A new approach to obtaining a 3D shear wave velocity model of the crust and upper mantle: An application to eastern Turkey

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1 **0. Abstract**

2 We present a new approach to the joint inversion of surface wave dispersion
3 data and receiver functions by utilizing Common Conversion Point (CCP) stacking to
4 reconcile the different sampling domains of the two datasets. Utilizing CCP stacking
5 allows us to suppress noise in the data by waveform stacking, and correct for
6 backazimuthal variations and complex crustal structure by mapping receiver
7 functions back to their theoretical location. When applied to eastern Turkey, this
8 approach leads to a higher resolution image of the subsurface and clearly delineates
9 different tectonic features in eastern Turkey that were not apparent using other
10 approaches. We observe that the slow seismic velocities near the Karlioiva Triple
11 Junction correlate to moderate strain rates and high heat flow, which leads to a
12 rheologically weak crust that has allowed for the upward propagation of Miocene
13 and younger volcanics near the triple junction. We find seismically fast, presumably
14 rigid blocks located in the southeastern Anatolian Plate and Arabian Plate are
15 separated by a band of low shear wave velocities that correspond to the East
16 Anatolian Fault Zone, which is consistent with the presence of fluids in the fault
17 zone. We observe that the Arabian Plate has underthrust the Eurasian Plate as far as
18 the northern boundary of the Bitlis Massif, which can explain the high exhumation
19 rates in the Bitlis Massif as a result of slab break-off of the Arabian oceanic
20 lithosphere. We also find a shallow (~33km) anomaly beneath eastern Turkey that
21 we interpret as a localized wedge of mantle that was underthrust by a crustal
22 fragment during the collision of Arabia and Eurasia. These observations are possible
23 because of the high-resolution images obtained by combining common conversion

point receiver function stacks with ambient noise dispersion data to create a data-driven three-dimensional shear wave velocity model.

Keywords: Continental lithosphere, Seismic Imaging Techniques, Eastern Mediterranean, Neotectonics

1. Introduction

Imaging how the crust and upper mantle deform in response to stresses is critical to the understanding of Earth's tectonic processes. A widely used and relatively high-resolution seismic method to do this is through receiver function analysis (Langston 1979). This method isolates P-to-S wave conversions at impedance contrasts in the Earth to recover Earth structure immediately beneath a seismic station. However, receiver function analysis suffers from an inherent non-uniqueness with respect to the absolute shear wave velocities that are responsible for the resulting receiver function profile (Ammon et al. 1990). More recently, the development of Ambient Noise Tomography (ANT) has led to the accurate recovery of short period Rayleigh waves sensitive to absolute shear wave velocities in the crust and uppermost mantle, which were previously difficult to obtain via earthquake-generated surface waves (Shapiro et al. 2005). Dispersion data obtained from surface wave inversions are widely used to recover the shear wave velocity structure of the Earth through shear wave inversions, but suffer from their own non-uniqueness, as the broad sensitivity kernels of Rayleigh waves sample a wide range of depths depending on their frequency (Fig. 1) and thus are not ideal for imaging sharp velocity discontinuities. Inverting these two datasets separately results in an inverse problem with a large number of models that satisfy the data,

which can lead to biases in velocity models due to their strong dependence on poorly constrained *a priori* models.

The joint inversion of surface wave velocities and receiver functions has resulted in a vast improvement in the calculation of shear wave velocity models by utilizing each method's individual strengths (Özalabey et al. 1997; Julia et al. 2000). Receiver functions constrain the depth to boundaries and their associated velocity contrasts, while Rayleigh wave dispersion data constrain the absolute shear wave velocities between the boundaries. By utilizing both receiver functions and high frequency surface wave dispersion data, many studies have been successful in gaining insight into crustal structure at a resolution unprecedented before the development of this joint technique (Shen et al. 2013b, Ward et al. 2014). However, the sampling regions and resolutions of these two datasets are vastly different (Fig. 1), and must be reconciled before presenting models of shear wave velocity via this technique.

Past studies have attempted to reconcile the different sampling regions of these two techniques using different approaches. The most common approach is the single station joint inversion (Julia et al. 2000; Kgaswane et al. 2009). This approach uses all receiver functions recorded at a single station, accounts for differences in ray parameter, and constrains shear wave velocities using a surface wave dispersion curve from a gridpoint near the station location. This results in the approximation of a 1D shear wave velocity profile as a function of depth beneath each individual station. This approach, however, suffers when large backazimuthal variations exist beneath an individual station. Shen et al. (2013a) attempted to account for

backazimuthal variation via “harmonic stripping”, which ideally creates azimuthally independent receiver functions. If a 3D velocity model is sought, joint inversion studies generally interpolate the single-station velocity profiles between stations (Shen et al. 2013b, Liu et al. 2014), leading to a 3D velocity model that is dependent on the interpolation method as opposed to a data-driven 3D velocity model. Ward et al. (2014) attempted to improve upon the single station method by broadening their data-driven shear wave velocity recovery through a multistep inversion process: first, individual receiver functions were inverted with dispersion data and the resulting shear wave velocity profile was back-projected along the theoretical raypath of the receiver functions to account for backazimuthal variations. This resulting profile served as the starting model for a shear wave inversion of dispersion data, followed by an interpolation where there was no ray coverage. This method is an improvement in creating a 3D volume of shear wave velocities using a joint inversion approach, but may suffer from the inversion of contaminating noise in individual receiver functions, which may lead to spurious velocity information that is later propagated through the model via mathematical interpolation. Conversely, Chai et al. (2015) smoothed receiver function waveforms over large distances to obtain a low-noise receiver function containing information about first order discontinuities. This leads to a good first-order model at the cost of local heterogeneities, and thus resolution.

In this paper, we present a new approach to the joint inversion problem to develop a more robust 3D shear wave velocity model. We utilize common conversion point (CCP) stacking (Dueker & Sheehan 1997), which is widely used to

create 3D volumes of receiver function amplitude as a function of depth to gain insight into impedance contrasts in the Earth. By using a depth-to-time migration on the resulting 1D amplitude profiles created by CCP stacking, we can create receiver functions that account for backazimuthal variations, dampen noise, mitigate the dependence of receiver function data on station location, and lead to a high-resolution data-driven 3D shear wave velocity model when jointly inverted with dispersion data.

2. Methods: The Creation of CCP-derived Receiver Functions

CCP stacking creates a 3D amplitude volume throughout a study area by averaging receiver functions that fall in a grid cell (or bin) after being ray-traced along theoretical raypaths assuming an average velocity model for an area. The average amplitudes in the volumes represent the location of discontinuities in the crust and mantle, successfully accounting for backazimuthal variations beneath a station, albeit rather smoothly. Commonly in the CCP method, the grid spacing and information in individual bins is user-defined by the radius of the bin and the bin spacing, which constrains the data that is used in the solution for amplitude in that bin.

In order to extract a receiver function for each grid-point from our CCP stacks, we must have a continuous amplitude profile as a function of depth. Due to the localization of raypaths beneath individual stations at shallow depths, bins are often empty between stations if the bin width is less than station spacing. In an attempt to alleviate this issue and create a more continuous image of the shallow crust, we allow our predefined bin width to dilate until a minimum number of

117 raypaths is incorporated into the bin. We use true receiver function amplitudes (i.e.
 118 non-normalized), migrate the receiver functions to depth for the CCP stacking
 119 analysis using a regional 1D velocity model, and then extract a vertical receiver
 120 function amplitude profile for each grid-point as a function of depth. Then, a depth-
 121 to-time migration is performed using the average ray parameter in the uppermost
 122 bin and the same 1D regional velocity model used for ray-tracing and time-to-depth
 123 migration in order to minimize the effect that an incorrect velocity model might
 124 have on the resulting CCP-derived receiver functions. To avoid aliasing in the CCP-
 125 derived receiver functions, we must choose thin CCP depth bins so that the time
 126 spent in each bin is less than the sampling rate in our original receiver function.
 127 Thus, for an input receiver function, we can use the following equation to find the
 128 maximum thickness for a CCP depth bin that, when migrated with a certain velocity
 129 model, will produced data points that sample at a desired sampling rate:

$$dz = \frac{dt}{\sqrt{\frac{1}{\min(V_s)^2} - \max(p)^2} - \sqrt{\frac{1}{V_p^2} - \max(p)^2}} \quad (1)$$

130 where V_p is the P-wave velocity corresponding to the minimum S-wave velocity
 131 layer (V_s) in the CCP profile, dt is the desired sampling rate of the CCP-derived
 132 receiver function, and p is the average ray parameter in the uppermost bin of that
 133 CCP profile. Choosing a depth bin thickness less than dz will ensure the signal is not
 134 aliased. The depth-to-time migration does not produce constant sampling rates if we
 135 migrate with a non-constant velocity model, and thus we must downsample our
 136 resulting CCP-derived receiver function to a desired frequency, producing an

equivalent, yet CCP-derived receiver function similar to our input receiver functions. This procedure is outlined in Fig. 2, and results in a single stacked receiver function for each gridpoint that can later be paired with dispersion data for the joint inversion following the method described by Julia et al. (2000).

3. Application to real data: Eastern Turkey Seismic Experiment

The Eastern Turkey Seismic Experiment (ETSE) consists of 29 stations deployed from 1999-2001 with the goal of investigating the crustal and upper mantle structure beneath the East Anatolian Plateau (EAP) (Fig. 3; Sandvol et al. 2003a). This region was initially thought to be a small-scale Himalayan-type orogen due to the broad deformation in the plateau and characteristic tectonic escape of the Anatolian Plate to the west along the North and East Anatolian Faults (NAF and EAF) as a result of the collision of the Arabian plate with Eurasia (Şengör & Kidd 1979). However, notable differences between this region and the Himalayan-Tibet system are apparent as shown by research done with the ETSE seismic data (Al-Lazki et al. 2003; Zor et al. 2003; Sandvol et al. 2003b; Gök et al 2003; Gök et al. 2007; Zor 2008; Ozacar et al. 2008; Özacar et al. 2010).

The evolution of what is now eastern Turkey is rather enigmatic before ~15 Ma due to extensive volcanic cover and the complexity of the termination of the Tethyan system, and many models for the tectonic history of eastern Turkey have been proposed (see Keskin 2007 for review; Şengör et al. 2008; Oberhänsli et al. 2012). Today, the EAP is generally characterized by high heat flow (Tezcan 1995), broad internal deformation in a crust composed largely of weak accretionary material formed during the closure of the Neotethys Ocean (Şengör et al. 2003;

Şengör et al. 2008; Özeren & Holt 2010), and Miocene to recent volcanics which young to the south and are interpreted to reflect a more asthenospheric geochemical signature as a result of rollback and detachment of the Arabian slab in the mid-late Miocene (Keskin 2003; Keskin 2007). The easternmost Izmir-Ankara-Erzincan Suture (here called the “Eastern Pontide Suture”, EPS, Fig. 3) separates the accretionary material of the EAP from the igneous rocks of the eastern Pontides in the north, while the crystalline rocks of the Bitlis Massif bound its southern extent. Internal deformation in the EAP is largely constrained between the eastern Pontides and Bitlis Massif. The Anatolian Plate is separated from the Eurasian Plate along the NAF and EAF, which converge at the Karliova Triple Junction (KTJ), where some of the highest strain rates in eastern Turkey are found (Özeren & Holt 2010).

Because of the large number of seismological studies in eastern Turkey as a result of the Eastern Turkey Seismic Experiment, it is an excellent test area for this new joint inversion approach. Our goal is to compare the results of both the single-station joint inversion technique and our CCP-derived receiver function technique and evaluate their spatial resolution. A total of 224 events from 29 stations with event magnitudes greater than 5.5 were used, leading to a total of 1929 receiver functions (Supplemental Section S1). We produced receiver functions with a 2.5 Gaussian pulse width, resulting in a vertical resolution of ~ 1 km using the iterative deconvolution method as described by Ligorria & Ammon (1999). Rayleigh wave dispersion data from ambient noise tomography (Delph et al. 2015) are paired with these receiver functions to constrain absolute shear wave velocity in the joint inversion.

3.1 CCP-Derived Receiver Function Parameters

We first produce CCP stacks of 10 samples per second (sps) non-normalized receiver functions across the region with a dilating bin width from 0.3 to 1.2 degrees until there is a minimum of 10 rays in each bin. We use a CCP bin spacing of 0.1 degrees so that we obtain a CCP-derived receiver function at each grid-point where we have dispersion data. The receiver function data are migrated to depth using a layer over halfspace velocity model consisting of a 3.4 km/s crust down to 44 km and 4.2 km/s mantle (Delph et al. 2015) with a constant V_p/V_s of 1.78. Based on our velocity model, the largest vertical slowness is defined by our slowest seismic velocity (0.16 s/km or $1/3.4 \text{ km/s}$) and highest ray parameter ($\sim 0.08 \text{ s/km}$). Using Eqn. (1), the maximum thickness for our depth bin in our CCP calculation can be 0.72 km to avoid aliasing the output of a 10sps receiver function. We choose a depth bin thickness of 0.5 km for ease of computation. The resulting CCP volume is similar to that obtained by Özacar et al. (2010), with small differences arising due to the varying bin width and velocity model used for migration. We extract depth-domain receiver functions from the CCP stacks and perform a depth-to-time migration using the same velocity model used in the CCP stacking analysis. These CCP-derived receiver functions are then resampled to a constant sampling rate of 10sps.

3.2 Joint Inversion Parameters

We use the same inversion parameters for the joint inversion of both the CCP-derived receiver functions and the single-station receiver functions so that any differences that arise between the models will be due strictly to the different approaches. For the single-station joint inversion, we pair the receiver functions

with dispersion data at the nearest gridpoint to the station, resulting in a single shear wave velocity profile at each station. Due to the ~ 1 km resolution of 2.5 Gaussian receiver functions, our initial models for the joint inversion consist of a halfspace of 4.5 km/s with 1 km thick layers. Our constant velocity starting model alleviates biases that *a priori* models may introduce, thus all “structure” in the resulting profiles is derived from the data itself. We perform the inversion a maximum of 40 times or until we reach a minima defined by a model change of less than 0.05% from the previous inversion iteration. The joint inversion method described by Julia et al. (2000) requires a dataset weighting parameter that ranges from 0 to 1, with 0 performing a receiver function only inversion and 1 representing an inversion of dispersion data only. We favor a heavier weight on receiver function fit in order to take full advantage of the information gained by incorporating receiver functions into the inversion. We choose a damping parameter of 0.5 and weighting parameter of 0.3 (70% Receiver function, 30% Dispersion data), which helps prevent the inversion from fitting noise that may be present in the receiver functions. Regardless of weighting parameter, the fit to the dispersion data remains relatively constant (Fig. 4). The normalized RMS fit of the inverted receiver functions and dispersion information to the observed data is shown in Fig. 5.

As seen in Eqn. (1), we only focus on Ps conversions and attempt to ignore strong seismic multiples that may be present in the receiver functions. By only inverting the first 10 seconds after the direct P arrival, we avoid inverting the high-amplitude multiples associated with the Moho. We also inspect each receiver function for evidence of shallow structural complexity, which will lead to a delayed

receiver function amplitude peak at $t > 0$ and will introduce multiples into our time window. Thus, stations with significantly delayed peak amplitude times (e.g. station HINS, peak at $t \sim 0.4$ s) were discarded. Multiples from mid-crustal discontinuities generally arrival in the 6 – 10 second time window and will effect shear wave velocities at depths > 50 km, where the sensitivity of the dispersion data used in this study is low (Fig. 1B). Thus, to avoid ambiguities created by crustal multiples and decreased dispersion sensitivities, we show only the top 50 km of our model.

4. Comparison of Results and Discussion

In Fig. 6 and 7, we show the comparisons between past techniques used to investigate crustal structure, such as CCP stacking and shear wave inversions of dispersion data only, with the joint inversion interpolation and CCP-derived receiver function approaches. In comparing these different techniques and approaches, we see that the main features of the models are recovered rather consistently, but the smoothness of the models decreases as shown in Fig. 6. This results from a lack of receiver function input in the shear wave inversion of dispersion data only, the averaging of receiver functions to produce one shear wave velocity profile at each station, and the incorporation of CCP-derived receiver functions. As we would expect from the inversion of the CCP-derived receiver functions, many of the features seen in the CCP stacks propagate their structure into the resulting joint inversion results. As we consider CCP stacking analysis as a robust way to interpret impedance contrasts within the Earth, this new approach allows us to gain a higher resolution image of structures and associated shear wave velocities in our region.

As seen consistently in past studies, the Arabian lithosphere shows up as the most prominent fast velocity anomaly at depth (Fig. 7D; Warren et al. 2013; Delph et al. 2015). The typical depth to a shear wave velocity of 4.2 km/s, which we use as a proxy for crustal thickness in the Arabian Plate, is ~34 km. These results are in general agreement with the H-k results of Özacar et al. (2010) for the Arabian Plate. This method also clearly delineates the abrupt lateral termination of the Arabian lithosphere's seismic signature near the Bitlis-Zagros suture in the 38 km depth slice (Fig. 7D) as opposed to the interpolation procedure (Fig. 7C).

A prominent seismic velocity contrast is also observed across the EPS between the eastern Pontides and the East Anatolian Plateau (Fig. 6D), consistent with the findings of Delph et al. (2015). By incorporating receiver functions into the inversion, we observe bands of low shear-wave velocities in the lower crust beneath the Miocene to recent volcanics not apparent via the other techniques (Fig 6D, 7B). Seismic velocities in the crust are mainly affected by anisotropy, fluids, and compositions. These velocities (<3.2 km/s) are too low for most rock compositions expected at mid-to-lower crustal depths (Christensen 1996), and the backazimuthal distribution of our station-event pairs is too poor to constrain crustal anisotropy. However, the CCP stacking method may help diminish the effects of anisotropy through stacking raypaths of different backazimuths in the same bin. Hence, assuming isotropy, while acknowledging that our shear wave velocities may be effected by anisotropy that we cannot characterize in this study, the low velocities, high heat flow, low resistivity, and high attenuation in this region (Tezcan 1995; Gök et al. 2003; Türkoğlu et al. 2008; Pasyanos et al. 2009) could indicate the presence

of fluids (most likely melt) in the lower crust of the Eurasian Plate related to the young volcanism, as proposed by Gök et al. (2007). By modifying the Raymer-Hunt-Gardner relationship between P-wave velocity and porosity, Dvorkin (2008) presented a relationship between shear wave velocity and porosity, or in this case, melt percentage. Using this equation, we calculate a melt percentage roughly between 5% and 8% for the lowest shear wave velocity (3.07 km/s) observed in the northern EAP. These estimates are in good agreement with the 3% - 10% melt percentage proposed by Türkoğlu et al. (2008) based on magnetotelluric data, which is very sensitive to the presence of fluids. Conversely, using the relationship for shear wave velocity as a function of melt percentage from the study of Hammond & Humphreys (2000), we obtain a melt percentage between ~1.5 %– 2% (Fig. 8, see Supplemental Material Section S2 for details).

Near the Karliova Triple Junction, where the North Anatolian and East Anatolian Faults merge, broad and moderate strain rates (Özeren & Holt 2010) correlate with slow lower crustal shear wave velocities (W1, Fig 7B,D). These strain rates, accompanied by high ($\sim 90 \text{ mW/m}^2$) heat flow (Tezcan 1995) are indicative of a weak lower crust. This broad region of strain may have allowed for the upward propagation of melt generated as a result of hot asthenosphere at shallow depths, leading to the observed correlation between extensive Miocene to recent volcanics seen at the surface and slow shear wave velocities at depth. Also, regions of low strain-rate in the southeastern Anatolian Plate and Arabian Plate correlate with fast seismic velocities at 38 km (R, Fig. 7D). The high shear wave velocities, low strain-rates, and lack of abundant earthquakes in this tectonically active zone (Fig. 6A)

indicate the presence of strong, rigid lithosphere. Relatively low shear wave velocities separate the two rigid blocks (Fig. 6B; W2, Fig. 7D) and spatially correlate with the young (~ 5 Ma) and broad (~ 20 km) East Anatolian Fault Zone (EAFZ; Bulut et al. 2012). A recent high-resolution magnetotelluric line across the EAFZ imaged a vertical high conductance anomaly located beneath the trace of the EAFZ starting ~ 10 km and extending to the bottom of their model (25 km), interpreted to be a highly permeable fault zone where fluids are present (Türkoğlu et al. 2015). We propose that this decrease in seismic velocities is due to a rheological weakening of the lower crust due to the fluids seen by Türkoğlu et al. (2015).

Our results show that the Arabian continental lithosphere appears to underthrust as far as the northernmost boundary of the eastern Bitlis Massif (Fig. 6C). While we do not have the depth resolution to image exactly how far north the Arabian lithosphere underthrusts the Eurasian Plate, high Sn attenuation (Gök et al. 2003) and low Pn velocities (Al-Lazki et al. 2003) are indicative of a shallow lithosphere-asthenosphere boundary beneath most of the EAP. The lack of earthquakes associated with the Bitlis-Zagros suture in this area and similar lateral displacement velocities from geodetic (GPS) observations in the northern Arabian Plate and near the northern edge of the Bitlis Massif (Reilinger et al. 2006) indicate that these systems are acting as a block, and may be responsible for transmitting stresses from the Arabian-Eurasian collision further into the plateau. Further evidence supporting the possibility that the Arabian lithosphere underlies the Bitlis Massif comes from apatite fission track dating, which shows the highest exhumation rates in the Bitlis Massif (0.5 – 0.6 km/My) occurring from 12 – 8 Ma

(Okay et al. 2010). This timing coincides with the break-off of the Arabian plate's oceanic lithosphere ~13 – 8 Ma (Şengör et al. 2003; Keskin 2003, 2007) and the presumed rebound of the Arabian continental lithosphere. These exhumation rates are consistent with the topographic responses of slab break-off (Duretz et al. 2011). In the western portion of the study area, it is difficult to image the relationship between the Arabian lithosphere and Anatolia due to deformation from the East Anatolian Fault.

Consistent with Özacar et al. (2010), we find an anomalously high-amplitude, positive-negative converter beneath the Lake Van region (near station AHLT, Fig. 3) at about ~33 km and 40 km (Fig. 6C). This anomaly was interpreted to be a lower-crustal conversion as opposed to a Moho conversion by Özacar et al. (2010).

However, with the joint inversion approach, we can determine the absolute shear wave velocity at these depths. We find a relatively high shear wave velocity (~4.2 km/s at 33 km, Fig. 6C) consistent with the velocity of the upper mantle beneath eastern Turkey (Delph et al. 2015), which suggests this anomaly may represent the Moho discontinuity. These relatively fast shear wave velocities continue to a depth of ~40 km, where there is a large decrease in seismic velocity to ~3.5 km/s, consistent with the shear wave velocities in the lower crust of our study area. We cannot unambiguously constrain the origin of such a structure, but we speculate that it may represent a locally uplifted wedge of mantle material underthrust by a thin wedge of continental crust formed during the collision of the Arabian and Eurasian Plates prior to break-off of the Arabian oceanic lithosphere. A similar seismic structure has been interpreted in the analog of the Adriatic – European

collision zone beneath the Alps (Schmid & Kissling 2000, plate 1), and underthrusting of lower crust in a collision zone is possible as shown by geodynamic modeling (Beaumont et al. 1996, fig. 2).

5. Conclusions

We present a new approach for creating a data-driven three-dimensional shear wave velocity model using the joint inversion of P-wave receiver functions and Rayleigh-wave dispersion measurements. We mitigate the different spatial sensitivities of the two techniques by computing CCP stacks to reduce the dependence of receiver function data on station location by placing more emphasis on ray path coverage. We then generate receiver functions at each gridpoint where we have dispersion data by performing a depth-to-time migration on vertically continuous profiles in the CCP stack. This resulting receiver function decreases noise through stacking and can be paired with a dispersion curve, allowing for the joint inversion for each gridpoint where both receiver function and dispersion data exist. This approach gives us the ability to resolve shear wave velocity anomalies at finer detail than previous joint inversion studies.

Consistent with past studies, we find slow shear wave velocities in the lower crust that correlate with surficial deposits of Miocene to recent volcanism. A broad region of moderately high strain-rate at the surface near the Karlioiva Triple Junction correlates with slow shear wave velocities in the lower crust. The strain rate and high heat flow in the region indicate a weak lower crust that may have allowed for the upward propagation of young volcanics in eastern Anatolia. We also observe a low shear wave velocity band paralleling the EAF that separates fast velocity

anomalies on either side. This slow velocity band correlates with a region of moderate strain rate and is consistent with fluids in a permeable fault zone, whereas the fast anomalies correspond to regions of negligible strain rate interpreted to represent either rigid and/or coherent lithosphere.

We image the Arabian lithosphere underthrusting the Eurasian Plate to the northern edge of the Bitlis Massif. An absence of seismicity and nearly constant GPS velocities across the Bitlis-Zagros Suture indicate that the eastern Bitlis Massif and Arabian Plate have coalesced, and transmit stresses further into the Eurasian Plate. The timing of high exhumation rates in the Bitlis Massif correlates well with the break-off of the Arabian oceanic lithosphere, and the rates are consistent with topographic responses to slab break-off from geodynamic modeling.

We observe an anomalously shallow, local converter (~ 33 km) located near Lake Van. When inverted with Rayleigh-wave phase velocities, we find that shear wave velocities below this converter reach 4.2 km/s and extend down to ~ 40 km depth, which we interpret to be mantle material. This is underlain by shear wave velocities of ~ 3.5 km/s, consistent with the shear wave velocities of lower crustal material in this study area. We propose this layered lithospheric structure developed as a result of collisional processes between the Eurasian and Arabian plates by the underthrusting of lower crustal material beneath mantle, in a structure similar to what has been imaged in the Alps.

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Figure 1. (A) Sampling regions of receiver functions (dark gray cones) and Rayleigh waves (light gray background). Rayleigh waves sample the entire region, while receiver functions are limited by station distribution. Black, circular arrows show conceptual particle motion of a Rayleigh wave as it propagates to the right. (B) Sensitivity kernels for Rayleigh waves at different periods for a simple 1D shear wave velocity model (black line). (C) Radial receiver function response to a velocity contrast at depth (shown by black line representing velocity structure).

Figure 2. Flow chart for the new approach to the joint inversion of surface wave dispersion data and receiver functions.

Figure 3. A map of the Eastern Turkey Seismic Experiment (operating from 1999-2001). The East Anatolian Plateau is shaded. Blue, Bitlis Massif; Red, Miocene to recent volcanics on the East Anatolian Plateau. BZS, Bitlis-Zagros Suture; EAF, East Anatolian Fault; EPS, Eastern Pontide Suture; KTJ, Karlova Triple Junction; NAF, North Anatolian Fault.

Figure 4. Average normalized RMS fit of dispersion data (gray) and receiver function data (black) and associated standard deviation as a function of weighting parameter in the CCP-derived dataset. The dispersion fit stays high (>97% fit) regardless of weighting, but the receiver function fit degrades rapidly as less importance is given to fitting the receiver functions. 0.0: Receiver function inversion only (Dispersion fit is off the graph), 1.0: Surface wave inversion only. Note that the receiver function fit is still ~86%, even when no information from the receiver function is used.

Figure 5. Fit to **(A)** CCP-derived receiver functions and **(B)** dispersion data for a weighting of 0.3.

Figure 6. (A) Station map with earthquakes $M > 3$ from the ISC catalog. Purple triangles, broadband stations; red triangles, Holocene volcanos; black dots, earthquakes. Black lines are cross sections, and white dots along line are increments of 100km. (B,C,D) Cross sections lines with 1:1 horizontal to vertical ratio. Elevation exaggerated 5x. **1)** CCP stacks from which we derived our receiver functions for each profile. Red indicates an increase in velocity with depth and blue indicates a decrease in velocity with depth across a boundary. Black line is interpreted Moho conversion. **2)** Shear wave velocity inversion of dispersion data only (Delph et al. 2015). **3)** Single station joint inversion collapsing all receiver functions at a station onto a single profile (white vertical lines). Interstation shear wave velocities derived using a linear interpolation. White, horizontal line is interpreted Moho conversion depth from CCP stack. **4)** Joint inversion of CCP-derived receiver functions. White, horizontal line is interpreted Moho conversion depth from CCP stack.

Figure 7: Depth slices through the shear wave velocity models derived from the single station joint inversion with interpolation (A,C) and the CCP-derived joint inversion (B,D). At 33 km, there is a correlation between slow shear wave velocities and Miocene to recent volcanic deposits (outlined in red). (D) At 38 km, the lateral seismic signature of the Arabian lithosphere abruptly terminates along the Bitlis-Zagros suture in the CCP model, but the interpolation model does not resolve this feature. Region inside the white outline has strain rates $> 1 \times 10^{-15} \text{ s}^{-1}$ with rates as high as $4 \times 10^{-15} \text{ s}^{-1}$ (Özeren & Holt 2010).

Figure 8: Shear wave velocity as a function of porosity based on the Raymer-Hunt-Gardner relationship (Dvorkin 2008) and the finite element analysis of Hammond & Humphreys (2000). For details of calculation, see Supplementary Material Section S2. **Orange:** Assumes V_s for a dry, non-porous rock is 3.5 km/s (near the average in this study). **Blue:** Assumes granite-gneiss shear wave velocity in the mid-to-lower crust (3.6 km/s; Christensen 1996). **Green:** Assumes granite-granodiorite shear wave velocity in the mid-to-lower crust (3.7 km/s; Christensen 1996). Shaded region is melt fraction estimate from Türkoğlu et al. (2008)