

Crustal S-wave structure around the Lake Van region (eastern Turkey) from interstation Rayleigh wave phase velocity analyses

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Received: 16.05.2016 • Accepted/Published Online: 27.12.2016 • Final Version: 13.01.2017

Abstract: This study focuses on the average crustal and the upper mantle structure throughout the Lake Van region of eastern Turkey. The study aimed to investigate the structure with the fundamental mode interstation Rayleigh wave phase velocities from the local and the regional earthquakes recorded by Kandilli Observatory and Earthquake Research Institute stations. Considering back azimuth differences of each source and station path, six different broadband station pairs and 27 earthquakes were selected to determine the 1-D shear-wave velocity structures throughout the region by using an interstation method (slant stacking technique). The linearized least squares algorithm was used to obtain the 1-D shear-velocity model that best fit the observed phase velocity dispersion curve. The normalized statistical resolution matrix was calculated to measure the reliability of the solution. Inversion results revealed that the solution quality of the upper crust is weak due to the high resolution lengths. The average shear-wave velocities in the lower crust scale down to approximately 3.5 km/s. It was inferred that this low-velocity zone shown in the lower crust may be associated with widespread volcanism. Final 2-D S-wave velocity models obtained from the inversion revealed that the crust-mantle boundary is ~42 km, and shear velocities vary from 3.6 to 4.2 km/s. Furthermore, the upper mantle (~45–70 km) velocities are slower than globally suggested models (e.g., IASP91), and this is possibly related to shallow hot asthenospheric material.

Key words: Lake Van region, crustal and upper-mantle structure, Rayleigh wave, phase velocity, inversion

1. Introduction

After two more devastating earthquakes, occurring on 23 October 2011 ($M_w = 7.2$) and on 9 November 2011 ($M_w = 5.6$), that hit the city of Van (eastern Turkey) and its surroundings, this region has received the special interest of geophysicists and geologists. Owing to the continental collision between the Arabian and Eurasian Plates in eastern Turkey, this region is a remarkable natural laboratory for earth scientists. The Eastern Turkey Seismic Experiment (ETSE) project was performed by Sandvol et al. (2003b) using the earthquake data of 29 V-shaped broadband PASSCAL stations in this region. Data recorded via these stations of this project have been used and processed by many authors with different methods (Al-Lazki et al., 2003, 2004; Gök et al., 2003, 2007; Sandvol et al., 2003a; Türkelli et al., 2003; Zor et al., 2003; Angus et al., 2006; Lei and Zhao, 2007; Özacar et al., 2008; Zor, 2008; Biryol et al., 2011; Gökalp, 2012; Delph et al., 2015b) and they have supplied new approaches for the 3-D crust and mantle dynamics of the region. Al-Lazki et al. (2003) interpreted low (<7.8 km/s) Pn velocities under the Anatolian Plateau

as most likely the absence of the mantle lid, just as Al-Lazki et al. (2004) observed the same velocities at the broad-scale in eastern Turkey and surrounding regions associated with hot, unstable mantle lid zones and active volcanism starting from the Late Miocene. In addition, due to unobservable S_n velocities in eastern Turkey, Gök et al. (2003) stated that asthenospheric upwelling brings about a lack of lithospheric mantle and this may be associated with Keskin's model (2003), while Gök et al. (2007), using receiver function modeling and surface wave group velocities, found extremity S/L_g waves attenuation for the East Anatolian Plateau, which is supported by asthenospheric material.

According to S-wave receiver function analyses carried out by Angus et al. (2006), the upper-mantle structure was relatively molten in eastern Turkey and the final results were consistent with slab steepening and the break-off model of Keskin (2003). Furthermore, the upper crust zone was about 10 km and related to volcanosedimentary lines. Delph et al. (2015b), utilizing joint inversion surface waves and receiver functions, obtained that slow shear-

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wave velocities of the lower crust are observed around the Karlıova Triple Junction regarding moderate strain (Özeren and Holt, 2010) and high heat flow (Tezcan, 1995). They thought that upwards of the hot asthenosphere had been triggered to get these findings and slab break-off model.

The crustal structures of eastern Turkey were studied by numerous authors using different geophysical methods in addition to the ETSE project (Mindevalli and Mitchell, 1989; Hearn and Ni, 1994; Çınar, 1998; Maggi and Priestly, 2005; Barazangi et al., 2006; Gans et al., 2009; Mutlu and Karabulut, 2011; Ateş et al., 2012; Bakırcı et al., 2012; Vanacore et al., 2013; Tezel et al., 2013; Toker, 2014; Komut, 2014; Skobeltsyn et al., 2014; Çınar and Alkan, 2015; Delph et al., 2015a; Türkoğlu et al., 2015). Hearn and Ni (1994) obtained a Pn velocity distribution for most of the Turkish–Iranian Plateau of as low as 7.9 km/s. Using surface waveform tomography analysis, Maggi and Priestly (2005) found that the low S-velocity zone arose from recent volcanism in the upper-mantle between eastern Turkey and the Iranian Plateau. In addition to the above-mentioned previous studies related to this region, Mutlu and Karabulut (2011) computed Pn first arrivals to identify lateral velocity variations, crust–uppermost mantle transitions, and tectonic processes in Turkey. They found low velocities (<7.6 km/s) underlying volcanic regions located in the east of Central Anatolia, which was associated with a slab break-off model (Şengör et al., 2003). Moreover, these researchers suggested that the mantle lid was not entirely absent because of some higher velocities (~8.0 km/s), and based on station delays from tomographic inversion, they determined that Moho depth varies between 38 and 42 km in the Lake Van region. Bakırcı et al. (2012) studied 3-D shear-wave velocity and the uppermost mantle structures beneath Turkey from Rayleigh wave phase velocity analyses. One of the important results in this study was the detachment of the Arabian slab appearing obviously at about 70–100 km, which may cause widespread volcanism and slow velocities. Using gravity anomalies, Ateş et al. (2012) interpreted the average crustal thickness, which varies from 38 to 43 km in eastern Turkey. Vanacore et al. (2013) calculated the crustal structure of Turkey by using receiver function data from teleseismic events. They determined the Moho depth in Eastern Anatolia as ~40–55 km and a high V_p/V_s ratio (>1.85) due to the recent volcanic activity. In a similar way, Tezel et al. (2013) found the Moho thickness and shear-wave velocity structure in the upper mantle beneath Turkey from the data of 120 broadband stations. A low velocity layer was defined as the depth between 20 and 40 km as a consequence of volcanic structures. Delph et al. (2015a) interpreted the crustal thickness and plotted 3-D shear-wave velocities in the entire Anatolian Plate by

using ambient noise tomography. According to shear-wave velocity models, these researchers found that the average crustal V_s changes between 3.2 and 3.5 km/s along the Eastern Anatolian Plateau. If the cross-sections given in this study are rigorously examined, it is clearly seen that the crustal thickness is approximately 40 km and mantle S-wave velocity about 4.2 km/s around the Lake Van region.

As is understood from the aforementioned studies related to this region, the crustal and the upper mantle structures of Lake Van and its surrounding areas are still a debatable issue for geophysicists and geologists. The main aim of this study is to determine the crustal and upper mantle structures of Lake Van and its surrounding area from the inversion of interstation fundamental mode Rayleigh wave phase velocity dispersion curves. For this purpose, station pairs were selected such that the interstation paths pass through Lake Van, and observed phase velocities were obtained from the slant stacking technique described by McMechan and Yedlin (1981) as implemented with the Computer Programs in Seismology package (Herrmann, 2013).

2. Geology and tectonic setting

Eastern Anatolia mainly encompasses an active tectonic region where the Arabian Plate collides with the Eurasian Plate in a direction of N-NW motion at 18 ± 2 mm/year (Figure 1), which initially began in the Early Miocene period (12–20 Ma age) along a deformation zone called the Bitlis-Zagros fold and thrust belt, situated south of the Lake Van region (Dewey et al., 1986; Yılmaz, 1993; Gökalep, 2012; Akinci and Antonioli, 2013; Çukur et al., 2013). This active continental collision, a part of the Alpine-Himalayan mountain system, leads to both a westward escape and counter-clockwise rotation of the Anatolian Plate along the two large strike-slip faults, called the dextral North Anatolian Fault Zone (NAFZ) and sinistral East Anatolian Fault Zone (EAFZ), which are connected at the Karlıova Triple Junction (KTJ) (Gök et al., 2007). Because of the N-S directional compression in the east of the KTJ, it leads to Plio-Quaternary volcanic activity and topographic upwelling of nearly 2000 m height (Akinci and Antonioli, 2013). A vast volume of the volcanic activity in the Eastern Anatolia (~43,000 km²) was produced after regional block uplift (11–13 Ma) and this process moved from the north in the Erzurum-Kars Plateau to the south-southeast in the Muş-Nemrut-Tendürek volcanoes (Keskin, 2007). Several different geodynamic models were created to explain the collision zone between the Arabian and Eurasian Plates (Şengör and Yılmaz, 1981; Dewey et al., 1986; Barazangi et al., 2006; Keskin, 2007; Zor, 2008; Gök et al., 2011; Skobeltsyn et al., 2014) and many of these were interpreted in detail by Keskin (2007). Şengör et al. (2003) revealed

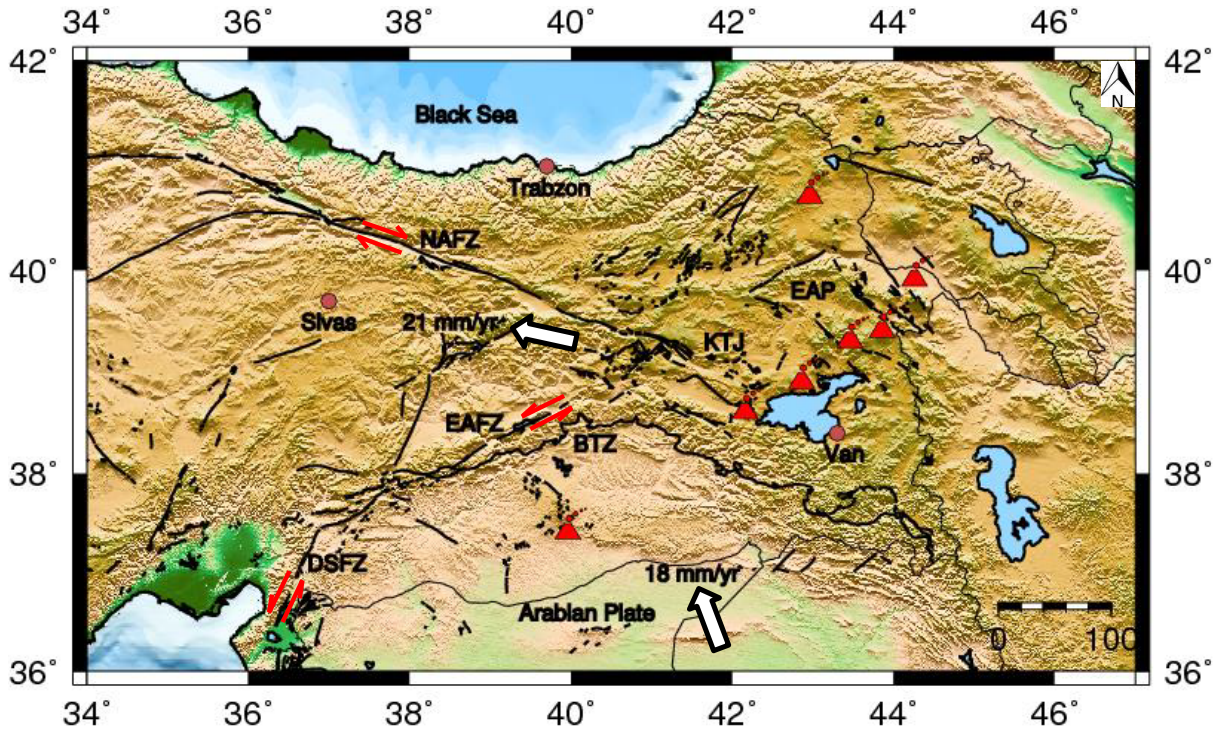


Figure 1. Tectonic map of Eastern Anatolia (modified from Delph et al. (2015a) and Utkucu et al. (2013)). Holocene volcanic centers are symbolized by red triangles and arrows denote relative plate and fault motions. DSFZ: Dead Sea Fault Zone, EAFZ: East Anatolian Fault Zone, BTZ: Bitlis Thrust Zone, KTJ: Karliova Triple Junction, NAFZ: North Anatolian Fault Zone, EAP: Eastern Anatolian Plateau.

that the East Anatolian High Plateau is deprived of mantle lithosphere, which is referred to break-off of northward subducted slab (started some 11 Ma ago) beneath the widespread melting due to direct contact with hot asthenosphere. According to Keskin (2007), a huge piece or the whole of the mantle lithosphere in Eastern Anatolia was detached from the overlying earth's crust, which took place at approximately 11–13 Ma, and he supported the slab steepening and break-off model (Keskin, 2003; Şengör et al., 2003). This model presents an upwelling hot asthenosphere to about 50 km in depth and extensive melting in the crust, which may support low velocity zones. Skobeltsyn et al. (2014) determined ultralow S-wave velocities at the depth of 70–150 km, which resulted from the asthenosphere with significant amounts of partial melt in the thinner lithosphere. They showed that two slab detachment events took place at roughly the same time beneath the Pontides in the north and the Bitlis slab in the south, which was the main cause of uplifting of the East Anatolian Plateau and the widespread collision-related intense volcanism.

The main tectonic feature of the Lake Van region is dramatically related to the aforementioned regional tectonic patterns. Interrelated effects between the Arabian and Eurasian Plates naturally created several strike-slip

and thrust faultings (e.g., Erciş Fault zone, Tutak Fault, Çaldıran Fault zone, Gevaş Fault zone, Malazgirt Fault, Van Fault, Kalecik Fault), prominent basins (Muş Ramp Basin, Tatvan Basin, Northern Basin, Deveboynu Basin), and widespread Pliocene and Quaternary volcanism (Nemrut volcano of the western shore and Süphan volcano of the northern shore) in and around the Lake Van region (Çukur et al., 2013; Utkucu et al., 2013). Regarding the formation process of the Lake Van Basin, it is thought that the flow out of the Muş Basin was closed by the Nemrut volcano approximately 100,000 years ago (Çukur et al., 2013) and evolution of Lake Van began. The Lake Van Basin has Eocene and Pleistocene clastic sediments (maximum thickness of sediments is ~500 m in the Tatvan Basin) outcropped Pliocene and Quaternary volcanic deposits, and upper Cretaceous carbonates and sandstones (Akinci and Antonioli, 2013; Çukur et al., 2016).

Because of the continental collision between the Arabian and Eurasian Plates, the Lake Van region and its environment have been faced with several moderate and large shocks, which usually occurred in the crust, but no subcrustal earthquakes. Generally, the mechanisms of these earthquakes are strike-slip. Considering historical and instrumental period reports, events such as the 1903 Malazgirt earthquake ($M_s = 7.0$), 1976 Çaldıran earthquake

($M_s = 7.3$), and 2011 Van earthquake ($M_w = 7.1$) caused enormous losses in the settlements (Utkucu, 2013). The latest devastating Van earthquake occurred in the NE part of Lake Van and aftershocks were over 3100 ($M_L \geq 1.8$) for 30 days. The main shock was very important evidence in terms of widespread seismicity (Bayrak et al., 2013).

3. Data and method

In this study, we used the inversion of the observed interstation phase velocity dispersion curves of the fundamental mode Rayleigh waves to obtain the 1-D shear-wave velocity structure in the studied region. For this purpose, vertical component broadband seismograms were obtained from the Observatories and Research Facilities for European Seismology (ORFEUS) databases of permanent stations (VANB, KARO, KOPR, MLAZ, GURO, AGRB, SVAN, CLDR) operated by the Kandilli Observatory and Earthquake Research Institute (KOERI). The most important criterion for interstation phase velocity studies is the degree of back-azimuth. The back-azimuth should be within $\sim 12^\circ$ between far station to

event and far station to near station (Horspool et al., 2006). According to this criterion, and in spite of examining many earthquake data, only 27 earthquakes could only be selected for this study (Figure 2) that had back-azimuth of $\sim 9^\circ$. Source parameters obtained from ORFEUS, names of the station pairs, and back-azimuth differences in degrees of each earthquake pair are seen in the Table.

Before calculating the observed phase velocities, the seismograms were demeaned, detrended, and tapered with 10% cosine-windows and instrument responses were removed in terms of seismometer types using the Seismic Analysis Code (SAC) obtained from the Incorporated Research Institutions for Seismology (IRIS). Furthermore, each seismogram bandpass is filtered from 5 to 100 s with a five-sided and two-poled Butterworth filter so that the signal-to-noise ratio can be improved (Figure 3). To eliminate possible errors due to higher-mode interference, coda, and noise, and also to extract the cleanest fundamental mode Rayleigh wave from the seismograms, secondly, we utilized useful preliminary analysis known as the multiple filter technique (MFT; Dziewonski et al., 1969)

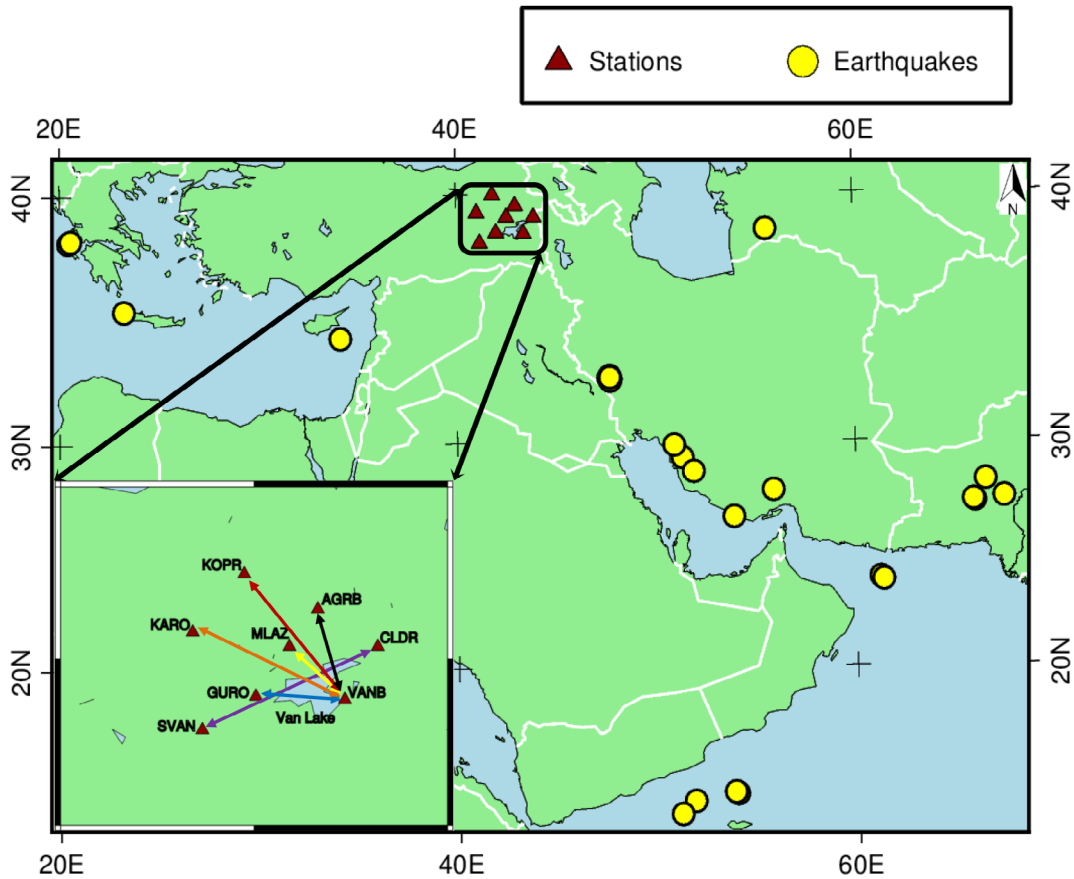


Figure 2. Claret triangles depict locations of KOERI broadband stations and yellow circles depict earthquake epicenters. In the lower-left map, stations and their codes, and each interstation path with the color scale, are displayed.

Table. Earthquakes and station pairs used for interstation phase velocities with back-azimuth difference in degrees (VANB-KOPR (dark red), VANB-MLAZ (yellow), VANB-KARO (orange), SVAN-CLDR (purple), GURO-VANB (blue), AGRB-VANB (black)).

Stations	Event date and time	Latitude	Longitude	Depth (km)	Mw	ΔBAZ
VANB-KOPR	2013.10.02T01:06:37	11.23°	57.59°	7.9	5.7	9.53°
VANB-KOPR	2013.11.28T13:49:11	29.32°	51.31°	8.0	5.8	0.80°
VANB-KOPR	2014.04.25T20:58:47	26.75°	53.87°	11.2	5.0	0.94°
VANB-KOPR	2014.06.20T22:54:53	29.87°	50.90°	10.0	5.0	0.77°
VANB-KOPR	2014.08.18T02:32:05	32.70°	47.70°	10.2	6.2	5.06°
VANB-KOPR	2014.08.20T10:11:00	32.64°	47.74°	17.7	5.6	5.07°
VANB-KOPR	2014.08.18T11:19:41	32.70°	47.62°	15.0	5.2	5.44°
VANB-KOPR	2014.08.18T11:51:31	32.78°	47.68°	10.0	5.4	4.88°
VANB-KOPR	2014.08.18T18:08:22	32.58°	47.70°	5.0	6.0	5.47°
VANB-KOPR	2014.08.18T18:08:36	32.72°	47.69°	12.0	5.6	5.04°
VANB-KOPR	2015.01.01T08:47:45	28.73°	51.86°	10.1	5.1	0.48°
VANB-MLAZ	2014.11.10T13:52:37	27.90°	55.86°	10.0	5.4	1.47°
VANB-MLAZ	2015.05.14T16:01:21	24.02°	61.21°	17.8	5.3	4.70°
VANB-MLAZ	2015.05.17T11:16:51	23.91°	61.35°	10.0	5.2	4.75°
VANB-KARO	2013.10.18T13:18:22	28.26°	66.53°	10.0	5.2	2.86°
VANB-KARO	2013.12.19T19:28:40	27.52°	67.45°	10.1	5.4	2.51°
VANB-KARO	2015.07.15T11:25:59	27.33°	65.99°	10.0	5.3	0.30°
VANB-KARO	2015.08.03T13:16:03	27.40°	65.91°	28.6	5.3	0.34°
SVAN-CLDR	2012.05.11T18:48:29	34.40°	34.14°	16.6	5.4	4.04°
GURO-VANB	2013.10.12T13:11:32	35.51°	23.25°	40.0	6.6	6.89°
GURO-VANB	2013.12.09T11:33:52	38.60°	55.60°	10.0	5.2	6.20°
GURO-VANB	2014.01.26T13:55:42	38.21°	20.45°	8.0	6.1	3.69°
GURO-VANB	2014.01.26.T13:56:19	38.29°	20.56°	10.0	5.6	3.90°
AGRB-VANB	2015.08.17T16:16:59	13.71°	51.78°	10.0	5.7	1.99°
AGRB-VANB	2015.09.02T01:13:51	14.04°	53.91°	10.0	5.3	6.54°
AGRB-VANB	2015.09.02T07:20:30	14.13°	53.79°	10.0	5.0	6.36°
AGRB-VANB	2015.09.22T04:03:14	13.09°	51.12°	10.0	5.2	0.30°

and phase match filter (PMF; Goforth and Herrin, 1979) just before the application of the slant stacking technique. The MFT is used to determine the group velocity dispersion for the single station method based on the properties of a dispersive signal that consists of different frequency components arriving at a station at different times (Perez, 2001). This method is applied on a series of narrow bandpass Gaussian filters centered at fixed frequencies that estimate the arrival time of the maxima of the surface-wave envelope for each center frequency (Cristiano et al., 2010). Once group velocities are obtained from the MFT, the PMF suppresses the waveform into a narrow time window centered near zero time, permitting noise to be windowed out by taking the spectrum of this narrow window rather than the full signal time window (Perez, 2001). Third, the vertical component

seismograms containing the fundamental mode were inputs for the slant stacking method, which is used to calculate Rayleigh wave phase velocities. This array technique was first described by McMechan and Yedlin (1981), consisting of two linear transformations to obtain a wave field in the $\rho - \tau$ domain (phase slowness–intercept time), followed by a 1-D Fourier transformation over τ to obtain the wave field in the $\rho - \omega$ plane (Mokhtar et al., 1988). This approach was tested on each seismogram pair to determine phase velocity dispersion and reasonable results were achieved in the Lake Van region. Figure 4 shows an example of the phase velocity dispersion curve for the data of the VANB-KARO stations.

The phase velocity dispersion curve inversion process is generally a nonlinear problem, and it is calculated by using

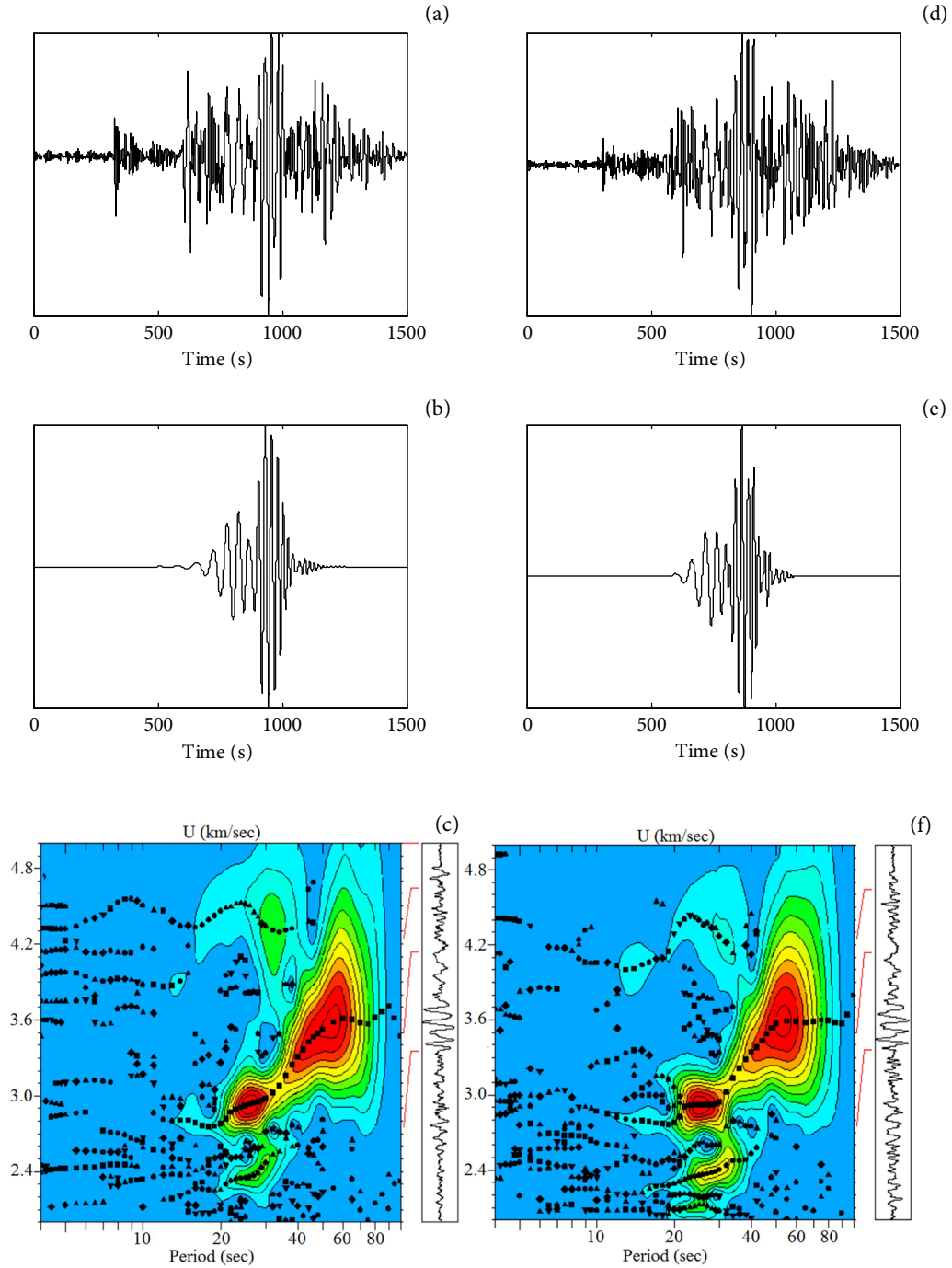


Figure 3. Panels (a) and (d) show vertical component broadband raw seismograms recorded at the KARO and VANB stations (Event No. 16 shown in the Table), respectively. Panels (b) and (e) are fundamental mode seismograms after MFT and PMF processing. Panels (c) and (f) are the contour maps of the relative amplitude of wave energy for KARO and VANB, respectively. The red area is associated with the largest amount of energy (see text for details).

the linearized least squares algorithm (Herrmann, 2013) to obtain the 1-D plane-layered shear-wave structure. To achieve this, we used the computer program SURF96 (Herrmann, 2013) using singular value decomposition

in the stochastic or differential form (Russel, 1987). The linearized least squares algorithm compares seismic velocity from each layer of the earth model and determines how well it fits to the observed data. The final earth

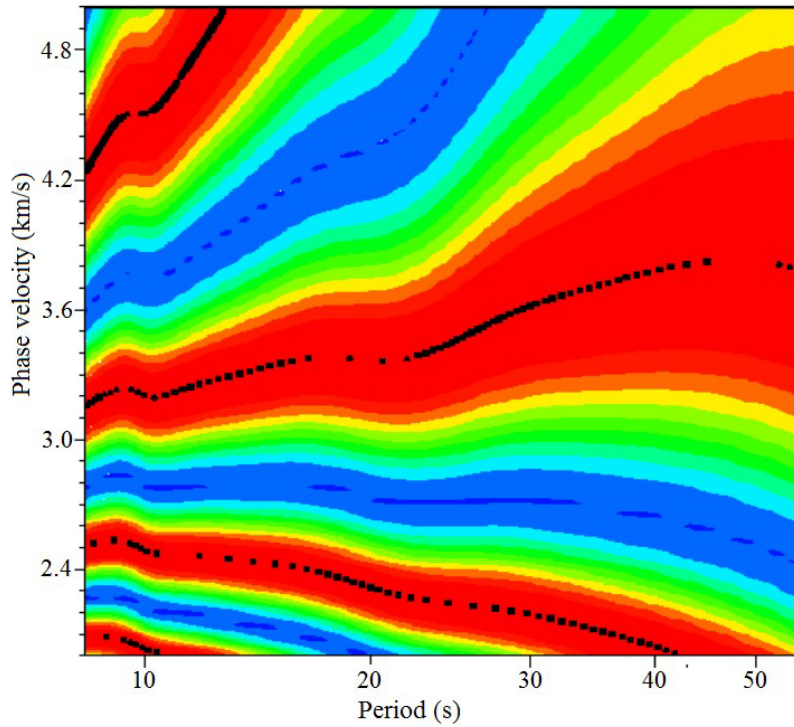


Figure 4. Interstation phase velocity dispersion curve vs. period obtained between VANB and KARO stations from Event No. 16. The largest red area indicates the maximum amplitude zone, and smaller red areas indicate misaligned waveforms. The black square symbols in the largest red area represent the fundamental mode Rayleigh wave phase velocity dispersion curve.

model defines 1-D shear-wave velocities, compressional-wave velocities, and density for each layer. According to Herrmann (1991), the damping value at the beginning of the inversion process should be selected as the highest value, and it must be reduced to the lowest one during the convergence process because of its closeness to the highest eigenvalue. We have chosen the appropriate damping values to stabilize the inversion for this reason, and it has decreased at each value in each of five iterations (10, 5, 2, 1, 0.1). We have an initial model for the inversion process with the combination of CRUST1.0 and IASP91 (IRIS-Reference Earth Models). The thickness and the numbers of layer in the combined model consist of 37 layers (first layer is 1 km, layers between 2 and 21 are 2 km, layers between 22 and 30 are 5 km, and layers between 31 and 37 are 10 km along increasing depth) and 150 km in total thickness. Poisson's ratio is constant for each layer and the empirical relation of Berteussen (1977) is only used to constrain IASP91 density values.

In an inversion problem, the resolution matrix clarifies a linear relationship in that each solution parameter is derived from the weighted averages of nearby true-model parameters, and the resolution matrix elements are the weights. Resolution matrices are generally used both to

calculate the obtainability of the solution and to isolate and extract spatial and temporal resolutions (An, 2012). The statistical resolution matrices described by An (2012), generated using Gaussian approximation, are employed to determine the resolution length information of the total inversion system. We have calculated the exactness in the solution of our inversion problem via this method.

4. Discussion and conclusions

Rayleigh waves mostly have the strongest arrivals recorded at the teleseismic distance, and they provide the best constraints on the structure of the earth (Perez, 2001). These waves are more sensitive to shear-wave velocity change along the propagation path (Tang and Chen, 2008). For this reason, we have generally used the far regional events to image the detailed major discontinuities (e.g., crust-mantle) and to obtain the shear-wave velocity structure of the Lake Van region, relying on the inversion of interstation phase velocity dispersion curves. However, the interstation phase velocity method has a significant disadvantage if sufficient station pairs and appropriate earthquake epicenter to interstation paths are unavailable. Therefore, we have considered and calculated observed Rayleigh wave phase velocity curves for only six different

station pairs (VANB-KOPR, VANB-MLAZ, VANB-KARO, SVAN-CLDR, GURO-VANB, AGRB-VANB). Since each station pair has a different number of dispersion curves, we have statistically averaged each individual phase velocity curve in order to minimize the influence of lateral heterogeneity along the wave propagation path (Erduran et al., 2008). Thus, we have inverted the average phase velocity curves to obtain the 1-D S-wave structure of the studied area, and we have added the inversion results of each individual phase velocity curve. For each station pair, in Figures 5–7, panels a and d show average (red line) and observed (black line) Rayleigh wave phase velocity dispersion curves, panels b and e show average (red line) and observed (black line) S-wave velocity structures, and panels c and f show the statistical resolution matrix, which depends on the fitting of the last model. Phase velocity curves show generally similar characteristics between periods of 15–50 s (Figures 5–7, panels a and d).

According to the normalized statistical resolution matrix figures (Figures 5–7, panels c and f), parameter index *i* and *j* coincide with row and column elements of the statistical resolution matrix, respectively. Each row of the resolution is equal to the contributions of all model parameters in the *i*th solution parameter. If the resolution matrix is an identity matrix, the model parameter can make a huge contribution to a solution parameter and each row from the resolution matrix can be approximated by a Gaussian function. An (2012) expressed that the average resolution length is taken to be equal to half of the recovered checker dimension (Lebedev and Nolet, 2003), which is very similar to the half width of a Gaussian function. Because of this, red-colored contours in the normalized resolution matrix figures refer to the resolution length. The 1-D S-wave velocity structure used as an initial model in the inversion has 37 layers constructed from the combination of CRUST 1.0 and IASP91 models. This combined model has carefully been inverted for a total of 25 iterations. The initial (black line) and the last (blue line) models (Figures 5–7, panels b and e) for all iterations in the inversion process have been indicated with colored lines varying from red to blue according to the iteration number. As shown in panels c and f of Figures 5–7, the spatial resolution lengths are generally compatible with each other in the medium depth range, except the SVAN-CLDR path. This path has a single event and observed phase velocity data in the period range from 15 s to 36 s, demonstrating large resolution lengths before the 8th parameter and after the 32nd parameter. As the spatial extent of each parameter represents model layer thickness, the depth of these parameters corresponds to approximately 15 and 101 km, respectively. Additionally, another important example in Figure 7 associated with the VANB-KOPR path shows that the resolution lengths

increase from 8 km for the 10th parameter to 15 km for the 20th parameter. While all resolution matrices related to the inversions of the average phase velocities exhibit poor results in the upper crustal depths owing mostly to lack of the observed phase velocity data in the lower period range, good resolutions are seen in the lower-crustal and upper-mantle depths.

Since the fundamental mode Rayleigh wave dispersion is very sensitive to S-wave velocity, we have calculated and plotted the sensitivity kernels ($dC/d\beta$) as a function of depth for the initial model at different periods (15 s, 22 s, 30 s, 38 s, 55 s) in Figure 8. The longer periods have more sensitivity to shear wave velocities at greater depths, whereas the shorter periods are sensitive to lower depths (below 30 km). In longer periods (30 s, 38 s, and 55 s), there is still sensitivity up to greater depths below 150 km. As there are no phase velocity data in longer periods for some of the interstation paths, we have limited the initial model used in the inversion processes with 150 km depth.

Figure 9 depicts 1-D S-wave velocity structures derived from all inversion results for both each individual phase velocity dispersion curve and their averages. General characteristics of the velocity models are approximately similar to each other. We have obtained the shear velocity value in the upper-crustal depths (nearly 1–10 km) as ~3.3 km/s, which corresponds to the thick volcanosedimentary successions (Angus et al., 2006). The shear-wave velocities gradually increase in the range of 10 to 20 km depths. At depths greater than 20 km, nearly the 30–40 km depth range, interstation paths exhibit that seismic velocities decrease to as low as 3.5 km/s in the studied region. According to Angus et al. (2006), a discontinuity existing at roughly 25 km in depth (longitude 42°E) is associated with a crustal low velocity zone representing a pocket of partial melt, with observation of geothermal centers and below Quaternary volcanic centers near the city of Van. Gök et al. (2007) determined very low S-wave velocities (about 3.6 km/s) at the lower crustal depth (30–38 km) in the vicinity of the Lake Van region. They interpreted such a low shear-velocity as the location/surface distribution of the Neogene and Holocene volcanoes in eastern Turkey. Gökalp (2012) suggested that high velocities are related to ophiolitic units at the upper crust and extensive low velocities stem from high temperature magma chambers or partly molten uppermost mantle at the middle crust (≥ 20 km) in the East Anatolian Plateau. Irmak et al. (2012) studied the source mechanism of the 2011 Van earthquake main shock and aftershocks. The continental–continental collision plate motion in the region has caused thrust faults, and one of them generated the Van earthquake with a NE strike and SW dip, and aftershocks. The aftershocks (trending NE–SW) on the main thrust fault plane generally occur within the upper nearly 25 km depth of the crust

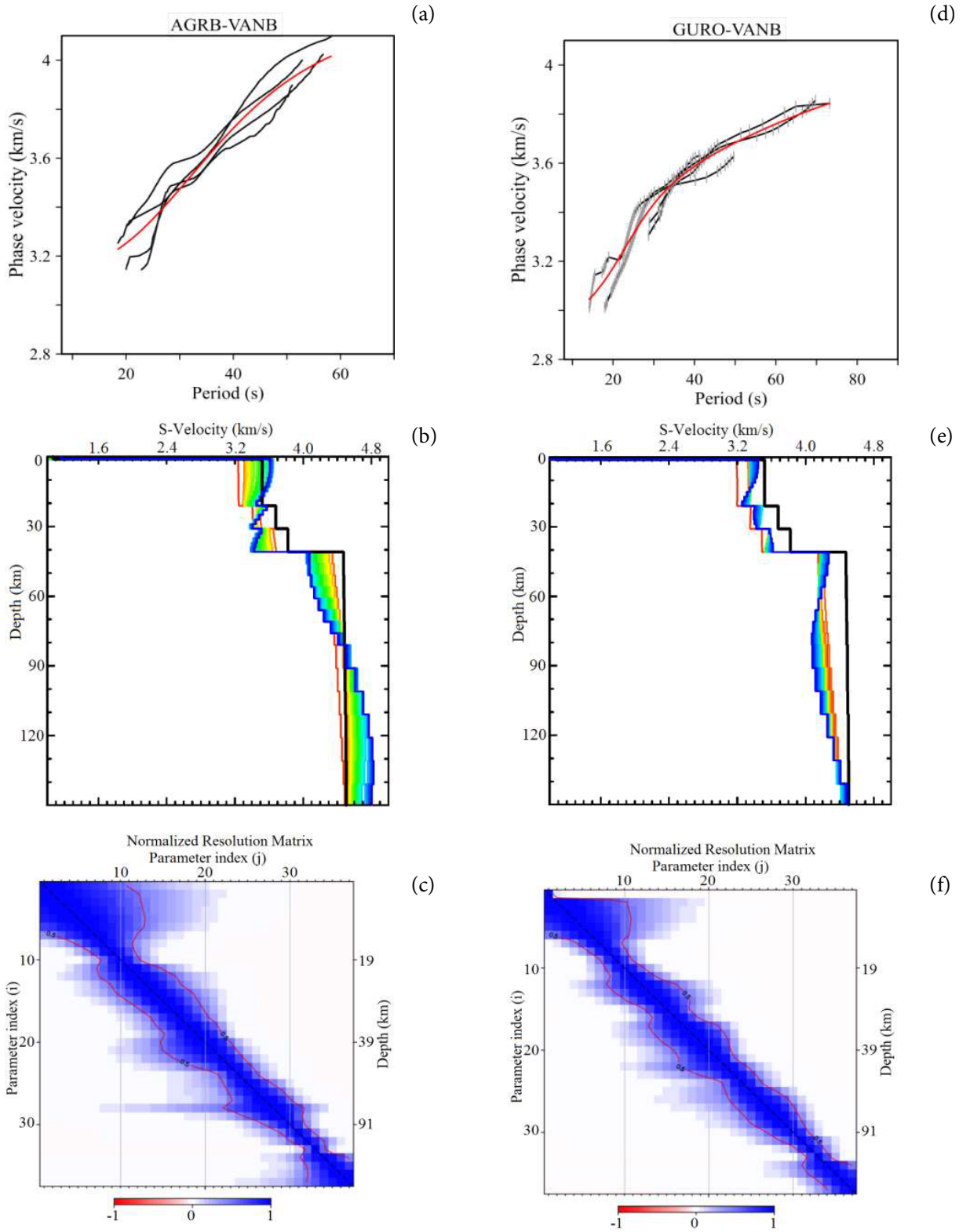


Figure 5. Results of interstation paths for AGRB-VANB (a, b, c) and GURO-VANB (d, e, f). Panels (a) and (d) depict individual observed phase-velocity dispersion curves obtained from the slant stacking technique (black line) and their average (red line). (b, e) 1-D shear-wave velocities derived from the inversion result of average dispersion curve; black line represents the initial model and blue line represents the fitting of the last model at the 25th iteration. (c, f) Normalized resolution matrix based on the fitting of the last model.

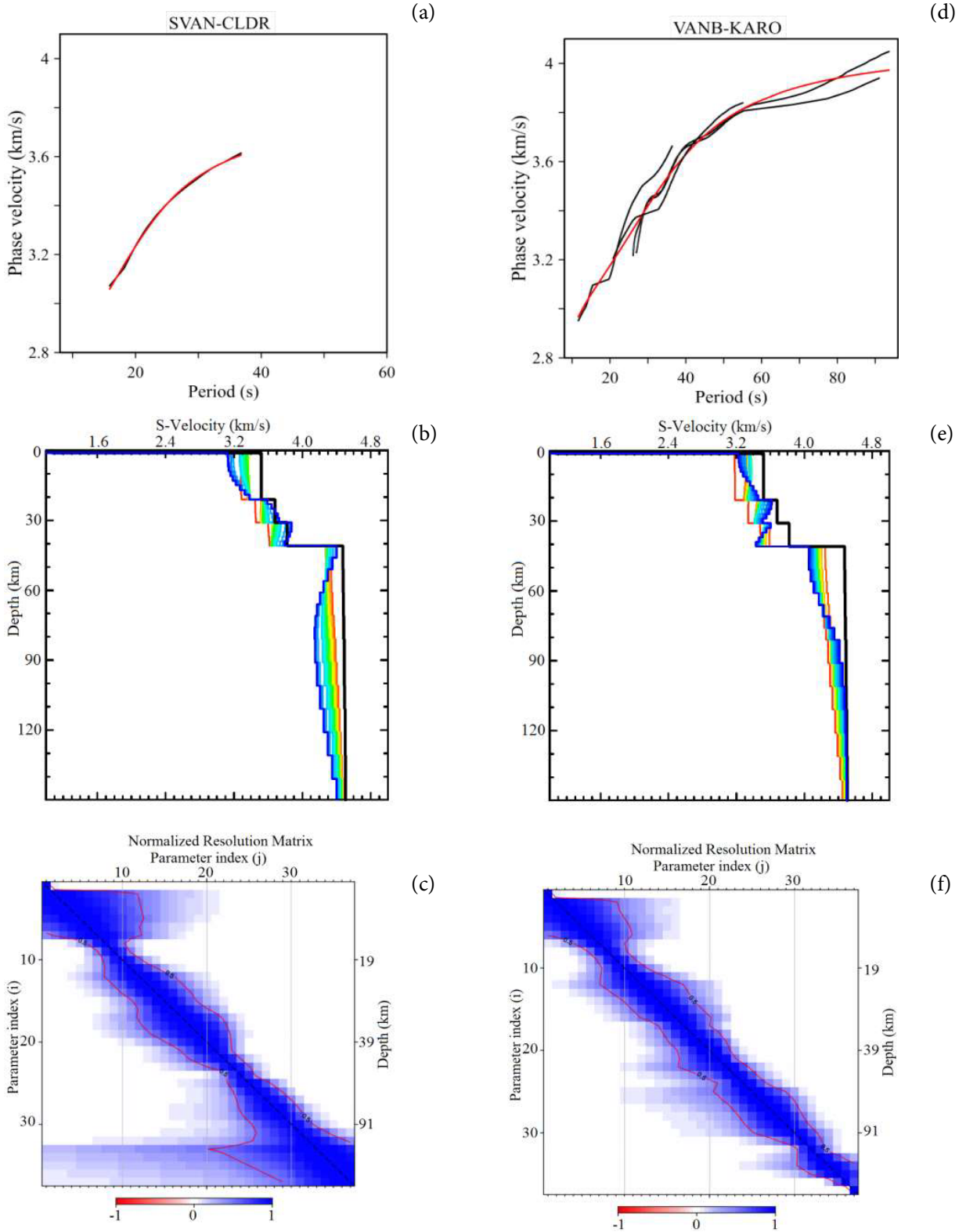


Figure 6. Results of interstation paths for SVAN-CLDR (a, b, c) and VANB-KARO (d, e, f). Panels (a) and (d) depict individual observed phase-velocity dispersion curves obtained from the slant stacking technique (black line) and their average (red line). (b, e) 1-D shear-wave velocities derived from the inversion result of average dispersion curve; black line represents the initial model and blue line represents the fitting of the last model at the 25th iteration. (c, f) Normalized resolution matrix based on the fitting of the last model.

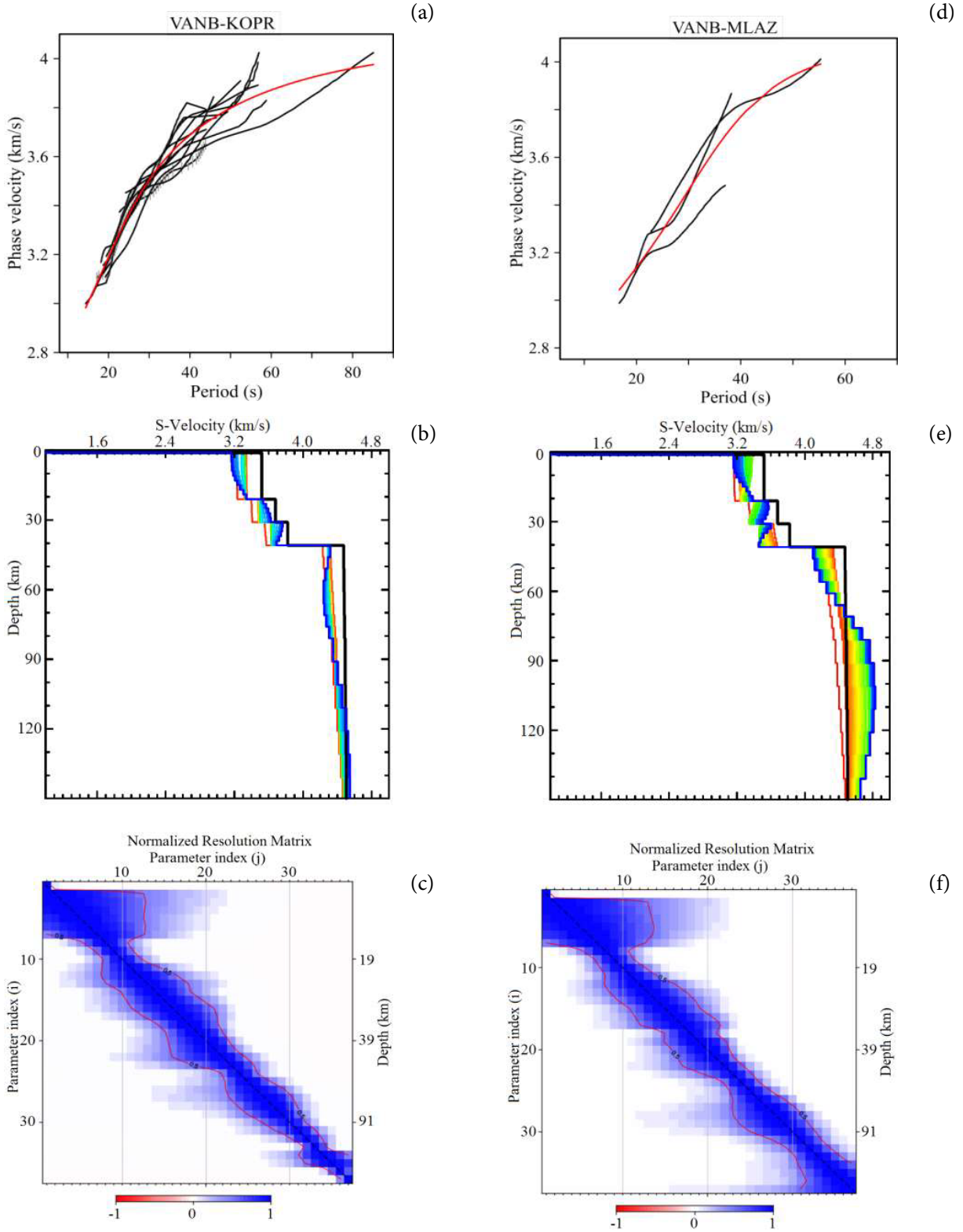


Figure 7. Results of interstation paths for VANB-KOPR (a, b, c) and VANB-MLAZ (d, e, f). Panels (a) and (d) depict individual observed phase-velocity dispersion curves obtained from the slant stacking technique (black line) and their average (red line). (b, e) 1-D shear-wave velocities derived from the inversion result of average dispersion curve; black line represents the initial model and blue line represents the fitting of the last model at the 25th iteration. (c, f) Normalized resolution matrix based on the fitting of the last model.

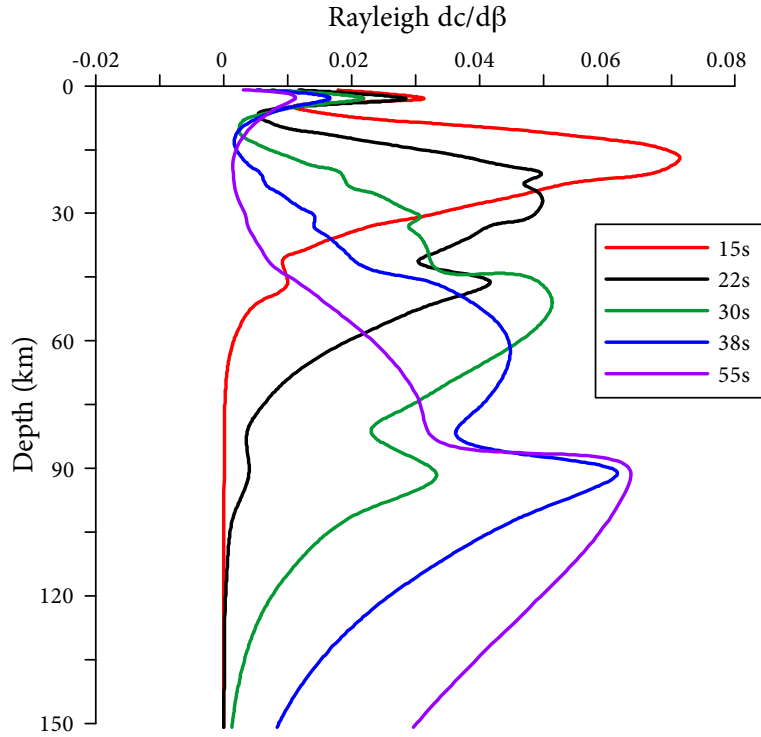


Figure 8. An example of Rayleigh wave sensitivity kernels ($dC/d\beta$ versus depth) calculated in different periods for VANB-GURO interstation path according to the initial shear velocity-depth model (modified from Crust1.0 and IASP91 model) used in the inversion.

(Angus et al., 2006; Irmak et al., 2012). In addition to these, Dragoni (1993) said that at sufficiently high temperatures and pressures, brittle behavior in the crustal materials gives way to ductile behavior. This creates a broad transition (a seismogenic layer), which takes place in an interval of depths depending on physical parameters such as geothermal gradient, strain rate, and activation energy, called the brittle ductile transition. As earthquakes are the consequence of a frictional instability, the depth limitation of the earthquakes must be controlled by a stability transition, determining a lower stability transition of the earth's crust at similar depths as the brittle ductile transition. For these reasons, this depth level (~25 km) may be related to the beginning of the low velocity zone, the brittle ductile transition, and no subcrustal earthquakes in the Lake Van Basin.

In accordance with Figure 9 (thick red line), Figure 10 depicts 2-D shear wave velocity models for the 6 broadband station pairs, and it is composed of Conrad (the upper crust–lower crust boundary) and Moho (the crust–mantle boundary) discontinuities. Vertical and horizontal axes show the depth (km) and latitude and longitude of each station pair, respectively. According to Figure 10, beneath the Lake Van region, the existence of the Conrad discontinuity (the upper crust–lower crust boundary)

and the Moho discontinuity (the crust–mantle boundary) is well established. The Conrad discontinuity marks the transition between the upper (granitic) and the lower (basaltic) crust boundary and has been identified on most continental parts of the earth's crust, but it is not present worldwide (Abdelwahed et al., 2013). It is at this depth that numerous intermediate-depth microearthquakes occur, suggesting that these earthquakes define the Conrad discontinuity (Merriam, 2006). In addition, Zhao et al. (1992) revealed that P-wave velocities in the upper crust and the lower crust at approximately 5.9 and 6.6 km/s and S-wave velocities in the upper crust and the lower crust are generally 3.5 and 3.8 km/s, respectively. Delph et al. (2015b) determined that the lower crustal shear velocity corresponds to ~3.5 km/s at the depth of approximately 25 km throughout 38–39°N in the Lake Van region. Based on joint inversion of surface wave and receiver function analysis results, Warren et al. (2013) found that the shear velocity is ~3.6 km/s at a depth of 30 km. According to these studies, by comparing these studies, we could determine that the Conrad discontinuity is generally located at a depth of 27 km and shear velocities vary from 3.4 to 3.7 km/s. The Conrad discontinuity seems to be shallow beneath the KOPR-VANB and GURO-VANB station pairs due to the use of more data and obtaining

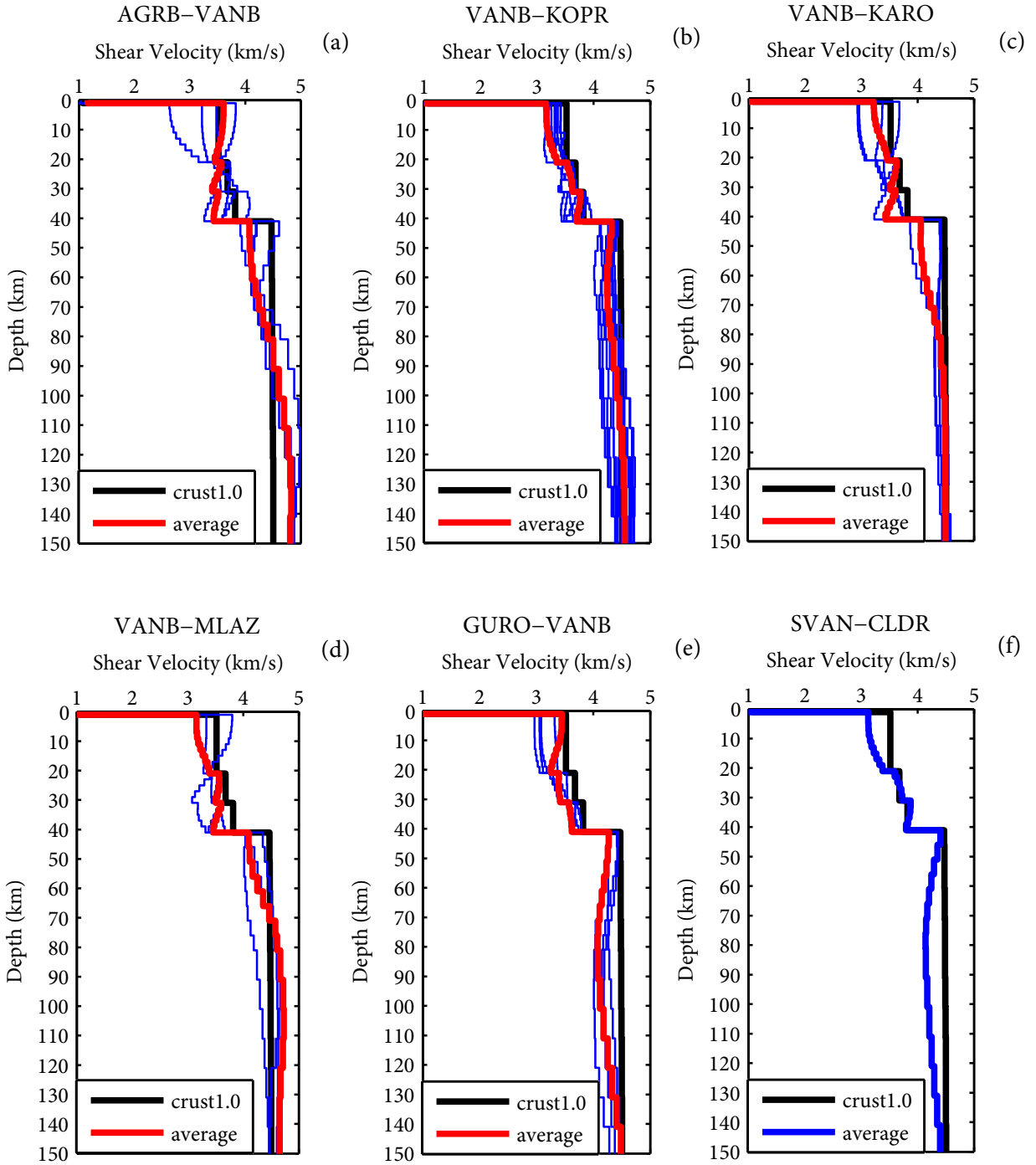


Figure 9. Comparison of 1-D shear-wave velocity models derived from the inversion profiles. The thick red line represents the fitting of the last model derived from the inversion of the average phase velocity dispersion curves, black line represents the initial model (CRUST 1.0+IASP91), and thin blue lines represent each individual inversion result. Panel (f) has a single event and the average velocity is shown in blue color.

shear wave velocities, whereas the depths of Conrad discontinuity beneath the other pairs are taken as 28 km.

The Moho discontinuity is located in a depth ranging from 42 to 44 km toward the north. Between the AGRB-

VANB, KOPR-VANB, and KARO-VANB station pairs, Moho discontinuity is 44 km, while the crustal thickness is shown as 42 km beneath the MLAZ-VANB, GURO-VANB, and SVAN-CLDR station pairs passing through

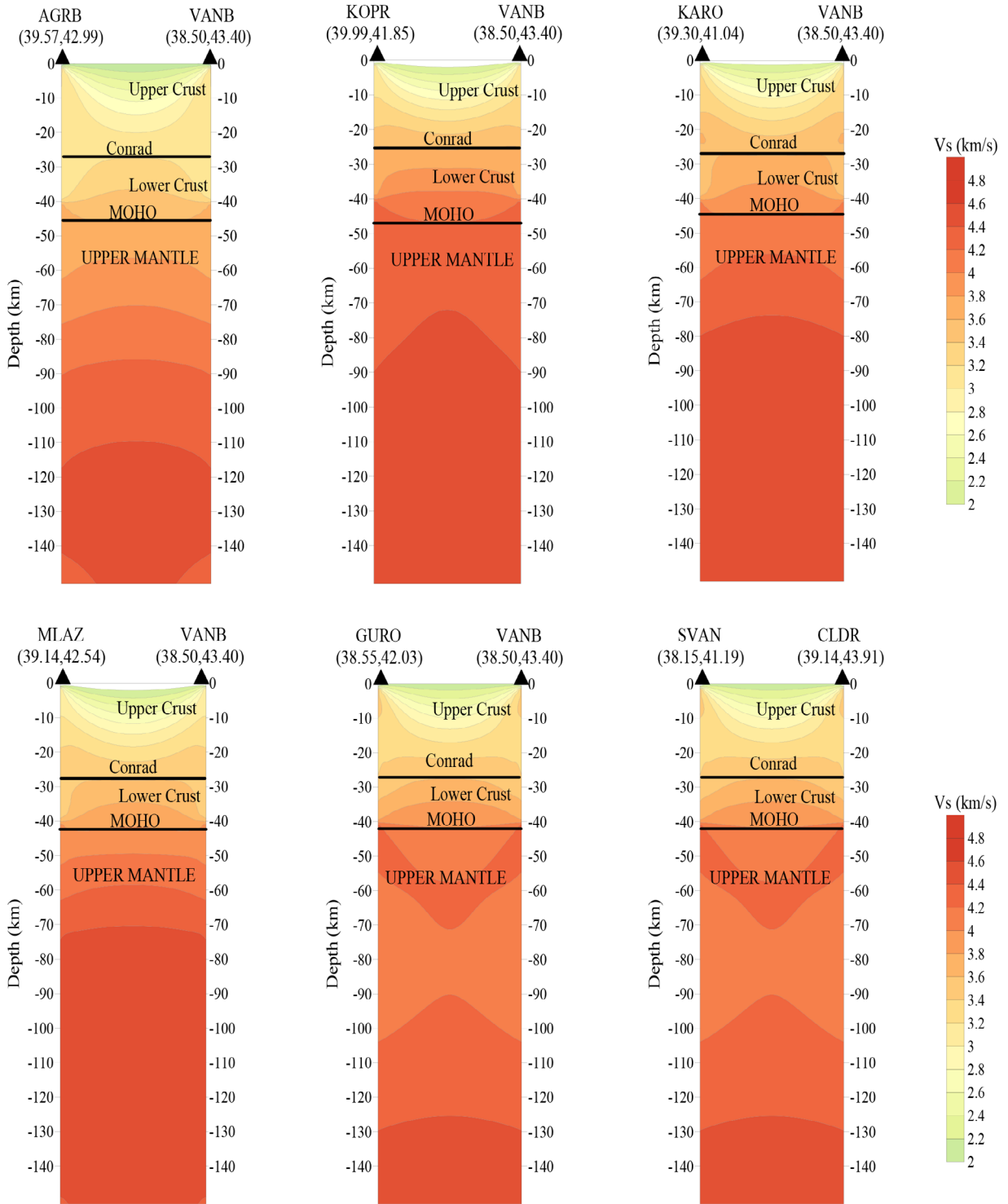


Figure 10. The 2-D shear velocity models of the crust and mantle for the six station pairs, obtained from the last models in Figure 9 (thick red line). The Conrad and Moho discontinuities are drawn with black lines. The locations of the profiles are shown in Figure 2.

Lake Van. S-wave velocities in the Moho discontinuity are generally 4.1 and 4.3 km/s. The Eastern Anatolian Plate is generally characterized by high heat flow, broad internal

deformation, slow shear wave velocities, and Miocene to Recent volcanics (Delph et al., 2015b). The KTJ, which separates the Anatolian and Eurasian Plates, has some of

the highest strain rates and it is located within the Bitlis Massif. According to Delph et al. (2015b), the Arabian continental lithosphere appears under the thrust as far as the northernmost boundary of the eastern Bitlis Massif; however, this move of the Arabian continental lithosphere does not reveal how far north it totally is. In particular, the KOPR-VANB station pair lies within the KTJ. As there are the high exhumation rates in the Bitlis Massif, we suggest that Moho depth becomes deeper toward the north. Due to the fact that the Moho dips slowly from the Arabian margin to the eastern Anatolian Block (Angus et al., 2006), crustal thickness in Figure 10 confirms the results of the each station pair.

Tezel et al. (2013) in eastern Turkey obtained Moho depths around 40–46 km and shear-wave velocities around 4.0–4.2 km/s. Özacar et al. (2010), using teleseismic P-wave receiver functions, found that thinner Moho depth is ~38 km near the west of Lake Van with high V_p/V_s (≥ 1.85) variation due to partial melt in the crust, while crustal thickness around Lake Van is approximately 40 km. Furthermore, a low-velocity zone interpreted as a decoupling zone is observed via strong negative phases at the lower crust beneath the north of Lake Van. Moreover, Mutlu and Karabulut (2011) depicted that the low Pn-velocity anomaly in Eastern Anatolia is related to volcanism following the break-off of the subducting oceanic lithosphere and the crustal thicknesses vary from 38 to 40 km in the Lake Van region. According to Gök et al. (2003), the S_n -wave was not observed in and around eastern Turkey. Their best explanation for observed S_n attenuation was the upwelling of the asthenosphere, which led to extensive partial melting throughout the uppermost mantle. Bakırcı et al. (2012) estimated that slow S-wave velocities (~4.0 km/s at 50 km depth) are seen in eastern Turkey.

In Figure 10, the upper mantle S-wave velocities are lower than 4.30 km/s between 45 and 70 km depth ranges. The reasons why results of average S-wave velocities are slower than global models may be connected to the shallow asthenosphere and young volcanism in the studied region. Gök et al. (2011) found that the Moho is at depths of 40–42 km around the Lake Van region and estimated that low shear-wave velocities are $V_s = 3.2$ km/s, $V_s = 3.7$ km/s, and $V_s = 4.1$ km/s for the depths of 10, 35, and 85 km, respectively. By checking regional waves in the crust (L_g) and the upper mantle (S_n), they observed both waves disappearing through the Eastern Anatolian Plateau, coinciding with the zones of inefficient S_n propagation, and heating of a thin to absent mantle. Zor's (2008)

tomographic results showed that the negative velocity anomaly above the slab-like positive anomaly may be an upwelling hot asthenosphere in the upper mantle beneath Eastern Anatolia and brings about slow shear-wave velocity anomalies. A magnetotelluric study carried out in Eastern Anatolia by Türkoğlu et al. (2008) showed that the upper mantle has a very low resistivity, which is related to the presence of shallow and partially molten material. The shear-wave cross-sections of Warren et al. (2013) indicate that low shear velocities at 10 km (3.4 km/s) and 30 km (3.6 km/s) are associated with young volcanism and the continued presence of hot rock. Due mostly to the possible hot asthenospheric rising, there is a slow velocity of ~4.2 km/s at the uppermost mantle. Results of Delph et al. (2015a) point out that the slower V_s velocity in the uppermost mantle (4.23 km/s) at 5 km below the Moho near Lake Van is associated with the influx of a hot asthenosphere.

The S-wave results of the crust and mantle structures obtained in the present study are reasonable and have good agreement with the existing geological and geophysical information. Even though shallow discontinuities (the first 10 km) have not been well resolved, our inversion results have clearly presented the low-velocity zones at ~30–40 km depth caused by the probably widespread volcanism since the Quaternary and partly molten uppermost mantle in and around the Lake Van region. The inferred average crust–mantle discontinuity has been displayed at the depth of ~42 km, with averaged S-wave velocity feebly jumping from 3.6 km/s to 4.2 km/s. Inversion results have shown that average S-velocities in the upper mantle throughout the interstation paths are less than 4.30 km/s, lower than the global average, inferred from an upwelling hot asthenosphere or a lack of significant lithospheric mantle. Finally, our S-wave velocity models have provided helpful and reliable results to understand the structure of the earth in the studied region.

Acknowledgment

The authors would like to thank the editor of the *Turkish Journal of Earth Sciences* and the anonymous reviewers for their valuable discussions and suggestions that greatly improved this manuscript. Figures in this manuscript were generated by using the Generic Mapping Tools (GMT) developed by Wessel et al. (2013). We would also like to thank the Kandilli Observatory and Earthquake Research Institute (KOERI) and the Incorporated Research Institutions for Seismology (IRIS) for providing the earthquake data.

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