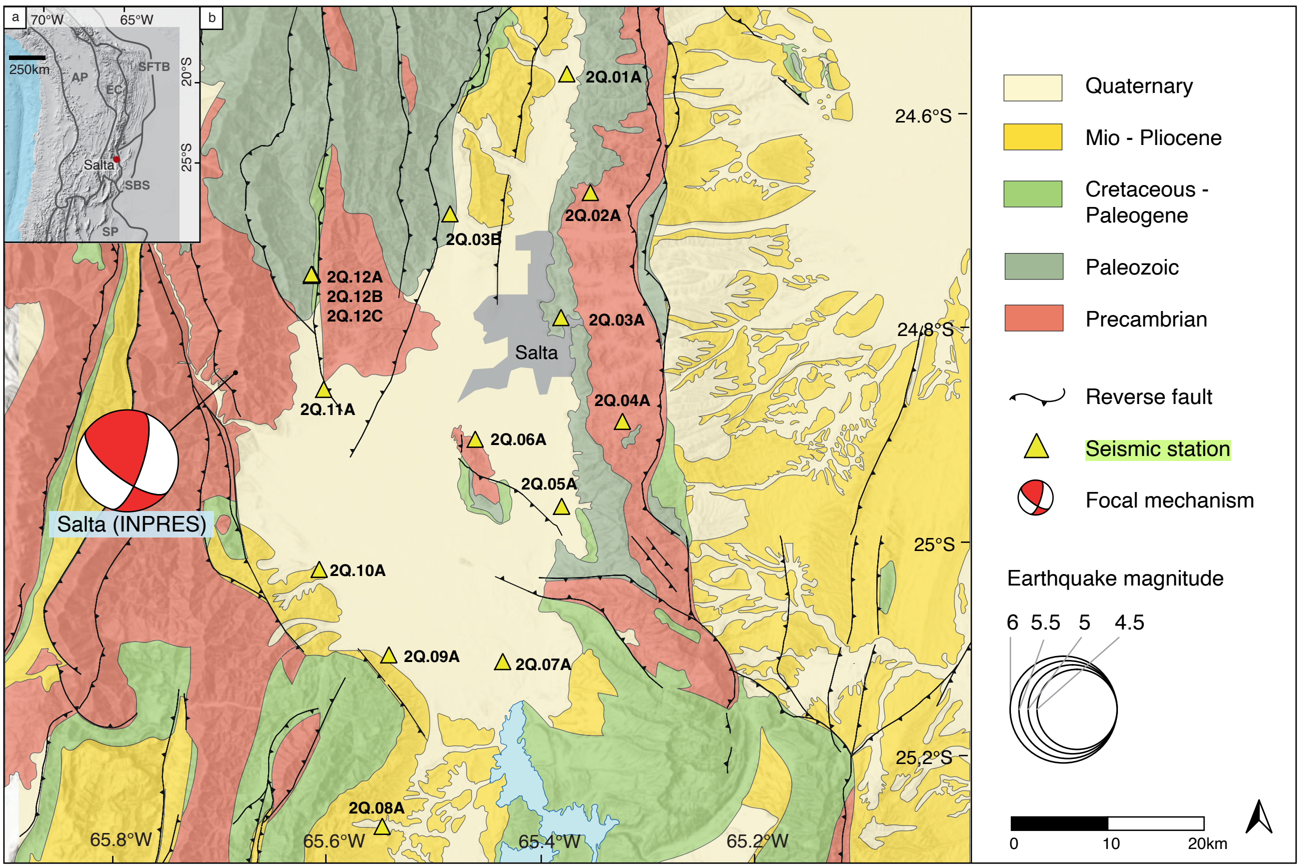

Figure 1. a) Location map in context of the geological provinces: SFTB: sub-Andean fold-and-trust belt, AP: Altiplano-Puna, EC: Eastern Cordillera, SBS: Santa Bárbara System and SP: Sierras Pampeanas (modified after Jordan et al. (1983)) b) LEVARIS Network over the Lerma Valley with lithologies and structures, modified from García et al. (2013).

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2 Geological setting

The studied area encompasses the Lerma Valley, an approximately 150-kilometer-long, north-south-oriented intermontane basin in the Eastern Cordillera of Argentina. The basin is flanked by basement-cored ranges (Pascha and Lesser ranges in the west, and Mojotoro-Castillejo-El Cebilar ranges in the east) delimited by reverse faults with both east and west vergence. These main structures correspond to inverted Cretaceous normal faults and Paleozoic faults which were reactivated during the Andean orogeny (Grier et al., 1991; Mon and Hongn, 1991; Mon and Salfity, 1995). One of the most important structures in the area is the regional Calama-Olacapato-Toro (COT) lineament. This NW-SE trending structure crosses the Lerma Valley and could have exerted a tectonic control over the Paleozoic depostis and the Salta Group rift sequences to the north and south, respectively (Moya, 1988; Marquillas et al., 2005). Marrett and Strecker (2000) and Hongn and Seggiaro (2001) postulated a main transcurrent sinistral movement for this segment of the lineament, which could reflect the differential blocks movements both towards the north and the south.

The stratigraphic succession that crops out along the valley and into the bounding ranges is composed by:

1. Neoproterozoic-Lower Cambrian metasediments of *Puncosviscana Formation* (Turner et al., 1979)
2. Cambro-Ordovician quartzites, marine shales and sandstones from the *Mesón* and *Santa Victoria* Groups (Turner, 1960)
3. Cretaceous-Paleogene rift deposits of *Salta* Group mainly composed of mudstones, sandstones and carbonates (Moreno, 1970)
4. Miocene-Pleistocene continental sequences from *Orán* Group includes conglomerates and sandstones (Russo, 1972)
5. Quaternary fill of the valley was separated into three main units, the Calvimonte, Tajamar and La Viña Formations (Gallardo et al., 1996) formed by fluvial-alluvial and lacustrine deposits.

A comprehensive review of the stratigraphy of the Lerma Valley can be found in García et al. (2013).

3 Data and Methods

3.1 Installation details of the seismic network

In August 2017, a temporary seismic network, the Lerma Valley Ring Installation of Seismometers (LEVARIS, (Criado-Sutti et al., 2017)) was installed in the studied area. The network spanned the central and northern regions of the valley and operated for a total of thirteen months. Prior to this deployment, there was only one permanent short-period station within the valley, managed by Argentine agency INPRES (code SLA, (INPRES, 2024)).The dimensions of our temporary network spanned approximately 80 km in a north-south direction and 30 km in an east-west direction, with stations strategically located to ensure safety, accessibility, and minimal interference from anthropogenic noise sources. The seismic stations were equipped with a DATA-Cube3 type digitizer paired with a Lennartz 3D/5s sensor. One of the installations (2Q.09A in Fig. 1) used a

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Mark L-4C-3D short-term sensor. In all cases, instruments were buried at an approximate depth of 60 cm. The Data-Cube3 digitizers were set to a sampling rate of 100 Hz, and the stations were powered by batteries connected to solar panels.

NETWORK	CODE	LOCATION	LATITUDE [°]	LONGITUDE [°]	ELEVATION [m]	RECORDING TIME [days]
2Q	01A	Campo Alegre	-24.56889	-65.37404	1460	397
	02A	Gallinato	-24.67945	-65.35223	1304	217
	03A	Cerron San Bernardo	-24.79603	-65.37949	1222	197
	04A	La Quesera	-24.89281	-65.32248	1445	396
	05A	Ceibalito	-24.97199	-65.37888	1146	395
	06A-B	Cerrillos	-24.90947	-65.45905	1220	396
	07A	Calvimonte	-25.11671	-65.43352	1122	398
	08A	Potrero de Díaz	-25.27033	-65.5453	1263	398
	09A	Chicoana	-25.11049	-65.53912	1270	394
	10A	Corralito	-25.03078	-65.60371	1359	397
	11A	La Silleta	-24.86299	-65.59959	1440	393
	12A-B	Potrero de Uriburu	-24.75678	-65.61088	1653	190
	12C	Potrero de Uriburu	-24.755603	-65.610981	1732	200

Table 1. Location of the stations of the LEVARIS temporary network, with their approximate recording time in days.

3.2 Methods

In order to study the various discontinuities of the crust below the grater Lerma Valley and to derive local velocity models, we employed three methods: receiver functions (teleseismic and local), ambient noise cross-correlation tomography, and joint inversions (forward modeling). The first two methods involved processing the raw data from the LEVARIS network (see section 3.1, (Criado-Sutti et al., 2017)) to produce receiver functions and dispersion curves. These latter results were then combined to be inverted using forward modeling and fitting the data with S-wave velocity model, thus obtaining a representative crustal model. In the following subsections we present and briefly describe each method and also provide a complete description of the parameters used in the data processing.

3.2.1 Teleseismic Receiver Functions (RFs)

As seismic waves from distant earthquakes (teleseisms) travel through the Earth's interior, they undergo reflections and P-to-S conversions at interfaces such as the crust-mantle boundary (the Moho). Receiver function (RF) analysis enables the detection of these converted phases, providing insights into the subsurface structure beneath the region covered by a seismic station.

The RF method, originally developed for teleseismic analysis (Langston, 1977; Vinnik, 1977; Burdick and Langston, 1977), involves deconvolving the vertical component from the horizontal components of a rotated seismogram to isolate the Earth's impulse response (Ligorría and Ammon, 1999) beneath the seismic station. This procedure suppresses the effects of the source-

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time function and distant propagation path, highlighting converted arrivals such as the Ps phase. Arrival times of these converted phases can be associated with structural discontinuities, provided a reference velocity model is available.

To improve spatial resolution and imaging of discontinuities such as the Moho, Common Conversion Point (CCP) stacking is employed. CCP stacking allows a pseudo-migration of RFs from the time domain to depth by tracing converted phases back into the Earth along theoretical ray paths using a local velocity model. This approach helps account for lateral heterogeneity and enhances structural imaging, particularly when focusing on strong, isolated phases like Ps, which are typically more prominent and interpretable than crustal multiples (Dueker and Sheehan, 1997; Audet, 2015).

3.2.2 Local Receiver Functions

Local deep earthquakes provide an alternative source for RF analysis. These events have shorter source durations and higher-frequency content than teleseismic events, enabling better resolution of fine-scale features and sharper discontinuities (Ammirati et al., 2016; Perarnau et al., 2012). The methodology for local RFs mirrors that of the teleseismic case, though tailored to events with steep incidence angles and originating from shallower depths (up to 150 km). In our study area, most local deep events are located in the Jujuy cluster (Mulcahy et al., 2014; Valenzuela Malebran, 2022), providing narrow coverage of the Moho, which lies at depths between 40 and 50 km (Cahill et al., 1992; Zeckra, 2020).

Teleseismic and local receiver functions were computed from three-component waveforms recorded at the LEVARIS stations. For teleseismic events, we selected earthquakes with magnitudes greater than or equal to 5.5 and epicentral distances between 30° and 90°, using data obtained from the IRIS web services. The theoretical arrival times of the direct P-waves were computed with the ak135 velocity model (Kennett et al., 1995) using the Cake software package (Heimann et al., 2017). For each event, we extracted time windows beginning 10 seconds before and ending 80 seconds after the expected P-wave arrival, and rotated the data into the LQT coordinate system to isolate the P, SV, and SH wave components.

Local deep earthquakes were analyzed using the same general workflow, with the main difference being the event catalog, which was constructed specifically for this study based on the LEVARIS network (Criado-Sutti et al., 2017). For both teleseismic and local events, we applied a bandpass filter from 0.01 to 2.0 Hz to isolate the relevant frequency band. To ensure data quality, we extracted 300-second noise windows ending 10 seconds prior to the P-wave arrival and computed the RMS of both noise and signal windows, discarding traces where the RMS ratio was below 1.5. Deconvolution was performed using the water-level method (e.g., Langston, 1977), with a Gaussian filter width of $a = 0.5$ and a water-level parameter of $c = 0.1$. A subsequent manual inspection step was used to remove traces with excessive noise or anomalous amplitudes. The resulting quality-controlled receiver functions were used to identify P-to-S converted phases and to estimate crustal properties, including Moho depth and v_p/v_s ratio, via the H-k stacking technique (Zhu and Kanamori, 2000).

3.2.3 *H*-*k* Analysis

The H-k stacking method, introduced by (Zhu and Kanamori, 2000), is a widely used technique for estimating crustal thickness (H) and the v_p/v_s ratio (k) by analyzing teleseismic receiver functions. The method relies on identifying the arrival times of converted and multiple seismic phases, such as Ps, PsPs, and PpSs. When appropriate values of H and k are found, the sum of

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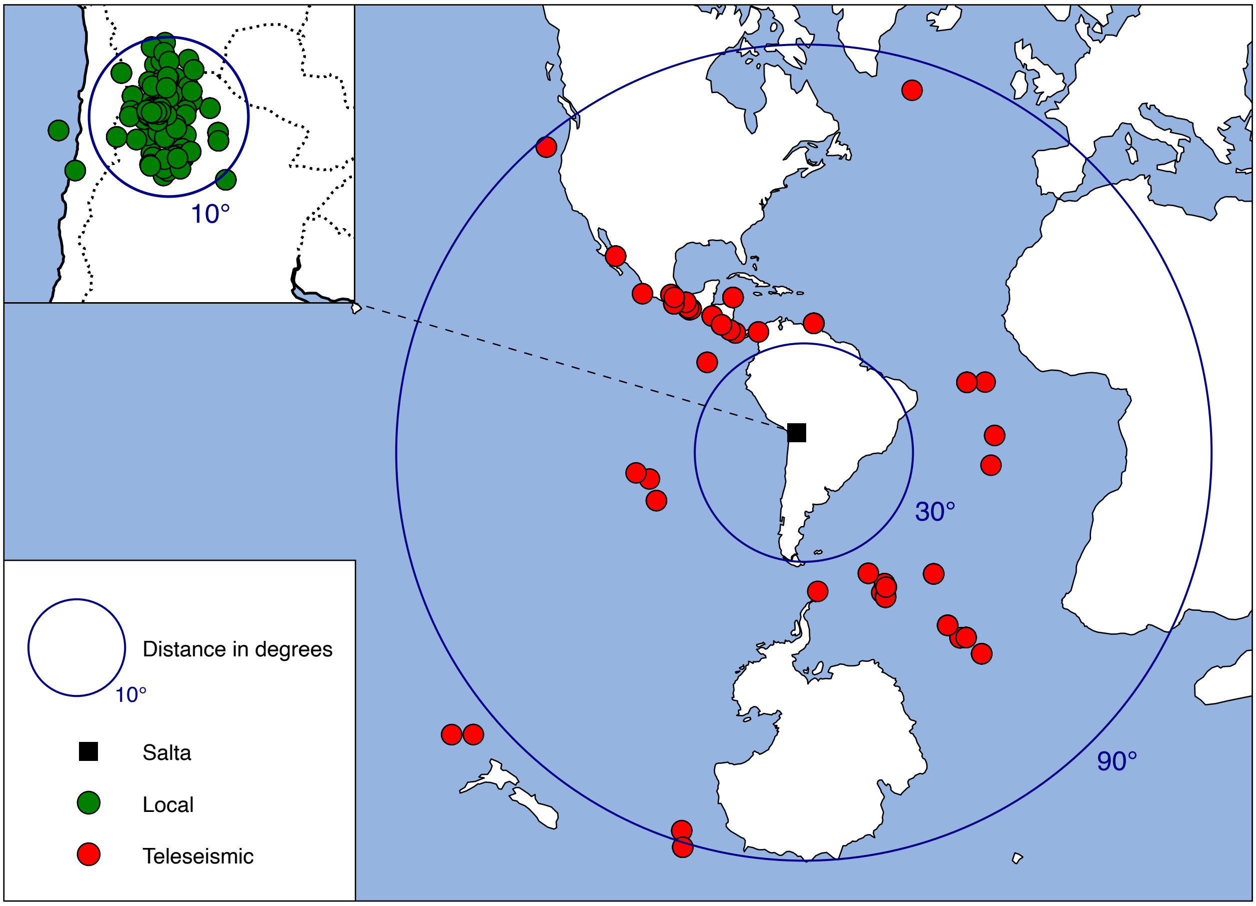

Figure 2. Events distributions for teleseismic (red) and local (green) events used for the receiver functions in an equidistant plot.

the amplitudes of the receiver functions at the corresponding travel-times interfere constructively, allowing the determination of crustal discontinuities by locating maxima in the stacking function.

In our implementation, we assumed a fixed P-wave velocity of 6 km/s. The analysis was performed on a grid with 2 km increments in depth (H) and 0.05 increments in the v_p/v_s ratio (k). The bounds of the grid search were set from 0 to 70 km for H and from 1.6 to 2.5 for k. These parameter ranges and step sizes were selected to ensure adequate resolution while maintaining computational efficiency. We estimated the uncertainties in the parameters H and k following the method of Eaton et al. (2006), who proposed defining a contour line at one standard error below the maximum stack amplitude. The standard error is given by $\left(\sigma^2/N\right)^{1/2}$, where σ^2 is the variance and N is the number of stacked receiver functions. This method implies that confidence in the estimated parameters increases with the number of receiver functions included in the stack.

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3.2.4 Estimation of Effective k Values.

The H-k technique provides only an average value for all the layers above a given seismic discontinuity. Therefore, to better understand the variation of the v_p/v_s ratio (denoted as k) with depth, we computed an *effective* k value for each observed layer stack thickness H using a 1D velocity model. This 1D model provides depth-dependent values of compressional (v_p) and shear-wave (v_s) velocities. For this purpose, the effective k is defined as the ratio of the travel-time integral of v_p to that of v_s down to the given depth H:

$$k_{\text{eff}}(H) = \frac{\int_0^H \frac{1}{v_s(z)}\,dz}{\int_0^H \frac{1}{v_p(z)}\,dz} \tag{1}$$

This formulation accounts for changes in seismic velocities with depth and provides a physically consistent comparison to the measured k values, which assume a constant velocity structure. The error associated with the effective k is estimated by propagating uncertainties in H using the bounds H_{min} and H_{max} reported from the receiver function inversion:

$$\sigma_k = \frac{1}{2}\left|k_{\text{eff}}(H_{\text{max}}) - k_{\text{eff}}(H_{\text{min}})\right| \tag{2}$$

It is important to note that the calculation assumes vertical incidence of incoming seismic waves. If the incidence angle deviates significantly (e.g., by more than $15°$), the approximation may introduce systematic bias in the resulting k values. This formulation helps us to lately derive a weighted average definition for each k, and thus establish a recursive strategy to correct the measured values, as we will see next.

3.2.5 Adjusting the Measured k Values.

To refine the measured k values based on our velocity model, we computed the *real* k for each layer. Since the shear-wave velocity v_s is typically not directly measured, we instead assume a constant average v_p/v_s ratio $\langle k\rangle$ with an associated uncertainty. This allows us to estimate v_s from v_p for the Moho, and then recursively reconstruct each true k_i value that contributes to the measured k.

The measured k_{meas} is treated as a weighted average, derived from the effective velocities of the layered medium. Considering a stack of n horizontal layers with thicknesses H_i, P-wave velocities $v_{p,i}$, and v_p/v_s ratios $k_i = v_{p,i}/v_{s,i}$, the total P- and S-wave travel times through the layers are respectively

$$t_p = \sum_{i=1}^{n} \frac{H_i}{v_{p,i}}, \quad t_s = \sum_{i=1}^{n} \frac{H_i}{v_{s,i}} = \sum_{i=1}^{n} \frac{k_i H_i}{v_{p,i}}.$$

The effective velocities are given by $v_p^{\text{eff}} = \frac{\sum_i H_i}{t_p}$ and $v_s^{\text{eff}} = \frac{\sum_i H_i}{t_s}$, so the effective k ratio becomes

$$k_{\text{meas}} = \frac{v_p^{\text{eff}}}{v_s^{\text{eff}}} = \frac{t_s}{t_p} = \frac{\sum_{i=1}^{n} \frac{k_i H_i}{v_{p,i}}}{\sum_{i=1}^{n} \frac{H_i}{v_{p,i}}}.$$

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Multiplying numerator and denominator by $v_{p,i}^2$ yields the weighted average expression

$$k_{\text{meas}} = \frac{\sum_{i=1}^{n} k_i \cdot v_{p,i} \cdot H_i}{\sum_{i=1}^{n} v_{p,i} \cdot H_i}, \tag{3}$$

which shows that the effective k is a velocity-thickness weighted average of the individual layer ratios.

We solve this equation recursively for each layer k_i, using known $v_{p,i}$, inferred $v_{s,n} \approx v_{p,n}/\langle k \rangle$ for the Moho, and layer thickness H_n. This procedure allows us to reconstruct a physically consistent, depth-varying k profile that agrees with the measured value at the surface while incorporating the velocity model and adjustable v_s values.

3.2.6 Bias Introduced by Non-Vertical Incidence.

The new method presented here for estimation of crustal thickness H and k in receiver function analysis, assumes vertical incidence of the incoming P-wave. However, for teleseismic events, the incidence angle θ may differ significantly from vertical. This introduces a systematic bias in both H and k, since the actual wave paths are longer and deviate from the vertical.

Assuming a plane-layered Earth and using the ray parameter p, the apparent slowness, the bias in δk can be estimated using the modified travel-time equations:

$$\delta k(\theta) \approx \frac{t_{Ps}(\theta)}{t_{PpPs}(\theta)} = \frac{\sqrt{\left(\frac{1}{v_s^2} - p^2\right)} - \sqrt{\left(\frac{1}{v_p^2} - p^2\right)}}{2\sqrt{\left(\frac{1}{v_p^2} - p^2\right)}} \tag{4}$$

For a typical incidence angle of $\theta = 15°$, we compute the ray parameter $p = \frac{\sin\theta}{v_p}$, and compare the result to the vertical case $\theta = 0$. Assuming representative crustal values (e.g., $v_p = 6.5\,\text{km/s}$, $v_s = 3.75\,\text{km/s}$), the relative bias in k can be estimated as:

$$\delta k(15°) = \frac{k(15°) - k(0°)}{k(0°)} \approx +5\%$$

This means that neglecting an incidence angle of $15°$ may lead to an overestimation of k by approximately 5%, depending on the exact velocity structure and event distance. Such biases should be considered when interpreting k values derived from steeply incident teleseismic arrivals.

3.2.7 Ambient Noise Tomography (ANT)

To estimate empirical Green's functions between receiver pairs within the LEVARIS network, we applied ambient noise cross-correlation techniques to continuous seismic data (Table 1). The available recordings were segmented into two-hour windows, detrended, cosine-tapered (5%), and corrected for instrument response. Cross-correlations were then computed by spectral multiplication in the frequency domain, following the method of (Ekström, 2014):

$$\rho_{ijk} = \frac{u_{ik}(\omega)\, u^{jk}(\omega)}{\sqrt{u_{ik}(\omega)\, u^{ik}(\omega)}\sqrt{u^{jk}(\omega)\, u_{jk}^{*}(\omega)}}, \tag{5}$$

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where ρ_{ijk} is the cross-correlation for stations i and j in time window k, u represents the Fourier-transformed time series, and * denotes complex conjugation. The resulting cross-correlograms were stacked across the entire deployment period, and a time-scale phase-weighting scheme (Ventosa et al., 2017) was applied to enhance signal-to-noise ratios prior to further analysis.

The ambient noise dataset from the LEVARIS network was organized using the Pyrocko-based "jackseis" tool (Heimann et al., 2017), with daily MiniSEED files sorted by component and stored in annual station-specific folders using Julian day naming conventions. Cross-correlations were computed as described above for all possible vertical-component station pairs using 1-hour windows, and were then stacked over the entire deployment period to improve coherence. Dispersion measurements were obtained using time-frequency analysis to pick group velocities, following the method of (Bensen et al., 2007), and phase velocities were estimated by numerical integration. To address the $2\pi N$ ambiguity in phase velocity curves, we selected the curve that remained closest to the corresponding group velocity without being slower, as recommended by (Bensen et al., 2007).

Subsequently, the derived dispersion curves were used to produce surface wave tomographic maps based on the method of Barmin et al. (2001), which assumes surface waves propagate along great-circle paths between stations. The tomographic inversion was conducted in two stages. In the first inversion, strong regularization parameters were applied ($\alpha = 1000$, $\beta = 50$, and $\sigma = 400$ km) to generate oversmoothed velocity maps for quality control, following procedures outlined in (Barmin et al., 2001). Measurements that deviated by more than two standard deviations from the mean phase or group velocity were flagged and removed. A second inversion was then performed using the same regularization parameters to produce the final phase velocity maps. The regularization involved a balance between smoothing and fidelity to the data, and parameter values were chosen through trial-and-error (Barmin et al., 2001), with visual inspections confirming that small perturbations in the chosen parameters did not significantly affect the resulting maps.

3.2.8 Joint Inversion of RFs and Phase Velocity Dispersion Curves using Hamiltonian Monte Carlo (JIHMC)

The Hamiltonian Monte Carlo (HMC) inversion method (Betancourt and Girolami, 2015; Betancourt, 2017) provides a robust framework for exploring complex posterior distributions by leveraging an energy-based sampling approach that minimizes the misfit between observed and synthetic data. This technique is particularly well-suited for seismic inversion problems due to its ability to efficiently explore high-dimensional parameter spaces with strong correlations.

For our local model inversion, we adopted a modified version of the velocity structure proposed by (Zeckra, 2020) as a baseline (Table 2). Although alternative models were considered, including a preliminary 1D model derived from a VELEST inversion of local events, these alternatives proved unstable and were ultimately not used.

The joint inversion was carried out using the RfSurfHmc software package (Quang-Duc, 2021), a Python-based framework with C-based computational kernels that implements the HMC approach developed by (Betancourt and Girolami, 2015; Betancourt, 2017) and later integrated with the EvodCinv platform (Luu, 2018). The RfSurfHmc (Quang-Duc, 2021) tool enabled the simultaneous inversion of teleseismic receiver functions and phase velocity dispersion curves to construct station-specific shear wave velocity profiles.

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Depth [km]	v_p [km/s]	v_s [km/s]	v_p/v_s
0	2.90	1.75	1.70
1	4.16	2.83	1.45
3.5	5.71	2.83	2.02
8.5	5.81	3.30	1.76
36	6.65	4.33	1.54
46	8.04	4.49	1.79

Table 2. Modified velocity model derived from Zeckra (2020), with the second layer subdivided into two layers of 1 and 3.5 km thickness, showing depth, velocities, and v_p/v_s ratios.

The input data included stacked receiver functions in the time range from 0 to 10 seconds and surface wave dispersion curves from 1.7 to 10 seconds. Inversions were performed using data from all LEVARIS stations (see Fig. 1) to resolve broader basin-scale features. The inversion was run for 200 iterations, with misfit weighting parameters set to $\sigma_{rf} = 1 \times 10^{-3}$ for receiver functions and $\sigma_{swd} = 0.7 \times 10^{-2}$ for surface wave dispersion curves.

The forward modeling step used a Gaussian filter with parameters $a = 1.5$ and $c = 0.001$, and a time step of $dt = 0.1$ s. We used a ray parameter of $0.045\,s/^\circ$ and explored the depth range from 0 to 50 km ($L =$ 0–50 km). These parameter choices were based on sensitivity tests and prior studies, and were verified to ensure that small variations in their values did not significantly alter the inversion results.

3.2.9 Inversion of Surface Wave Phase Velocity Dispersion Curves with Evolutionary Algorithm (IEA)

Evolutionary algorithms (EAs) are optimization techniques inspired by the principles of natural selection and genetics. These methods are particularly well suited for exploring large, complex solution spaces where conventional optimization strategies may struggle due to non-linearity, high dimensionality, or multimodal objective functions (Mitchell, 1998; Deb, 2001). EAs have seen widespread application in various fields such as machine learning, computational biology, and geophysical inversion, offering a flexible and robust approach to finding globally optimal solutions (Mitchell, 1998; Deb, 2001).

In this study, we employed an evolutionary algorithm to invert surface wave phase velocity dispersion curves, following the approach described by (Luu, 2018). This method was particularly effective in enhancing resolution in the upper five kilometers of the crust, where conventional methods often lack sensitivity.

The inversion was carried out on phase velocity dispersion curves measured over periods ranging from 1.7 to 9.9 seconds. The evolutionary algorithm was initialized with a population size of 20 and a random seed set to zero to ensure reproducibility. The optimization process was iterated for a total of 200 generations. These settings were chosen based on prior benchmarking to ensure a balance between computational efficiency and solution robustness.

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4 Results

4.1 $H - k$ Analysis

The solution obtained from the $H-k$ analysis showed to be stable and constrained in depth for both teleseismic and local receiver functions station stacks, which resulted in good results, for the deepest discontinuities at $\sim 53-43km$ and $\sim 36-30km$. However, the k values fluctuated considerably for the local receiver functions for the layers above the shallower $\sim 10-8km$ and $\sim 1.5-1.2km$ discontinuities. In figure we present the results for the Moho for teleseismic receiver functions for stations *01A* and *05A*.

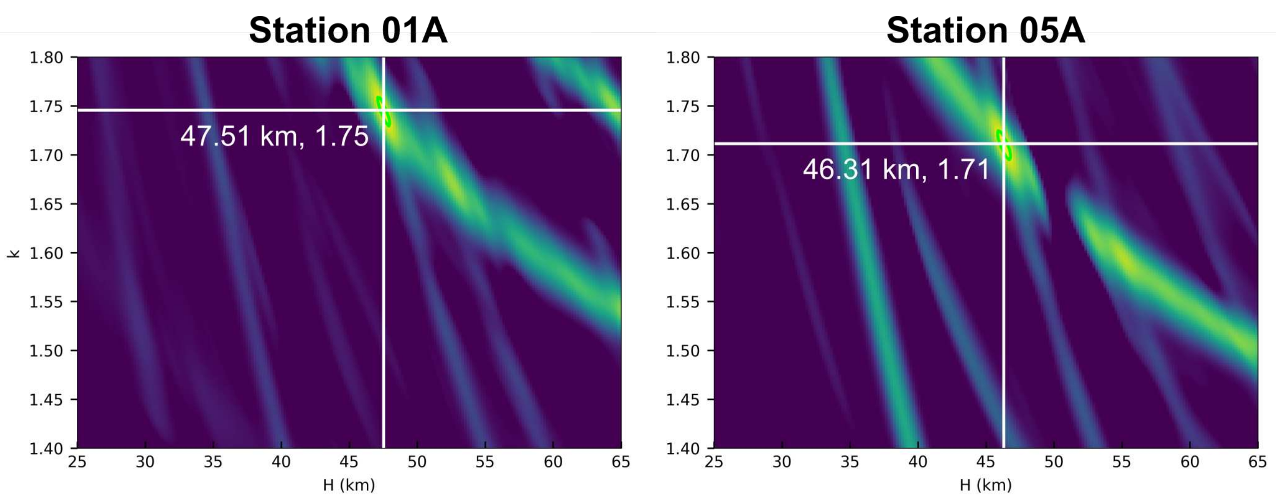

Figure 3. Sample $H-k$ stacking results for stations $01A$ and $05A$. The white lines indicate the position of the maximum value.

We see in figure 3 that the measured discontinuity, the Moho, depth and k parameters are well constrained for both cases, local and teleseismic. The v_p/v_s ratio is the one expected for the Moho's region, being ~ 1.75 and ~ 1.71 for the teleseismic and local cases, respectively.

4.2 Receiver Functions

Figure 4 shows the receiver functions for station *05A* for the *Q* and *T* components, marking the Moho conversion times for *Ps* on the *Q* component at about *5* seconds. The traces were then stacked using a binning of 15 degrees in backazimuth with an overlap of 5 degrees. In the *T* components there is a clear azimuthal conversion near 200deg, evidently more present in the *local* RFs.

Table 3 presents a comprehensive overview of all potential discontinuities for each station, extracted from the $H-k$ analysis of the stacked receiver functions, where the k values were corrected using the model by Zeckra, spanning the shallow $1-2km$ depth range up to the deeper $43-53km$ region of the Moho. Specifically, four discontinuities were identified, from the lowest

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STATION	EVENTS	$H \in [H_{min}, H_{max}]$ [km]	$k_{meas} \in [k_{min}, k_{max}]$	$k_{corr} \pm \Delta k_{corr}$
		Teleseismic RFs:		
01A	32	$1.1 \sim [1, 1.2]$	$2.00 \sim [1.8, 2]$	1.64 ± 0.47
		$47.5 \sim [46.9, 47.9]$	$1.75 \sim [1.7, 1.8]$	1.64 ± 0.08
02A	14	$8.2 \sim [6.9, 9.1]$	$1.64 \sim [1.6, 1.8]$	1.4 ± 0.05
		$31.5 \sim [28.5, 38.1]$	$1.70 \sim [1.6, 1.8]$	1.4 ± 0.06
		$31.8 \sim [31.3, 31.9]$	$1.69 \sim [1.69, 1.7]$	1.64 ± 0.07
		$48.1 \sim [47.3, 53.1]$	$1.6 \sim [1.6, 1.7]$	1.64 ± 0.08
03A	24	$1.6 \sim [1.6, 1.8]$	$1.80 \sim [1.7, 1.8]$	2.0 ± 0.56
		$27.3 \sim [26.8, 27.7]$	$1.68 \sim [1.68, 1.7]$	2.0 ± 0.08
		$45.6 \sim [45.4, 45.8]$	$1.80 \sim [1.79, 1.8]$	1.64 ± 0.08
04A	20	$3.6 \sim [3.4, 3.9]$	$1.8 \sim [1.6, 1.8]$	2.0 ± 0.49
		$31.5 \sim [30.3, 32.7]$	$1.70 \sim [1.6, 1.8]$	2.0 ± 0.08
		$42.4 \sim [41.8, 43.2]$	$1.73 \sim [1.7, 1.8]$	1.64 ± 0.08
05A	25	$1.6 \sim [1.5, 1.7]$	$1.75 \sim [1.7, 1.8]$	1.57 ± 0.44
		$27.7 \sim [27.3, 28.6]$	$1.75 \sim [1.7, 1.8]$	1.57 ± 0.06
		$46.3 \sim [45.7, 46.7]$	$1.71 \sim [1.71, 1.74]$	1.64 ± 0.08
06A	7	$9.4 \sim [9.2, 9.9]$	$1.75 \sim [1.7, 1.8]$	1.57 ± 0.06
		$29.8 \sim [29.5, 30.1]$	$1.63 \sim [1.63, 1.65]$	1.57 ± 0.06
		$44.6 \sim [44.3, 44.9]$	$1.76 \sim [1.76, 1.78]$	1.64 ± 0.08
07A	3	$3.1 \sim [3, 3.6]$	$1.80 \sim [1.6, 1.8]$	1.64 ± 0.41
		$52.6 \sim [52.3, 52.9]$	$1.80 \sim [1.8, 1.81]$	1.64 ± 0.07
08A	2	$8.1 \sim [8, 8.6]$	$1.65 \sim [1.6, 1.8]$	1.64 ± 0.06
		$30.3 \sim [28.3, 32.5]$	$1.71 \sim [1.6, 1.8]$	1.64 ± 0.07
		$49.0 \sim [48.5, 49.6]$	$1.78 \sim [1.76, 1.78]$	1.64 ± 0.08
09A	35	$1.7 \sim [1.6, 1.9]$	$1.80 \sim [1.6, 1.8]$	2.0 ± 0.55
		$6.1 \sim [5.7, 6.5]$	$1.76 \sim [1.7, 1.8]$	2.0 ± 0.31
		$26.4 \sim [26.1, 26.5]$	$1.64 \sim [1.6, 1.7]$	1.64 ± 0.06
10A	19	$8.5 \sim [7.9, 8.9]$	$1.70 \sim [1.6, 1.8]$	2.0 ± 0.07
		$27.6 \sim [27.1, 28]$	$1.63 \sim [1.6, 1.7]$	2.0 ± 0.08
		$43.1 \sim [42, 43.5]$	$1.78 \sim [1.78, 1.8]$	1.64 ± 0.08
11A	7	$0.5 \sim [0.5, 0.6]$	$1.80 \sim [1.6, 1.8]$	2.0 ± 0.59
		$35.0 \sim [34.6, 35.4]$	$1.79 \sim [1.79, 1.81]$	2.0 ± 0.09
	8	$47.0 \sim [45.6, 48.3]$	$1.75 \sim [1.7, 1.8]$	1.64 ± 0.08
		Local RFs:		
01A	65	$27.2 \sim [27, 27.4]$	$1.80 \sim [1.79, 1.8]$	1.64 ± 0.06
		$49.0 \sim [48.1, 49.9]$	$1.60 \sim [1.6, 1.7]$	1.64 ± 0.08
02A	21	$24.9 \sim [24.6, 25.1]$	$1.50 \sim [1.5, 1.51]$	1.64 ± 0.06
		$43.8 \sim [43.8, 43.9]$	$1.60 \sim [1.59, 1.6]$	1.64 ± 0.08
03A	51	$1.0 \sim [0.5, 1.4]$	$2.40 \sim [1.8, 2.5]$	1.73 ± 0.5
		$31.3 \sim [31.1, 31.6]$	$1.70 \sim [1.7, 1.8]$	1.73 ± 0.07
		$43.0 \sim [42.2, 43.3]$	$1.80 \sim [1.7, 1.8]$	1.64 ± 0.08
04A	41	$0.6 \sim [0.5, 0.6]$	$1.90 \sim [1.8, 1.9]$	1.64 ± 0.48
		$22.5 \sim [22.3, 22.5]$	$1.70 \sim [1.69, 1.7]$	1.64 ± 0.06
05A	62	$29.5 \sim [29, 29.7]$	$1.60 \sim [1.6, 1.7]$	1.64 ± 0.07
		$49.3 \sim [48.7, 50.1]$	$1.70 \sim [1.6, 1.7]$	1.64 ± 0.08
06A	24	$38.9 \sim [37.6, 40.3]$	$1.60 \sim [1.5, 1.6]$	1.64 ± 0.07
07A	8	$3.8 \sim [3.8, 4.4]$	$1.70 \sim [1.5, 1.8]$	1.64 ± 0.39
		$39.9 \sim [39.2, 40.6]$	$1.90 \sim [1.8, 1.9]$	1.64 ± 0.08
08A	6	$1.5 \sim [1.3, 1.8]$	$2.50 \sim [2.2, 2.5]$	1.64 ± 0.46
		$27.6 \sim [26.7, 28.3]$	$1.90 \sim [1.8, 1.9]$	1.64 ± 0.06
09A	178	$13.7 \sim [13.5, 13.9]$	$1.80 \sim [1.7, 1.8]$	1.64 ± 0.06
		$46.7 \sim [46.6, 47.2]$	$1.60 \sim [1.59, 1.6]$	1.64 ± 0.08
10A	53	$13.2 \sim [12.6, 14.7]$	$1.90 \sim [1.7, 2]$	1.74 ± 0.06
		$23.0 \sim [22.4, 23.9]$	$1.70 \sim [1.7, 1.8]$	1.74 ± 0.06
		$29.4 \sim [27.7, 30]$	$1.60 \sim [1.59, 1.6]$	1.64 ± 0.07
11A	15	$0.4 \sim [0.3, 0.6]$	$2.50 \sim [1.7, 2.5]$	2.0 ± 0.6
		$7.7 \sim [7.4, 8.4]$	$2.50 \sim [2.3, 2.5]$	2.0 ± 0.11
		$15.1 \sim [14.2, 15.7]$	$1.70 \sim [1.7, 1.8]$	1.73 ± 0.06
		$19.2 \sim [19.1, 19.7]$	$1.60 \sim [1.59, 1.6]$	1.64 ± 0.05

Table 3. Corrected v_p/v_s ratios at each depth and station for teleseismic and local receiver functions, with depths H and measured k and their related error. v_p/v_s ratio of 1.64 ± 0.02 used at the Moho.

to the greatest depth: at depths of $43-53$, $30-35$, $8-10$, and $1.2-1.5$ km. It is crucial to acknowledge that the majority of these discontinuities are not discernible for all stations simultaneously; rather, only the Moho Ps conversion is visible at all stations. The errors have been constrained for the deeper discontinuities, with values of less than 5% relative error for both

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parameters. However, for the shallower interval of $1.2 - 1.5$ km, the resulting k was significantly increased in relative error from 50% to 70% of the measurement.

In figure 5, the vertical variations of the v_p/v_s ratio is also distinguished. This ratio is observed to slightly increase monotonically, depending on depth from the bottom to upper discontinuity, although it is poorly contained, particularly in the uppermost layer. Although some of the teleseismic k corrected values oscillate around 1.64 between 1.4 and 2.0 for depths ~ 30 and ~ 10 km, we consider these to be under- or overestimated by at least %5 due to non-vertical incidence and predominance of events having a narrow backazimuth interval between 300° to 360°.

The migrated receiver function stacks (see Fig. 8) clearly reveal four discontinuities listed in Table 3, appearing as continuous zones. In the north-south profile *A–A'*, three of these discontinuities—located at approximately 47 km, 30 km, and 10 km depth—are observed in both local and teleseismic RFs, with sharper resolution in the local RFs. Additionally, a detachment horizon at around 15 km depth is identified exclusively in the local RFs. In both profiles, the Moho region thickens and dips southward, reaching depths exceeding 50 km.

For the east-west directed profiles *B–B'* and *C–C'*, situated in the south and north respectively, the same discontinuities are identified. However, in the northern profile, they appear more diffuse. Notably, the higher frequency content of the local RFs significantly enhances the clarity of structures in the east-west profiles.

4.3 Ambient Noise Cross-correlation

Figure 6 shows the acausal and causal parts of the ambient noise cross-correlation traces in terms of the inter-station distance, where there is a clear one-sided tendency towards positive times. This, in principle will appear to be a contraposition to the homogeneity assumption of the ambient noise wavefield on which ANT is based. However, Pedersen and Krüger(2007) showed that even in this scenario of a dominant noise direction the cross-correlations are not significantly affected (less than 10%). On the other hand, this also points to clear difference between the northern and southern sectors.

A combination of phase and group velocities was obtained from the cross-correlations. The maps, computed for periods of 10 seconds, showed similar characteristics within the above time span. However, the quality of the phase velocities proved to be more consistent for shorter periods, particularly between 1.6 and 2.2 seconds. As a group, the velocities in all cases have an unstable (sharp oscillatory) behavior in the processed periods.

For the period of 2 seconds, a weak zone of relatively slow velocities appears between the area enclosed by the triangle formed by stations 09A-07A-08A (see Fig. 7) and 12A-06A-10A. This zone is only visible in the phase velocity maps. On the other hand, the group velocity shows a zone of relatively high velocity in the line formed by stations 12A-11A-10A and a zone of relatively low velocity in the area bounded by stations 06A-05A-07A-08A-09A. For the period of 3 seconds, two distinct zones appear in both group and phase velocities: A zone of high relative velocity in the area bounded by stations 01A-02A-04A-05A-06A-03B, which will be called the northern sector, and a zone of relative low velocity between stations 11A-07A-08A-10A, which will be called the southern sector. The maps for the period of 4 seconds, show for the high velocity zone an increase in contrast and extension in the group velocity map, and a decrease in extension and contrast of the low and

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high velocity zones in the phase velocity map. Finally, for the period of 5 seconds, the two zones of high and low velocity, begin to merge into a homogeneous layer of the same velocity.

4.4 Joint Inversion

We observe that the best model derived from the joint inversion of RFs and dispersion curves reproduces the model proposed by Zeckra(2020) with the main difference that velocity of the shallowest layer are decreased to lower values and the discontinuities at $35km$ and $46km$ change slightly to increased depths.

In Figure 8, we present the inversion results for stations 01A, 05A, 07A and 10A. All four stations share similar depth and velocity characteristics, though subtle differences emerge. Notably, the upper layers in the northern region exhibit slightly higher S-wave velocities compared to those in the south, while the lower layers show consistently lower velocities across all four stations without significant variation.

The receiver function fits are reliable for the selected stations, with station 05A displaying the best fit. At this station, the model closely follows the observed data, capturing not only the shape but also the amplitude of all maxima and minima.

Moreover, the results from the evolutionary algorithm inversion (see Fig. 8) align well with those from the joint inversion in the middle layers of the upper layers, between 3 and 5 km. A primary distinction is the presence of a low-velocity layer, approximately 0.5 to 0.8 km thick, which appears at stations 01A through 07A, is absent at stations 08A and 09A, and reappears at stations 10A–11A and 12C.

The combined velocity model for all stations comprises five distinct layers: an upper sediment layer (0.8 km thick), below it a consolidated sediment layer (3.7 km), a lower consolidated sediment layer (2 km), an upper crustal layer (32 km), and a lower crustal layer (10 km). The Moho is located at a depth of 48-49 km.

5 Discussion

The distribution of local earthquakes provides essential context for interpreting the crustal discontinuities identified throughout the analysis. The earthquake catalog comprises well-located events with relatively low hypocentral uncertainties (less than 5%), broadly distributed across the study region. Event magnitudes range from 0.5 to 4.8 ML, with an average magnitude of 1.5 ML. In terms of depth, the seismicity exhibits a distinctive bimodal distribution, with two prominent peaks at approximately 15 km and 35 km—the majority of events occurring at the deeper level. Within the Lerma Valley, seismicity is sparse and lacks a clear spatial pattern. However, on the western flank of the valley, two distinct clusters—each consisting of eight events—likely delineate the rupture zone of the 2010 Salta earthquake. These clusters are characterized by NE–SW striking, full-reverse or reverse-transpressional focal mechanism solutions.

The results presented in the previous section (see Sec. 4) highlight the complexity of the crustal structure in the Lerma Valley on multiple levels. In this trend, the discontinuities identified through the $H-k$ analysis of the receiver functions (both teleseismic and local) align well with previously proposed regional crustal models, e.g. by Cahill et al. (1992). Specifically, regarding the depth of the Moho, all stations showed constrained and stable solutions at 48 ± 5 km, a feature that aligns closely

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with the findings of Zeckra (2020), who positioned the Moho depth at 46 km in the Santa Bárbara system. Similarly, the corrected v_p/v_s ratios remained in the range of 1.48 to 1.8, with a mean value of 1.65. This, too, agrees with the results of Zeckra(2020), who attributed this stable ratio to a felsic composition in the lower crust. However, this stability in v_p/v_s ratios is only apparent for the upper discontinuities in the teleseismic receiver functions, as teleseismic signals, due to their long-period frequency, are less sensitive to minor changes in layer velocities. In contrast, local receiver functions reveal a gradual increase in the v_p/v_s ratio from 1.6 to 2.5, spanning from the Moho upward. This behavior is also present in Zeckra (2020), where a v_p/v_s ratio of about 2 is measured for the second layer.

In addition to the felsic layer identified above the Moho region, a shift in azimuth in the T components of the receiver functions indicates a dip along the north–south axis, centered around 200 deg. As shown in Figure 7 for both local and teleseismic data, this feature suggests a gradual change in the Moho surface. Similar observations have been reported in New Zealand Savage (1998), where azimuthal analysis of receiver functions revealed Moho dips associated with variations in the geometry of the subducting plate.

Detachment zones play a key role in accommodating crustal shortening and deformation in orogenic systems, particularly within the Andean orogen. In the Eastern Cordillera, these zones are commonly associated with mid-crustal decoupling, where strain is partitioned between upper and lower crustal levels, often facilitated by the presence of weak layers or fluids (Grier et al., 1991). Such detachment structures have been invoked to explain the style and distribution of deformation in the Eastern Andes, where thick-skinned tectonics transitions to more complex, distributed strain at depth (Kley and Monaldi, 2002; Pearson et al., 2013).

Continuing with the middle crust discontinuities, the one at an average depth of 30 km, well defined in teleseismic receiver functions but more dispersed in local receiver functions, likely represents the mid-lower crustal boundary. This finding is consistent with the velocity model of Zeckra(2020). Further, a discontinuity at 15 km, exclusive to local receiver functions, likely marks the upper detachment horizon with an extensive fracture network occupying the middle crust. Notably, similar features have been proposed on different scales by Grier et al. (1991), Pearson et al. (2013), and Kley and Monaldi (2002). This distinction by the local receiver functions is due to the high-frequency content in the spectra of Zapla cluster events (Mulcahy et al. (2014); Valenzuela Malebran (2022)), which effectively detect the fracture zone, despite uniform rock composition.

Within the upper crust discontinuities, a boundary at 8 km depth appears prominently in teleseismic receiver functions and at one station in local functions, while additional discontinuities between 5 and 1 km depth are evident in both RF types. The former indicates a significant change in rheology, distinguishable by long-period signals and likely defining the upper crustal boundary with sediment layers. Meanwhile, the latter depth marks the basement of the basin represented by the Puncoviscana Formation (see Section 2), which is overlain by the Santa Victoria and Mesón Groups, the Salta Group, and more recent Orán Group and Quaternary sediments. According to ambient noise cross-correlation tomography, two zones—one slow ($1.75 km/s$) and the other fast ($3.5 km/s$)—are defined by marked velocity contrasts with depth. We interpret, these zones would correspond to the dense quartzites of the Santa Victoria Group, which form a high-velocity zone while the Quaternary units comprise the low-velocity zone.

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This feature directly corresponds to the differences observed between the northern and southern basin, as noted by Salfity(1985). The northern section lacks outcrops of the Salta Group, which dominate in the southern division controlled by the COT lineament (see section 2).

The migrated receiver function stacks (Fig. 8) provide compelling evidence for the presence of multiple seismic discontinuities, consistent with those listed in Table 3. These features appear as continuous zones across all profiles, supporting the interpretation of laterally coherent crustal structures. In the north–south profile *A–A'*, three prominent interfaces—located at approximately 47 km (the Moho), 30 km, and 10 km depth—are identified in both local and teleseismic receiver functions. The improved sharpness of these features in the local RFs highlights their higher resolution and sensitivity to fine-scale crustal layering, consistent with previous findings on the advantages of local RFs (Yuan et al., 2000; Ozacar and Zandt, 2008).

In this context, a distinct detachment horizon at approximately 15 km depth is observed exclusively in the local RFs. This feature may reflect mid-crustal shearing or the presence of fluids—both commonly associated with deformation and metamorphism in active orogens (Levander and Miller, 2006). These observations are consistent with interpretations of widespread mechanical decoupling and intra-crustal strain partitioning in other Andean foreland systems (Kley and Monaldi, 2002; Oncken et al., 2006).

In contrast, the Moho, evident in both local and teleseismic profiles, exhibits a clear southward-deepening trend, reaching depths greater than 50 km. This pattern may indicate crustal underplating or lithospheric flexure associated with ongoing convergence and crustal thickening (Zandt et al., 2004; Thybo, 2006). Similar Moho deepening has been reported in seismic studies across the central Andes and is often linked to magmatic additions or lower crustal flow in response to long-term tectonic loading (Beck and Zandt, 2002; Beck et al., 2015; Heit et al., 2014). These structural features are further supported by receiver function and seismic tomography results, which reveal significant heterogeneities in crustal structure tied to the evolution of the Andean orogen (Bianchi et al., 2013).

In the east–west oriented profiles *B–B'* and *C–C'*, which cross the southern and northern segments of the study area, the same discontinuities are observed. However, in the northern profile, these features appear more diffuse. This may indicate lateral heterogeneity in crustal composition or increased attenuation due to structural complexity or varying seismic properties (Ammon, 1990).

Importantly, the higher frequency content of the local receiver functions significantly enhances structural clarity in the east–west profiles, emphasizing the utility of high-resolution RF analysis for imaging crustal discontinuities. The combined use of local and teleseismic data provides a more comprehensive image of crustal architecture and reveals important spatial variations that contribute to our understanding of the geodynamic evolution of the region (Julia et al., 2000; Kind et al., 2002).

The model derived from the joint and SWD inversions closely aligns with that obtained from the receiver functions and phase velocity dispersion curve, delineating four primary boundaries at depths of 47 km, 36 km, 6 km, and 4 km. These interfaces, first identified by Zeckra (2020), correspond well with the discontinuities observed in the teleseismic receiver functions (see Subsec. 3.2.2). However, a notable discrepancy exists in both depth and shear wave velocity: our model systematically indicates lower velocities and greater depths across all discontinuities.

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It is worth noting that a preliminary inversion using a base model derived from local VELEST results was also tested. However, this approach proved unstable, producing poor fits and yielding unphysical results, including negative velocity gradients in the lower crust. Such artifacts are geologically implausible and were therefore excluded from further consideration.

In addition, it should be noted that the upper layers of the model mentioned above, down to five kilometers, include a low velocity layer of about 0.7 km/s. This feature can be attributed to the Tajamar Formation, for the southern stations (see Sec. 2). The extent of this unit will be of key importance, since it is conformed by fine-grained siltstones that are expected to suffer liquefaction when water oversaturates during strong motion produced by high magnitude events (Elías et al., 2022).

6 Conclusions

In conclusion, the results of this study provide a detailed and coherent image of the crustal structure beneath the Lerma Valley, derived from the analysis of both local and teleseismic receiver functions in conjunction with surface wave dispersion data. The observed crustal stratification is broadly consistent with previous models proposed by Zeckra (2020) and Cahill et al. (1992), particularly in the alignment of the upper layers with the known sedimentary basin structure, characterized by low velocities reaching down to $2.5\ \mathrm{km/s}$.

The structural interpretation revealed four major discontinuities at approximate depths of 53–43 km, 35–30 km, 10–8 km, and 1.5–1.2 km. These were clearly imaged in the migrated receiver function stacks and supported by the CCP analysis. The deepest discontinuity corresponds to the Moho, which exhibits a southward-dipping geometry as observed in the L-component of the teleseismic RFs. The second interface marks the transition between the lower and middle crust, while the third delineates the upper limit of a possible detachment zone. The shallowest interface defines the basement of the sedimentary basin.

Importantly, the Common Conversion Point (CCP) migration reconfirms a pronounced north–south contrast in crustal architecture. In the north–south profile (*A–A'*), the Moho and intermediate discontinuities appear sharper and better defined, particularly in the local receiver functions, with a clear deepening of the Moho towards the south—reaching depths greater than $50\ \mathrm{km}$. Additionally, a detachment zone at $\sim 15\ \mathrm{km}$ depth is only evident in the local RFs, suggesting a mid-crustal feature that may be tectonically significant, in terms of seismicity.

This north–south differentiation is further supported by the internal velocity variations observed across the valley, ranging from 1 to $3.5\ \mathrm{km/s}$. The southern sector is characterized by lower velocities, likely reflecting thicker or less consolidated sedimentary sequences, while the northern sector presents higher velocities associated with more competent crustal material.

The velocity model resulting from the joint inversion of receiver functions and Rayleigh wave phase velocities is robust and well-constrained. It comprises five distinct layers: (1) a soft upper sediment layer (0.8 km thick, $1.25\ \mathrm{km/s}$), (2) a medium-consolidated sediment layer (3.7 km, $2.83\ \mathrm{km/s}$), (3) a lower consolidated sediment layer (2 km, $3.25\ \mathrm{km/s}$), (4) a middle crustal layer (32 km, $3.9\ \mathrm{km/s}$), and (5) a lower crustal layer (10 km, $4.1\ \mathrm{km/s}$). These results provide key insights into the crustal architecture and geodynamic context of the Lerma Valley and establish a valuable reference for future seismic and tectonic investigations in the region.

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Data availability. The data sets used for the process are currently available at **?** (**?**) and Criado-Sutti et al. (2025).

Appendix A: Instability of the Correction Method in Upper Layers

The new correction method for estimating the true v_p/v_s ratio k_i in each layer is based on a weighted average relation:

$$k_{\text{meas}} = \frac{\sum_{i=1}^{n} k_i v_{p,i} H_i}{\sum_{i=1}^{n} v_{p,i} H_i}, \tag{A1}$$

where H_i is the thickness, $v_{p,i}$ is the P-wave velocity, and k_i is the true v_p/v_s ratio of the i-th layer. This formulation implies that the observed (measured) k value reflects a bulk average over all layers within the depth sensitivity of the receiver function.

To isolate the contribution from a shallow layer, such as a sedimentary unit, one can rearrange Eq. A1 to solve for k_1 in a two-layer system:

$$k_1 = \frac{k_{\text{meas}}\,(v_{p,1}H_1 + v_{p,2}H_2) - k_2 v_{p,2} H_2}{v_{p,1}H_1}. \tag{A2}$$

This expression clearly shows that when the shallow layer is thin or has low $v_{p,1}$, the denominator $v_{p,1}H_1$ becomes small. In such cases, the estimate of k_1 becomes highly sensitive to even small uncertainties in the measured k_{meas} or the assumed value of k_2.

To quantify this instability, we examine the sensitivity of k_1 to both the measured value and the physical properties of the shallow layer. First, the partial derivative of k_1 with respect to k_{meas} is:

$$\frac{\partial k_1}{\partial k_{\text{meas}}} = \frac{v_{p,1}H_1 + v_{p,2}H_2}{v_{p,1}H_1}. \tag{A3}$$

This factor becomes large when $v_{p,1}H_1$ is small, confirming that the inferred k_1 is highly sensitive to measurement noise in k_{meas} for thin upper layers.

Additionally, the sensitivity of k_1 to the denominator $v_{p,1}H_1$ itself is:

$$\frac{\partial k_1}{\partial (v_{p,1}H_1)} = \frac{k_{meas} - k_{meas}(v_{p,1}H_1 + v_{p,2}H_2) + k_2 v_{p,2} H_2}{(v_{p,1}H_1)^2}. \tag{A4}$$

This expression highlights that small uncertainties in either the P-wave velocity or thickness of the shallow layer result in quadratically amplified variations in the corrected k_1.

Moreover, shallow layers often exhibit high true k_1 values (e.g., up to 2), in contrast with more stable deeper crustal values around 1.64. Attempting to reconcile this contrast via the weighted average (Eq. A2) can easily lead to biased or anomalously low k_1 estimates unless the deeper contributions are accurately constrained. In regions with thick sedimentary cover or sharp vertical velocity contrasts, these effects are particularly pronounced.

In summary, the method is inherently less stable for thin upper layers, and results derived from such corrections must be interpreted with caution. Explicit sensitivity analyses, such as those above, are recommended when applying the method to shallow, low-velocity strata.

Author contributions. Criado-Sutti, E., Olivar-Castaño, A., Krüger, F., Montero-López, C., Aranda-Viana, G., and Zeckra, M. In this study, I performed with the assistance of AOC the receiver function (RF) analysis and ambient noise tomography (ANT) analysis using modified scripts originally written by AOC for his PhD thesis. I then carried out the inversion using adapted C and Python scripts. All authors contributed to the review and editing of the manuscript.

Competing interests. We hereby declare there was no conflict of interest

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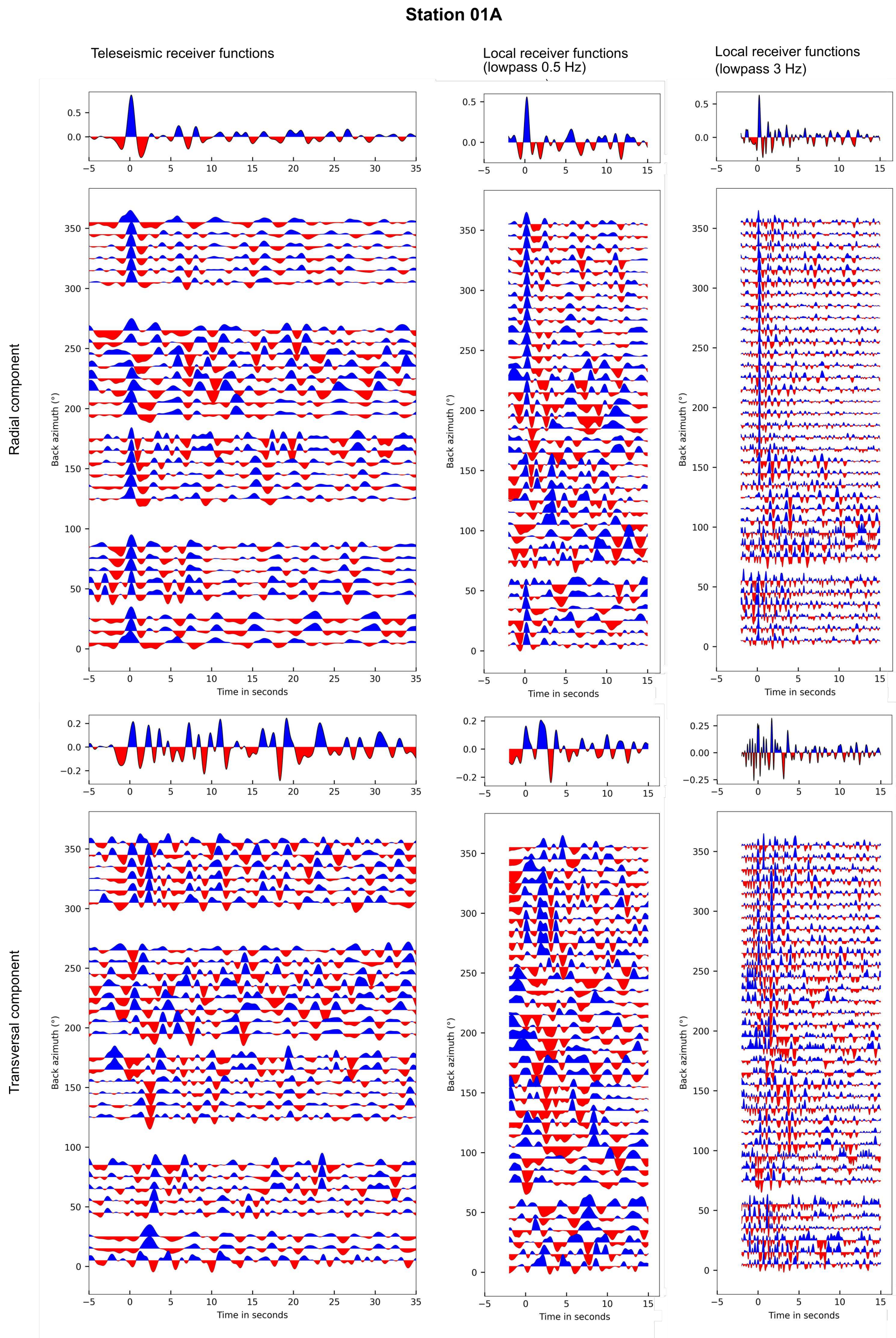

Figure 4. Teleseismic and local receiver functions computed for station $01A$. The individual receiver functions are binned in 10° intervals, with an overlap of 5°. The linear stack is represented on top of each panel. The first row shows the radial component, while the bottom row shows the transverse component.

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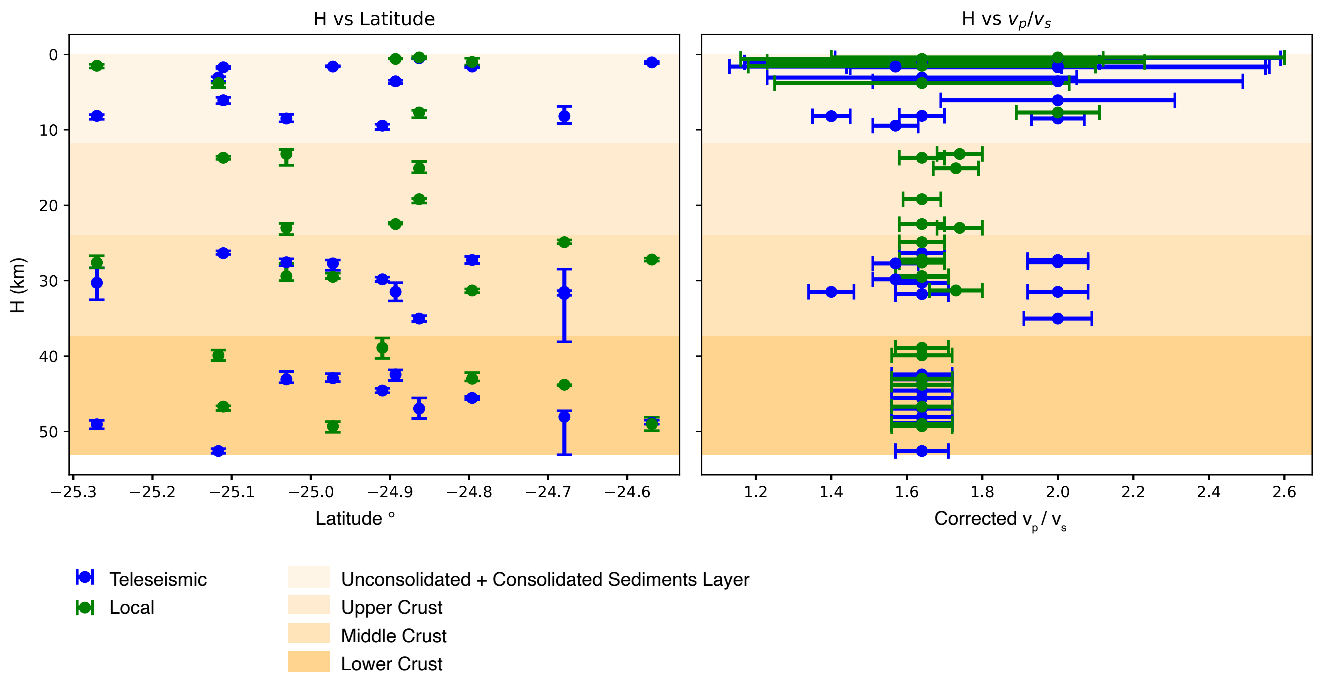

Figure 5. Corrected k values and discontinuities as a function of latitude for local and teleseismic events.

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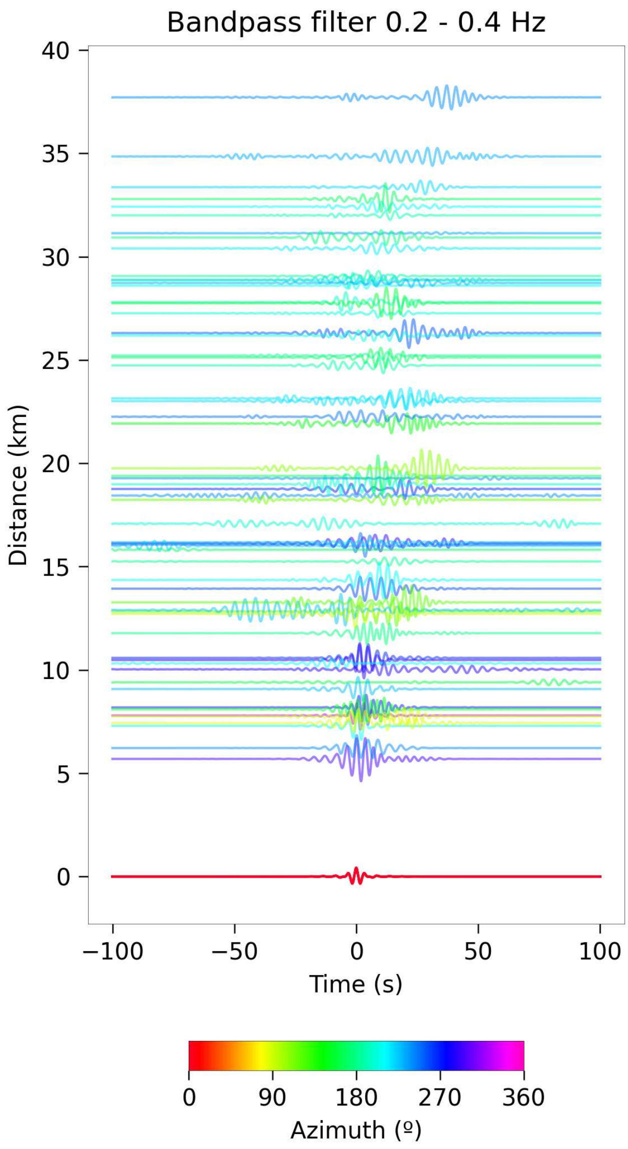

Figure 6. Acausal and causal parts of the cross-correlation for each station pair in terms of the inter-station distance band-passed with corner frequencies $0.2 - 0.4Hz$.

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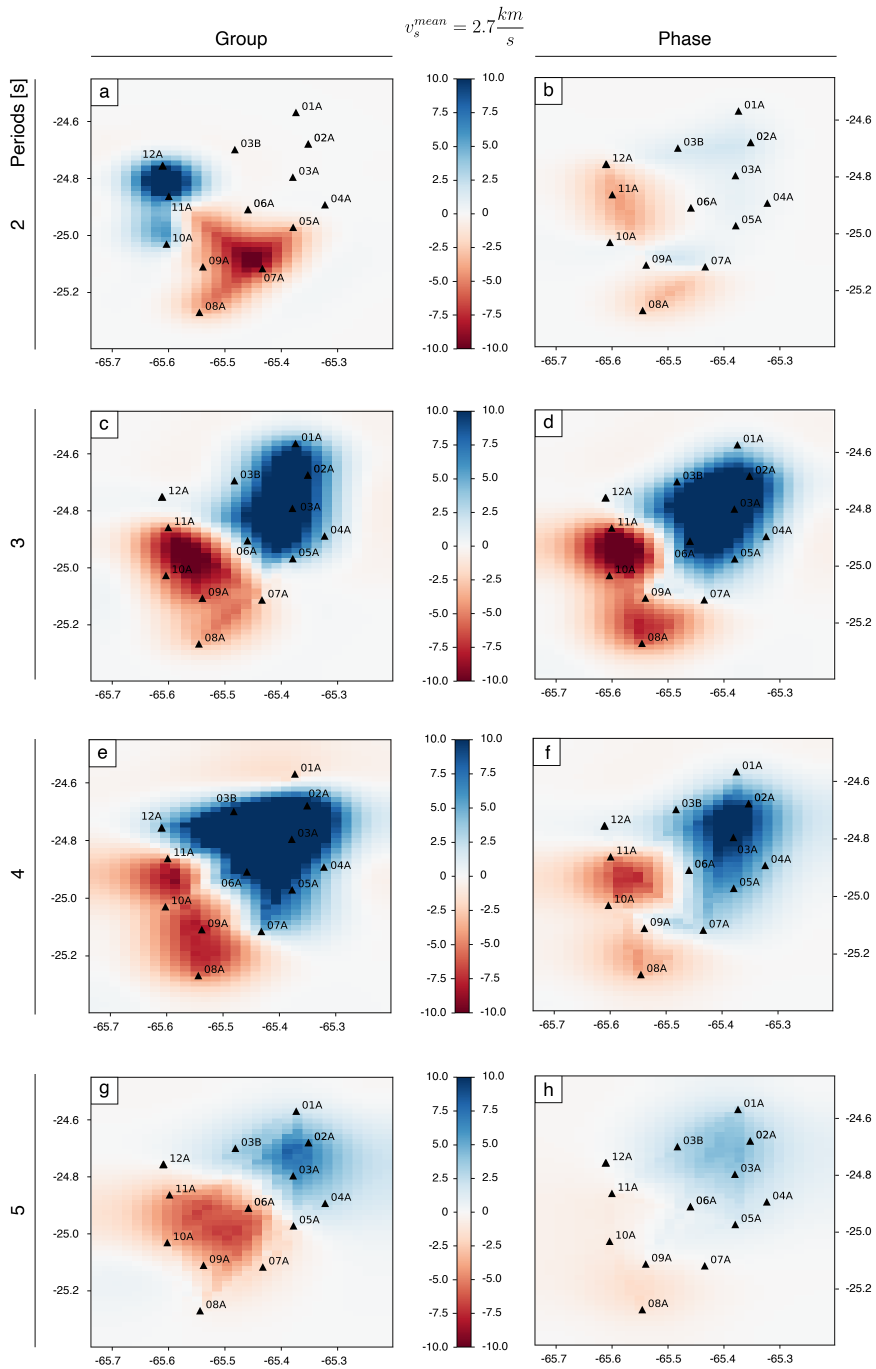

Figure 7. Group and phase velocity maps as function of period ranging from 2 to 5 seconds with station locations.

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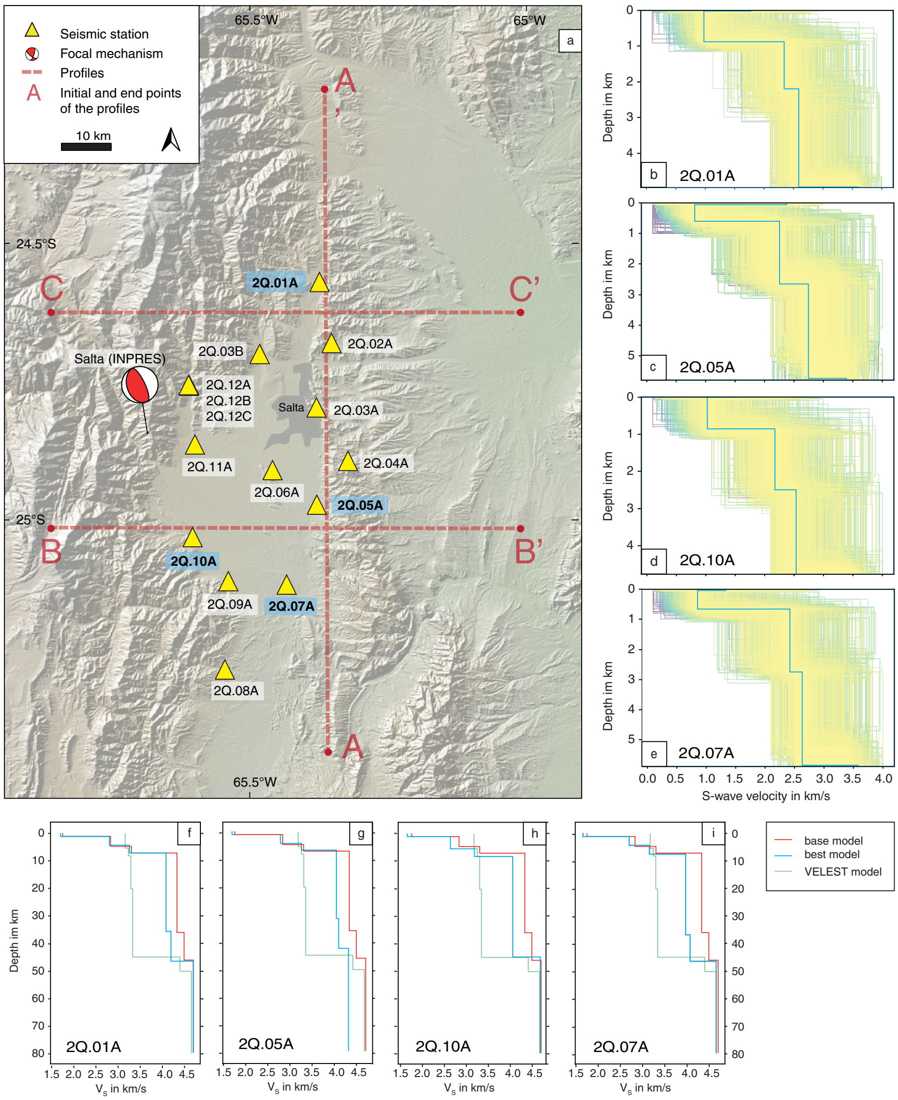

Figure 8. Base model (Zeckra (2020))(in red), VELEST model (in green) and best model inversion (in blue) for four representative stations; 01A, 05A, 07A and 10A namely.

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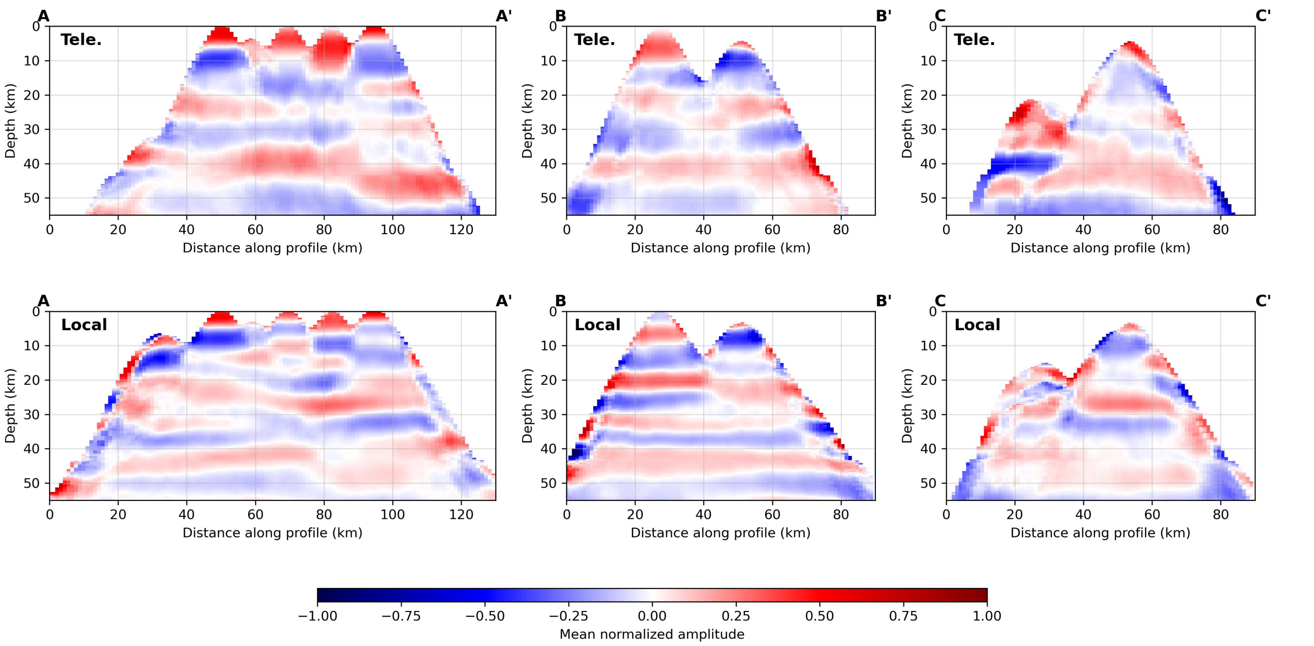

Figure 9. Pseudo-migrated sections of teleseismic (Tele.) and local (Local) receiver functions using the CCP stacking technique. The locations of the cross sections are shown in Fig. 8.