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Three-dimensional *S*-wave structure of the upper mantle beneath Turkey from surface wave tomography

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SUMMARY

A 3-D upper-mantle structure beneath Turkey is investigated using phase speeds of fundamental-mode Rayleigh waves employing a conventional two-station method with high-density seismic networks in Turkey. We analyse 289 seismic events with moment magnitude 5.5 and greater, and with focal depth shallower than 100 km between 2006 and 2008. Waveform data are derived from 164 three-component broad-band seismic stations operated by two national seismic networks. At first, Rayleigh-wave phase speed maps are obtained from the inversion of two-station phase speeds using about 1000–3000 paths, depending on the period of Rayleigh waves. The three-dimensional *S*-wave model is then obtained in the depth range from 40 to 180 km using the phase speed maps in the period range from 25 to 120 s. Our model reveals the fast anomalies in the north of Cyprus associated with the subducted portion of the African oceanic lithosphere from the Cyprus trench. We identify a vertical discontinuity of the fast anomaly associated with the Cyprus slab starting at 60–80 km depth which may represent a minor tear of the Cyprus slab. We observed that the western part of the Cyprus slab is getting closer to the edge of the Hellenic slab beneath the Isparta Angle (IA) and Antalya Basin. Our model also indicates a slow wave speed anomaly beneath the IA and Antalya Basin probably due to hot materials of asthenosphere rising from a tear of the subducted African oceanic lithosphere; that is, a slab tear between the Cyprus and the Hellenic subductions. In the eastern part of Turkey, a widespread slow anomaly appears in the model that corresponds to the Eastern Anatolian Accretionary Complex (EAAC). Our model shows a fast anomaly beneath the EAAC that can be interpreted as the detached portion of the subducted Arabian lithosphere.

Key words: Surface waves and free oscillations; Seismic tomography; Subduction zone processes; Continental margins: convergent; Dynamics of lithosphere and mantle.

1 INTRODUCTION

Improvements in the methods of seismic tomography over a few decades have enabled us to map high-resolution images of 3-D Earth structure. With such tomographic images, we can detect a variety of structural features within the Earth that allows us to discuss the tectonic and dynamic processes in the mantle. Surface wave tomography on a regional and global scale is a powerful tool for investigating the large-scale structure of the crust and upper mantle to better understand the geodynamic processes and evolution (e.g. Woodhouse & Dziewonski 1984; Trampert & Woodhouse 1995; van Heijst & Woodhouse 1999). High-resolution imaging of the Earth's internal structure has recently become possible owing to the rapidly increasing numbers of high-quality digital seismic data from newly installed stations, and development of recording and tomographic imaging techniques.

Turkey and the surrounding regions are one of the most active regions in the world with its active and complex tectonics encompassing plate subduction, collision, extension, crustal thickening and high seismicity. This area has been studied by using body and surface waves as a part of broader scale regional models; for example, European or Eurasian models with *P* wave (Romanowicz 1980; Spakman *et al.* 1993; Al-Lazki *et al.* 2004), and with surface waves (Levshin *et al.* 1992, 1994; Marquering & Snieder 1996; Ritzwoller & Levshin 1998; Villasenor *et al.* 2001; Marone *et al.* 2003; Pasyanos 2005; Kustowski *et al.* 2008). Turkey is also included in many local-scale studies in Mediterranean (*P* wave: Panza *et al.* 1980; Spakman 1991; Spakman *et al.* 1993; Piromallo & Morelli 1997, 2003, surface waves: Meier *et al.* 2004; Erduran *et al.* 2008), in Alpine-Himalayan (*P* wave: Koulakov *et al.* 2002; Alpine–Mediterranean–Piromallo & Morelli 2003), and in Turkish–Iranian Plateau (surface wave: Maggi & Priestley 2005),

not to mention in global-scale studies (Surface wave: Ekström *et al.* 1997; Ritzwoller *et al.* 2002). Recently, investigations on the crustal thickness and seismic velocities of smaller areas from local networks or temporary deployments of seismic arrays have been done in the Marmara region (Bariş *et al.* 2005; Zor *et al.* 2006), in the western Anatolia (Saunders *et al.* 1998; Akyol *et al.* 2006; Tezel *et al.* 2010), in the eastern Anatolia (Gök *et al.* 2003; Sandvol *et al.* 2003; Zor 2008), and in the Aegean Sea (Karagianni *et al.* 2002, 2005; Bourova *et al.* 2005). Tomographic studies of the lithospheric mantle structure of Turkey are performed using *P* wave by Mindevalli & Mitchell (1989), Biryol *et al.* (2011), and using surface wave by Cambaz & Karabulut (2010). Although a variety of seismological studies have been performed in and around Turkey, a high resolution 3-D imaging of seismic structure beneath whole Turkey with surface waves has yet to be performed, and the detailed 3-D structure of the uppermost mantle beneath Turkey is still largely unknown.

In this study, we investigate the upper-mantle structure beneath Turkey by constructing a high-resolution 3-D *S*-wave speed model using phase speeds of the fundamental-mode Rayleigh waves. We obtain phase speed dispersion curves with the conventional two-station method (Dziewonski & Hales 1972), using 164 broad-band stations in Turkey and its surroundings operated between 2006 and 2008. The phase speed models are derived from the measured interstation phase speeds employing a mapping method developed by Yoshizawa & Kennett (2004). We obtain the regional-scale phase speed maps of Rayleigh waves covering the entire region of Turkey as a function of period between 25 and 120 s with a 5 s interval. We then construct the 3-D *S*-wave speed model by inverting local dispersion curves that are assembled from the phase speed maps. This high-resolution isotropic *S*-wave speed model down to the depth of

180 km is used to discuss regional tectonics and dynamic processes in the uppermost mantle beneath Turkey.

1.1 Geology and geodynamic settings of Turkey

Turkey is one of the most active earthquake zones in the world, located along the east–west direction in the Alpine–Himalayan zone. During Palaeozoic and Mesozoic era, Anatolia is formed as an orogenic belt consisting of different continental and oceanic segments related to closing regime of the Tethys Ocean (Şengör & Yilmaz 1981; Görür *et al.* 1984; Şengör 1987; Ricou 1994; Stampfli 1996; Okay & Tüysüz 1999; Bozkurt 2001). Turkey is currently formed as a single landmass by the conjoined terranes; that is, different continental fragments comprising ophiolites and accretionary prisms and are separated by sutures, which were divided by oceans during Phanerozoic era (Okay 2008). Main fragments and suture zones of Turkey and major volcanoes (Aydar & Gourgaud 1998; Keskin *et al.* 1998; Yilmaz *et al.* 1998; Pearce *et al.* 1990; Rojay *et al.* 2001; Aydar *et al.* 2003; Bridgland *et al.* 2007; Lustrino *et al.* 2010; <http://www.volcano.si.edu/world>) are displayed in Fig. 1.

In northern Turkey, the Rhodope–Strandja (RSZ), İstanbul Zone (İZ) and Sakarya Zone (SZ), namely the Pontides, exhibit Laurasian affinity (Okay 2008). The RSZ consists mainly of quartzo-feldspathic gneisses and crosscutting Late Carboniferous and Early Permian intrusions (Okay 2001). The İZ is defined as a continental fragment consisting of late Precambrian crystalline. SZ zone is characterized by Carboniferous crystalline basement, Palaeozoic granitoids and Permo-Triassic low-grade metamorphic subduction–accretion complexes (Karakaya Complex). The Kırşehir Massif (KM) located in Central Anatolia is defined by the medium to high-grade metamorphic rocks and is

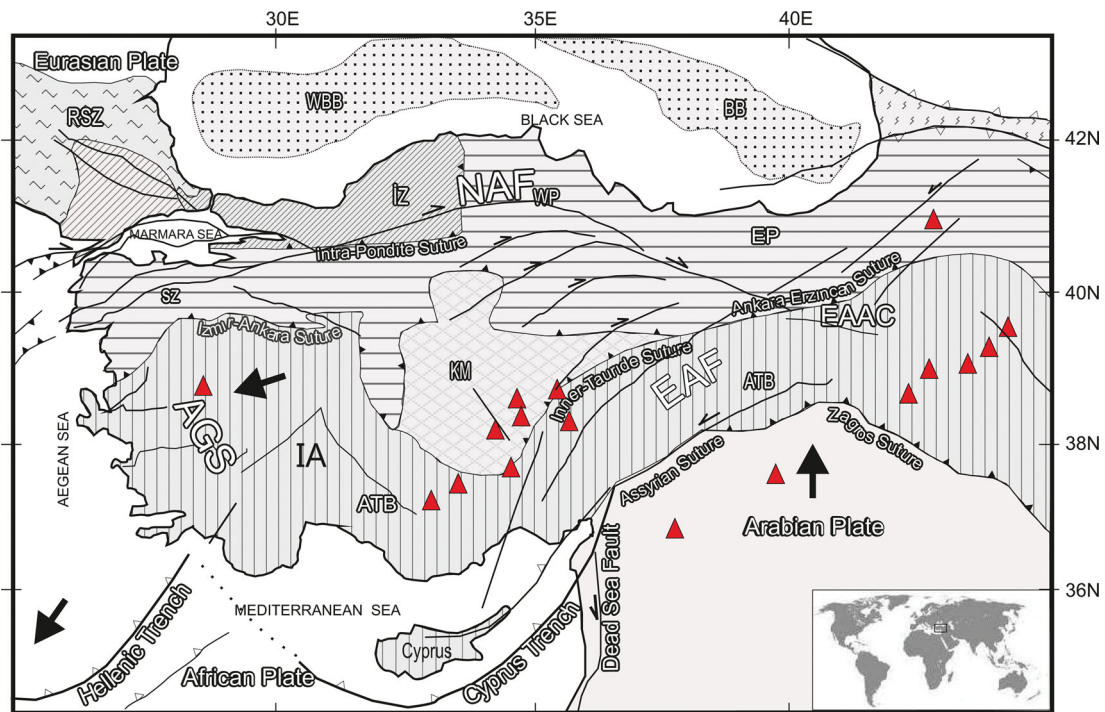


Figure 1. Tectonic and geologic map of Turkey. RSZ: Rhodope–Strandja Zone, İZ: İstanbul Zone, SZ: Sakarya Zone, WP: Western Pontides, EP: Eastern Pontides, WBB: Western Black Sea Basin, EBB: Eastern Black Sea Basin, ATB: Antolide–Tauride Block, KM: Kırşehir Massif, EAAC: Eastern Anatolian Accretionary Complex, IA: Isparta Angle, AGS: Aegean Graben System, NAF: North Anatolian Fault, EAF: East Anatolian Fault, triangles: volcanoes, thick arrows: direction of the plate motions, (Okay and Tüysüz 1999; Şengör *et al.* 2003).

crosscut by Cretaceous and younger granitoids (Göncüoğlu 1986; Akıman *et al.* 1993). The Anatolides–Taurides Block (ATB) exhibiting the Gondwana affinity consists of different age and type of metamorphic and sedimentary rock units (Özgül 1984; Okay & Tüysüz 1999; Okay 2008). The ATB, Pontides and KM are separated each other by the İzmir–Ankara–Erzincan suture zone. The ATB is amalgamated with Arabian Plate along the Assyrian and Zagros sutures in the southeastern part of Turkey. The Isparta Angle (IA) is a triangular-shaped region, bounded by the African Plate to the south and the Anatolian Plateau to the north. The tectonic evolution of the IA is based on the rifting of the North African continent at early Mesozoic era (Poisson *et al.* 1984; Robertson & Dixon 1984; Şengör *et al.* 1984). The IA exhibits the magmatism from calc-alkaline to ultra-potassic affinity (Francalanci *et al.* 2000; Koçyiğit *et al.* 2000; Çoban & Flower 2006). The Arabian Plate is characterized mostly by sedimentary rock units accumulated during Cambrian–Miocene era. The Black Sea region has two different basins; the western and eastern Black Sea basins separated by a ridge segment. The western Black Sea Basin (WBB) consists of sedimentary basin overlain by oceanic to suboceanic crust and the eastern Black Sea Basin (EBB) consists of thinner sediments overlain by continental crust (Nikishin *et al.* 2003).

The tectonic frame of the eastern Mediterranean region has been controlled by three major tectonic processes since the Late Cretaceous; the convergence between Africa and Eurasia, the collision between terranes derived from Gondwana and Eurasia, and the subduction of oceanic lithosphere beneath Eurasia (McKenzie 1978; Dercourt *et al.* 1986; Gealey 1988; Westaway 1994; Stampfli & Borel 2004). The continental collision between Arabian and Eurasian plates leads to the westward escape of the Anatolian Plateau since Middle Miocene (McKenzie 1972, 1978; Şengör *et al.* 1985).

The western part of Turkey is under extensional tectonic regime as indicated by Global Positioning System (GPS) data and the tectonic features (e.g. Aegean graben system; Reilinger *et al.* 1997; McClusky *et al.* 2000; Reilinger *et al.* 2006). Widespread latest Oligocene–early Miocene magmatism in the Hellenic subduction zone has been attributed to extensional deformation and crustal weakening (Thomson & Ring 2006). The crustal thickness of western Anatolia has been reported about 25–35 km (Saunders *et al.* 1998; Horasan *et al.* 2002; Akyol *et al.* 2006; Zhu *et al.* 2006; Zor *et al.* 2006; Tezel *et al.* 2010). The cause of the extension in the region is still a controversial issue. Some researchers propose the collision-related tectonic escaping model (Dewey & Şengör 1979; Şengör 1979; Taymaz *et al.* 1991; Jackson *et al.* 1992), in which the extension of the region is supposed to be caused by the westward motion of the Anatolian Plateau along the North Anatolian Fault (NAF) and East Anatolian Fault (EAF) since Middle Miocene (Bozkurt 2001). The orogenic collapse model (Seyitoğlu *et al.* 1992; Seyitoğlu & Scott 1996) supposes that the extension of the Aegean region is associated with the gravitational collapse and thinning of the thickened crust because of the closure of the Neotethys Ocean (Meulenkamp *et al.* 1988; Jolivet 2001; Faccenna *et al.* 2003; van Hinsbergen *et al.* 2005). The subduction roll-back model (McKenzie 1978; Angelier *et al.* 1982; Thomson *et al.* 1998; Gautier *et al.* 1999; ten Veen & Postma 1999; Reilinger *et al.* 2006) proposes that the roll-back movement of the Hellenic slab causes the backarc extension in western Anatolia. This movement of the Hellenic slab has been determined by the steeply dipping Mediterranean oceanic lithosphere underneath the Aegean continental lithosphere in former tomography models (Spakman *et al.* 1988; Faccenna *et al.* 2003; Piromallo & Morelli 2003; Faccenna *et al.* 2006).

The central part of Turkey has flatter topography than western and eastern Anatolia, and this area can be described as wedge-like region between NAF and EAF. In this region, there are several strato-volcanoes, which are considered as post-collisional (<13 Ma) (Pasquare *et al.* 1988; Notsu *et al.* 1995).

The eastern part of Turkey is one of the highest plateaus of the Alpine–Himalayan Mountain system, and has undergone crustal shortening and thickening due to the collisional events during Late Eocene to Oligocene. The eastern part of Turkey is under compressional tectonic regime as inferred from GPS measurements and the tectonic features (e.g. Eastern Anatolian Accretionary Complex (EAAC) with high topography, McClusky *et al.* 2000; Reilinger *et al.* 2006). Its average elevation is approximately 2 km above the sea level (Şengör & Kidd 1979). There are several different models to explain geodynamic evolution of eastern Anatolia, which have been summarized by Keskin (2005) as follows:

- (1) tectonic escape of microplates (McKenzie 1972, 1976; Şengör & Kidd 1979; Jackson & McKenzie 1988),
- (2) subduction of the Arabian Plate beneath the eastern Anatolia (Rotstein & Kafka 1982),
- (3) slab break-off and the northward movement of a subducting slab beneath the eastern Anatolia (Innocenti *et al.* 1982),
- (4) rifting along E–W directed basins (Tokel 1985) and melting of the asthenosphere due to extension (McKenzie & Bickle 1988),
- (5) the crustal thickening model (Dewey *et al.* 1986),
- (6) lithospheric delamination model (Pearce *et al.* 1990; Keskin *et al.* 1998), and
- (7) slab steepening and detachment beneath an EAAC (Keskin 2003; Faccenna *et al.* 2006).

Recent investigations indicate anomalously thin crust and lithosphere (Zor *et al.* 2003; Angus *et al.* 2006; Özacar *et al.* 2008), strong Lg and Sn attenuation (Gök *et al.* 2003), slow wave speed anomalies of S wave (Villasenor *et al.* 2001; Maggi & Priestley 2005; Gök *et al.* 2007) and Pn wave (Al-Lazki *et al.* 2003, 2004) in this region. Seismic tomography models indicate slow wave speed anomaly down to the mantle transition zone (Piromallo & Morelli 2003; Faccenna *et al.* 2006). Previous geological models indicate that the slab was presumably steepened and was finally broken-off beneath the EAAC around 10–11 Ma (Keskin 2003; Şengör *et al.* 2003). Şengör *et al.* (2003) suggest that the lack of the mantle lithosphere is related to the slab break-off and the widespread volcanism exists due to direct contact with hot asthenosphere. Şengör *et al.* (2008) consider that eastern Anatolia was occurred as subduction–accretion complex, covered by at least 15 000 km³ of volcanic rocks.

2 DATA AND SURFACE WAVE DISPERSION MEASUREMENTS

2.1 Two-station method

We employ the conventional two-station method (Dziewonski & Hales 1972) to measure average phase speeds of the fundamental-mode Rayleigh waves between a pair of stations, following the recent work by Yoshizawa *et al.* (2010). The two-station method is one of the most classical but useful techniques to determine surface wave phase speeds for regional- or local-scale studies. This method has been carried out to construct 3-D upper-mantle structures for many regional-scale studies (e.g. Isse *et al.* 2006; Yao *et al.* 2006; Yoshizawa *et al.* 2010).

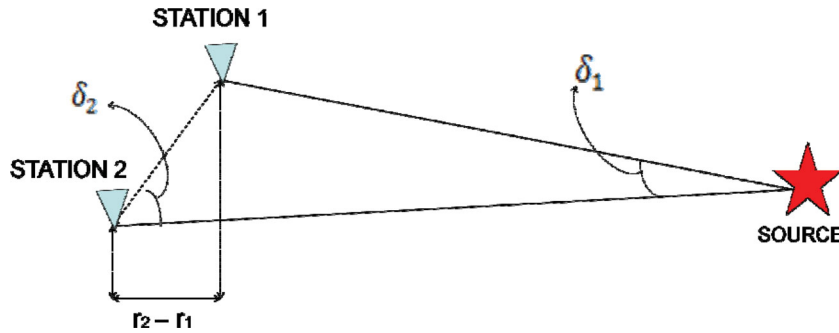


Figure 2. An illustration of geometrical configuration of the two stations and seismic source.

In this method, a station-pair is supposed to be located on or near a common great circle. The conditions of the station-pair selection are illustrated in Fig. 2. In this figure, δ_1 shows the difference of azimuths from a source to the two stations, which is selected to be less than 0.5° , δ_2 indicates the difference between backazimuths from the stations 2 to 1 and to the seismic source, which is set to be less than, $\delta_2 \leq 5.5^\circ$.

We determine the phase speed dispersion curve between two stations by computing the phase differences (Dziewonski & Hales 1972). Average phase speed between the pair of stations is given as

$$C(\omega) = \frac{\omega(r_2 - r_1)}{\phi_2 - \phi_1 + 2l\pi}, \quad (1)$$

where $C(\omega)$ is phase speed as a function of frequency, r_1 and r_2 are distance between the epicentre and stations 1 and 2, respectively, ϕ_1 and ϕ_2 are unwrapped phase function of the stations 1 and 2, respectively, and l is an integer that controls 2π ambiguity of unwrapped phase cycles.

2.2 Seismic data and their processing

We analysed 289 seismic events with moment magnitude 5.5 and greater with focal depth shallower than 100 km, located at epicentral distances longer than 2000 km in the period between 2006 January and 2008 December (Fig. 3). We used 164 three-component broadband stations [99 Boğaziçi University Kandilli Observatory and Earthquake Research Institute (KOERI), 59 Prime Minister Disaster and Emergency Management Presidency Earthquake Research Department (ERD), two IRIS and four GEOFON in Fig. 4]. The waveforms are provided from two national seismic networks operated by KOERI and ERD. First, we selected the waveforms carefully by checking their quality (e.g. without time gaps) and availability of two stations that are located on (or near) a common great circle path. Before proceeding to the two-station analysis, all waveforms are initially corrected with an instrument response of each station since the networks include different types of seismometers (e.g. CMG-3T, 3TD, 3ESPD, 6T, 40T, STS1, STS2) with different eigenperiods. We deconvolved the instrument response function of each seismometer, and then convolved with that of CMG-3T type seismometer with the eigen period of 120 s, which is the most widely used seismometer in our seismic network. We then extracted the fundamental-mode Rayleigh waves in a time window with the limits defined by group velocities of 2.6 and 5.1 km s⁻¹, for which the phase functions are calculated as a function of frequency. Based on the great circle approximation, the average phase speeds between two stations for each measurement are calculated using eq. (1).



Figure 3. Distribution of 289 seismic events used in this study (circles).

The phase difference term in eq. (1) can also be derived from cross-correlation between the seismograms at two stations, though the dispersion curves obtained by the phase difference and the cross-correlation are generally very similar. We can obtain satisfactory results using both methods, if the data quality is good enough with only minor effects from scattered or off-great-circle signals, and if the length of the path between two stations is long enough with respect to the wavelength.

2.3 Observed dispersion data

All dispersion curves are visually checked to choose a proper period range in which the observed dispersion curve is smooth enough with smaller variance. In Fig. 5, we display two examples of dispersion curves of Rayleigh waves measured for two station pairs: AGRB-KEMA and VANB-KEMA stations. In these examples, we selected the phase speeds in a period range between 20 and 65 s for AGRB-KEMA (Fig. 5a), and between 40 and 120 s for VANB-KEMA (Fig. 5b).

We usually observed that phase speed measurements for two closely located stations exhibit significant errors in the longer period for which the wavelength is close to the propagation distance (Fig. 5a) On the contrary, for the excessive long distance paths

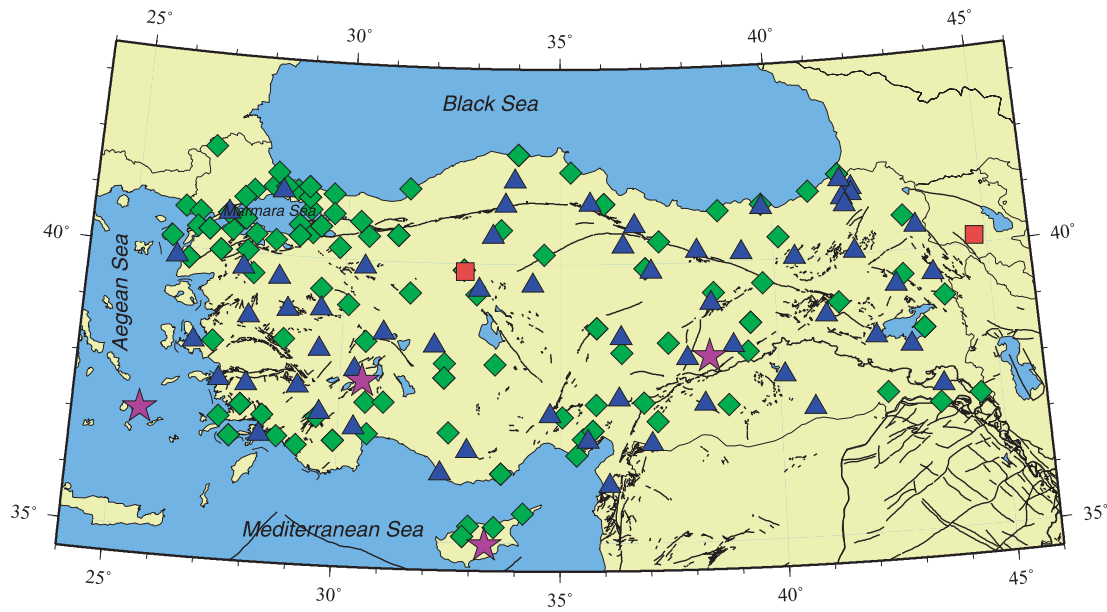


Figure 4. Locations of broad-band seismic stations (triangles: ERD stations, diamonds: KOERI stations, squares: IRIS stations, stars: GEOFON stations, thin lines: National borders, thick lines: known faults).

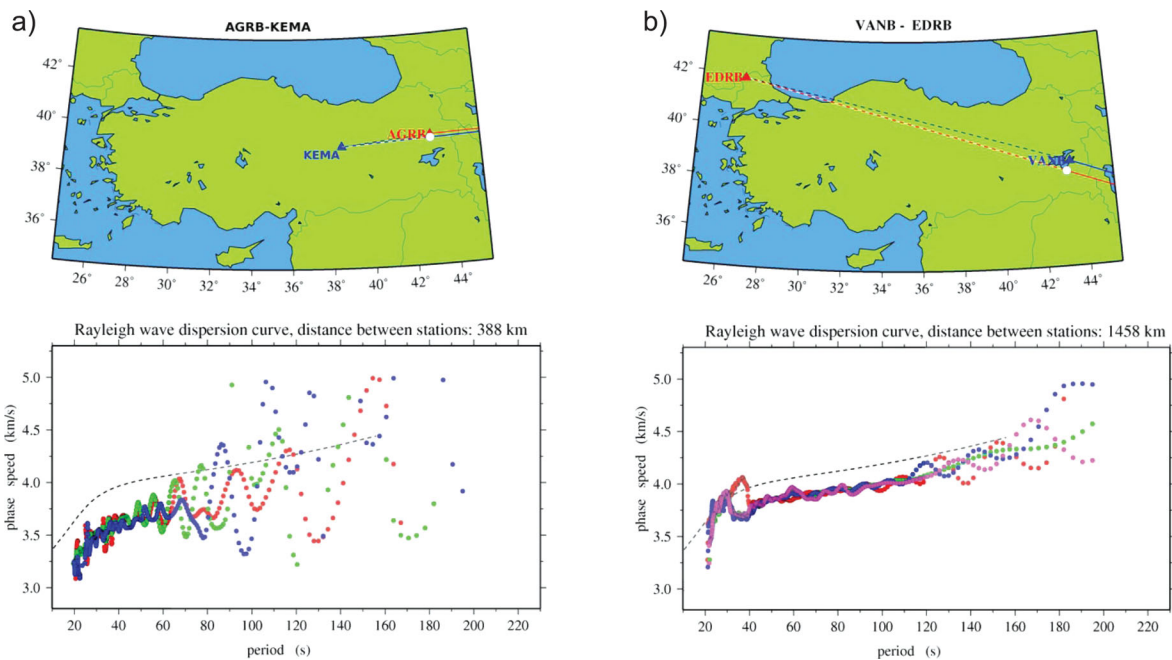


Figure 5. Two examples of observed dispersion curves. (a) Two-station paths between AGRB and KEMA stations (top panel), measured dispersion curves of Rayleigh waves for paths between AGRB and KEMA stations (bottom panel), white circle: projected new point of the near station on great circle for the further station (AGRB), dashed curve (bottom panel): the reference model (PREM + 3SMAC), grey area (bottom panel): discarded period range, dashed curve (bottom panel): the reference model (PREM+3SMAC). (b) Same as (a) but for paths between stations VANB and EDRB.

between two stations, observed dispersion curves in the shorter period tend to be affected significantly by 2π ambiguities in phase cycles (Fig. 5b). The stability and reliability of the two-station measurements depend on the distance between the stations and the wavelength of surface waves to be considered. Moreover, the influence of non-plane waves, which propagate with a variety of amplitudes and directions, on our two-station measurements may not be neglected. The energy of the non-plane waves is strongly affected by diffrac-

tion caused by the lateral heterogeneities in the structure (Pedersen 2006).

Fig. 6 shows the average dispersion curve of the fundamental-mode Rayleigh waves, derived from all the phase speeds for all station pairs in the period range from 20 to 160 seconds measured in this study. The phase speed dispersion curve (open circles) is significantly slower than the reference dispersion curves (dashed lines). The reference dispersion curve is obtained by using PREM

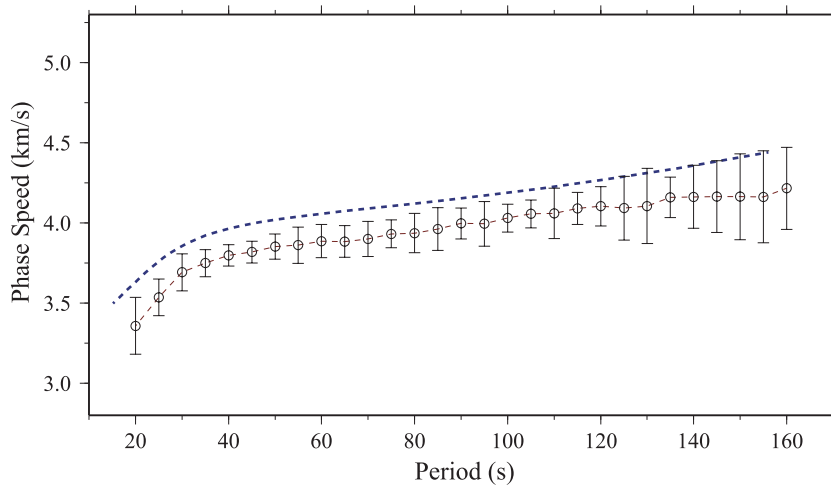


Figure 6. Average dispersion curves and standard errors for Rayleigh waves, dashed line: reference phase speed for a continental reference model with a correction for average crustal model of Turkey (PREM and 3SMAC), circles and error bars: average phase speed of Turkey (this study).

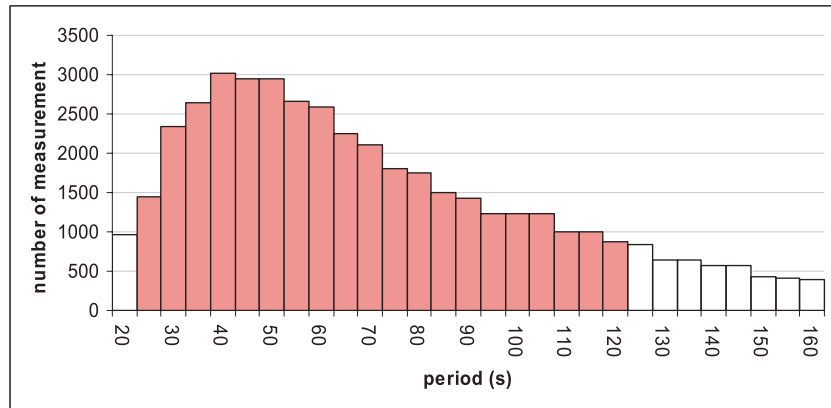


Figure 7. Number of measurements of two-station phase speeds of Rayleigh wave as a function of period. Filled columns represent the data used in subsequent phase speed mapping, and the blank columns indicate the data unused for phase speed mapping in this study.

(Dziewoneski & Anderson 1981) and 3SMAC models (Nataf & Ricard 1996). The crust of PREM is corrected by using the 3SMAC model to better represent the average crustal structure of Turkey. The period range of the reliable dispersion measurements with reasonable standard errors is observed between 25 and 120 seconds. Our results are consistent with previous global and large-scale regional tomographic studies, in which the average velocity of entire Turkey tends to be significantly slower than the global average (Ekström *et al.* 1997; Debayle *et al.* 2001; Villasenor *et al.* 2001; Maggi & Priestley 2005).

The numbers of measurements of Rayleigh wave phase speeds are plotted as a function of period in Fig. 7. The large number of the stable and reliable measurements has been obtained in the period between 25 and 120 s, which will be used in the subsequent phase speed mapping.

3 INVERSION FOR PHASE SPEED MAPS

3.1 Inversion method for phase speed mapping

To obtain Rayleigh wave phase speed maps as a function of period, we employ the method of tomographic inversion developed by

Yoshizawa & Kennett (2004). The principles of method are briefly summarized in this section.

Observed phase speed perturbations $\langle \delta c / c_0 \rangle$ can be represented by the following linear relationship:

$$\left\langle \frac{\delta c}{c_0} \right\rangle^{\text{obs}} = \frac{1}{\Delta} \int_{\text{ray}} ds \frac{\delta c(s)}{c_0}, \quad (2)$$

where δc is observed phase speed perturbation, s the coordinate along the path, c_0 a reference phase speed and Δ the interstation distance. Phase speed maps are obtained based on the approximation of surface wave propagation along the ray path between stations using the relationship (2). To obtain phase speed perturbation on a sphere, we use spherical B-spline functions $F(\theta, \phi)$ (Lancaster & Salkauskas 1986; Wang & Dahlen 1995), which are defined at rectangular geographical knot points. The knot interval used in this study is 0.75° .

The linear equation is solved using the damped least-squares scheme with the LSQR algorithm (Paige & Saunders 1982)

$$\begin{bmatrix} \mathbf{G} \\ \lambda \mathbf{I} \end{bmatrix} \mathbf{m} = \begin{bmatrix} \mathbf{d} \\ \mathbf{0} \end{bmatrix}, \quad (3)$$

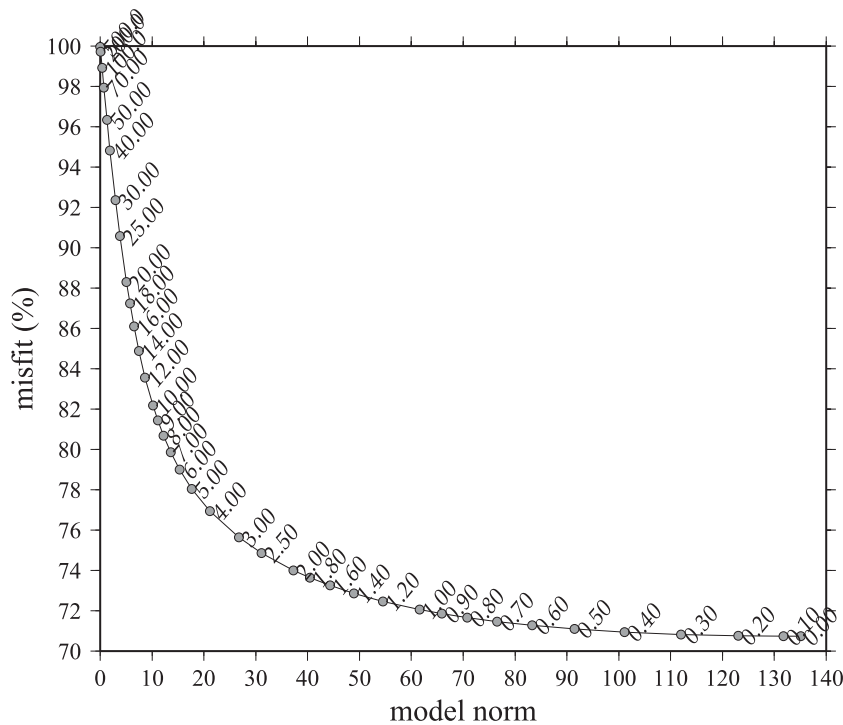


Figure 8. A trade-off curve for different values of damping parameter λ .

where $\mathbf{d}$ is the data vector comprised of observed phase speed variations, $\mathbf{G}$ the kernel matrix, $\mathbf{m}$ the model vector, $\mathbf{I}$ identity matrix and λ an arbitrary damping parameter that controls the trade-off between the model variance and resolution. The details of our method are explained by Yoshizawa (2002) and Yoshizawa & Kennett (2004).

In this study, the damping parameter for each phase speed maps is determined from a trade-off curve between the misfit norm and the model norm. An example of the trade-off curve for 40 s phase speed model is displayed in Fig. 8. Damping parameters λ in this example varies in a wide range from 0 to 500, and we selected $\lambda = 2.5$ as a preferred damping value. Data misfit is calculated by $[100 - \text{variance reduction (per cent)}]$, where the variance reduction is defined as $(1 - |\mathbf{d} - \mathbf{Gm}|^2 / |\mathbf{d}|^2) \times 100$ (per cent). We achieved the variance reductions between 20 and 42 per cent for our phase speed models, depending on periods of surface waves.

3.2 Horizontal resolution analyses

Checkerboard tests are performed to visually examine the effect of path coverage on the final model. The input checkerboard models contain ± 5 per cent perturbations from the reference phase speed (c_0) for each frequency.

We employ a variety of sizes of cell patterns to evaluate the spatial resolution of our phase speed models. The results of checkerboard tests at 40 and 80 s with cell sizes of 1.0° , 1.5° and 2.0° are displayed in Fig. 9. These tests indicate that our data set generally provides us with a good resolution for the structure with the scale-length of 1.5° or larger beneath Turkey. The horizontal resolution tends to be deteriorated at longer period that sample deeper part of the mantle, due to the smaller numbers of available paths.

All retrieved patterns indicate that the western part of Turkey (the Marmara and the Aegean region) has been better resolved owing to denser ray coverage. On the contrary, the eastern part of Turkey (the

Middle and the east Anatolian region) shows somewhat smeared patterns because of relatively limited numbers of crossing paths, and the northern part of Turkey (Black Sea region) shows lower resolution due to poor ray coverage. In spite of the smeared patterns in the eastern and northern areas, the numbers of paths used in our inversion enable us to attain satisfactory resolution of tomographic models that allows us to discuss the large-scale structure beneath Turkey.

The resolution of the tomography models strongly depends on the coverage of the paths. The number of ray paths counted in $0.75^\circ \times 0.75^\circ$ cells for Rayleigh waves at 40 and 80 s are plotted in Fig. 10. The cells with greater numbers of hits are located in the centre of Turkey and the number of the paths in a cell is decreasing near borders of our target region because of the insufficient number of seismic stations especially in the southern and eastern areas of Turkey.

3.3 Phase speed maps

The phase speed maps are obtained in the period range between 25 and 120 s. Four examples of the phase speed maps are shown in Fig. 11. All maps are represented as perturbations from the average phase speed c_0 indicated in each map.

We observe widespread fast phase speeds in the western part of Turkey (the Aegean and the Marmara region) in short-period phase speed maps (< 65 s). Remarkably, fast phase speed anomalies can be seen in the longer period (> 65 s) beneath southern Turkey associated with subduction of the African lithosphere. In the south-eastern part of Turkey, a contrast between fast and slow phase speed anomalies is notable across the northern border of the Arabian Plate (Zagros Suture zone) in the shorter period (< 50 s). We can identify the slow phase speeds beneath the volcanoes in the central part of Turkey at periods shorter than 45 s. An extremely slow phase speed in the eastern part of Turkey reflects EAAC.

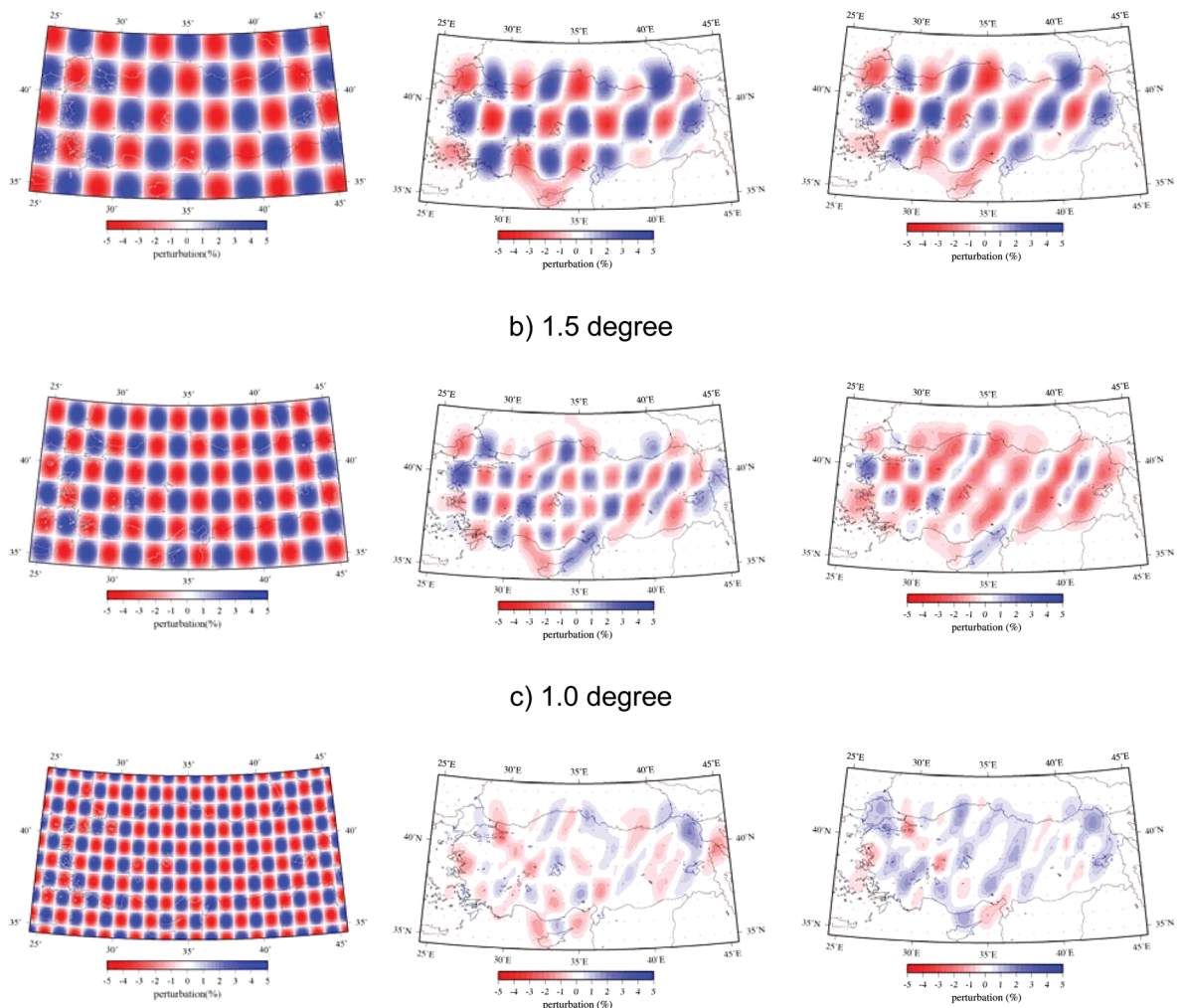


Figure 9. Examples of checkerboard resolution tests using the path coverage at 40 and 80 s with a variety of cell sizes; (a) 2.0° cells, (b) 1.5° cells and (c) 1.0° cells. (Left-hand side) Input checkerboard patterns, (centre), retrieved checkerboard patterns for 40s, (right-hand side) same as centre column but for 80 s.

4 INVERSION FOR S-WAVE SPEED MAPS

3-D S-wave speed maps are derived from the ensemble of phase speed maps. At first, we extract local phase speed information at every $0.5^\circ \times 0.5^\circ$ grid from phase speed maps in a period range from 25 to 120 s. Local phase speeds are then used to invert for a local 1-D S-wave speed profile, which are eventually assembled to form a final 3-D S-wave speed model.

4.1 Inversion method for local S-wave speed profile

To obtain local S-wave speed profiles, we perform the method employed by Yoshizawa & Kennett (2004), which is based on generalized non-linear least-squares inversion of Tarantola & Valette (1982). Considering generalized relation of inversion $\mathbf{d} = \mathbf{g}(\mathbf{p})$, where $\mathbf{d}$ is a data vector that includes a set of local phase speed perturbation, $\delta c(\omega)$, as a function of frequency ω . The model parameter vector $\mathbf{p}$ involves a local S-wave speed perturbation, $\delta\beta(r)$, as a function of radius r . We use the classical linearized relation between the perturbation of a local phase speed and an isotropic local 1-D structure (e.g. Takeuchi & Saito 1972; Dahlen & Tromp 1998). The fundamental-mode surface waves in the period range used in this study are far more sensitive to the perturbations of shear wave speed than those of P-wave speed and density. We therefore consider

only the perturbations of shear wave speed, and we fixed P-wave speed and density to the reference model in this study.

In this inversion, the amplitude of the S-wave speed perturbation and the smoothness of the model are managed by two *a priori* parameters; a standard deviation (σ) and a correlation length (L), which are defined for estimating a model covariance matrix. Through some empirical tests, we chose $\sigma = 0.05 \text{ km s}^{-1}$ and $L = 5 \text{ km}$ above the Moho, and $L = 12 \text{ km}$ below the Moho, so that the S-wave speed perturbations in the mantle vary smoothly, whereas those in the crust can vary rapidly. A local reference model, which is used to initiate inversion for a local 1-D model, is created by the combination of PREM and the local average crustal structure calculated from the 3SMAC model.

The inversion of the observed dispersion curves for a 1-D S-wave speed model is generally non-unique. To confirm the stability of the S-wave speed models, we summarize an example of test inversions (Fig. 12) with dispersion data at a location of 39.5°N and 36.5°E employing four different initial models in addition to the local reference model (PREM + 3SMAC). We construct four different initial S-wave speed models by perturbing our local reference S-wave model with ± 1.5 per cent and ± 4.0 per cent. Fig. 12 displays the resultant 1-D S-wave speed models derived from the inversions, estimated dispersion curves and sensitivity kernels. We calculate the sensitivity kernels of phase speeds for the fundamental-mode Rayleigh waves with respect to the shear wave speed ($\delta c(\omega)/\delta\beta(r)$) by using the original local reference model at 39.5°N and 36.5°E ,

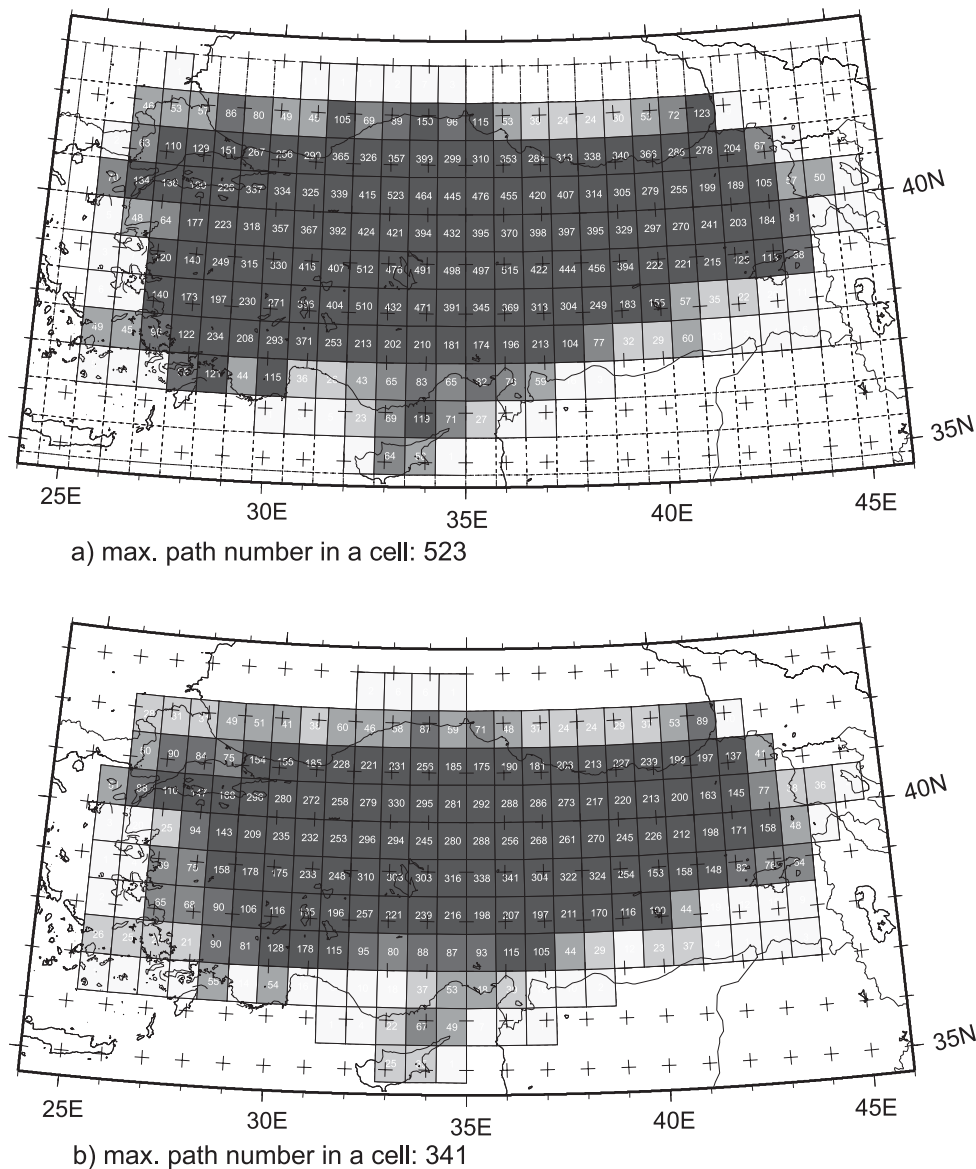


Figure 10. Density plot of the number of ray paths counted in $0.75^\circ \times 0.75^\circ$ cells for (a) 40 s and (b) 80 s.

and the maximum amplitude of sensitivity kernels for each period are normalized to be unity.

In Fig. 12, we recognize a trade-off between the crustal and mantle wave speeds, although we could achieve the same level of satisfactory fit between observed and calculated dispersion curves. This fact indicates intrinsic non-uniqueness in the determination of 1-D *S*-wave profile. The slight variations of the resultant 1-D *S*-wave models in Fig. 12 can be considered as a representation of the plausible range of uncertainties in our final *S*-wave model, which is normally less than about 2 per cent in the uppermost mantle.

4.2 Vertical resolution

An example of the vertical resolution analyses is displayed in Fig. 13. We test the vertical resolution using synthetic 1-D *S*-wave models (dashed red line in lower panels). The synthetic models are created by adding 10 per cent perturbation to a local reference *S*-wave speed model in narrow depth ranges (40–70, 70–100, 100–130 and 130–160 km). Synthetic dispersion curves (red open

circles in upper panels) are calculated using the synthetic 1-D *S*-wave models. Assuming the synthetic dispersion curve as observed data, we performed the inversion with the same initial model and parameters as the real-data inversion. The retrieved peak anomaly is in a very good agreement with synthetic *S*-wave model in shallow depths above 100 km. The obtained *S*-wave profiles (green line in lower panels) tend to be smeared with increasing depth, primarily due to the long wavelength of surface waves and the broadened shape of the sensitivity kernels at longer periods. Although we must be careful about such intrinsic limitations in surface wave inversions due to the smearing effects, the retrieved models indicate that we can reconstruct the large-scale features of the *S*-wave structure down to the depth of about 150 km.

5 THREE-DIMENSIONAL *S*-WAVE SPEED MODEL

We construct the final 3-D *S*-wave model using the ensemble of the local 1-D *S*-wave speed models of the entire region of Turkey.

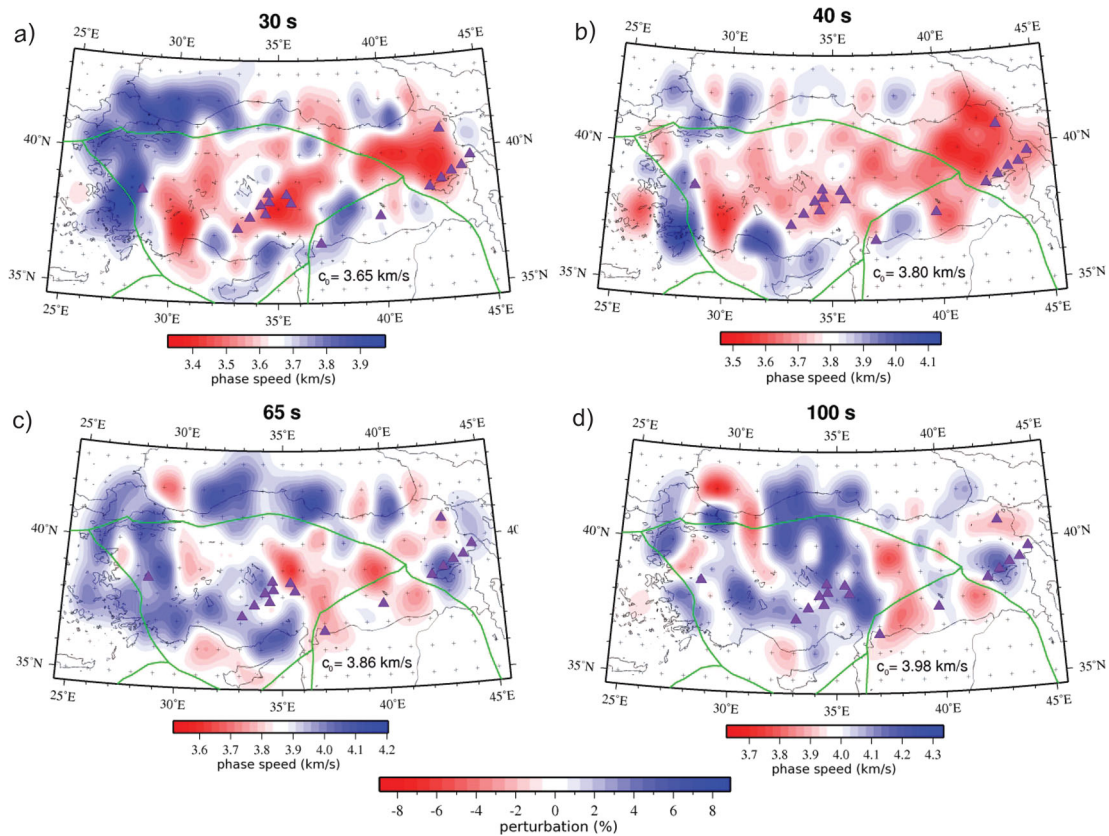


Figure 11. Examples of the phase speed maps of the fundamental-mode Rayleigh waves at (a) 30 s, (b) 40 s, (c) 65 s and (d) 100 s.

We plot the horizontal slices at various depths and the vertical cross-sections across the distinct tectonic features (Figs 14–16). All the maps are plotted as perturbations from an average S -wave speed (V_0) model shown in each map. Locations of volcanoes and plate boundaries are plotted on all maps, and the focal depths of earthquakes are also displayed in the cross-sections. Hypocentre locations with magnitude 3.0 and greater are provided by the ISC catalogue between 1900 and 2008.

One of the most prominent features of shallower depths in the horizontal slices (50–80 km) is the fast S -wave speed anomalies in the western part of Turkey (Fig. 14). It should be noted that the S -wave speeds are generally faster in the west and becomes slower in the east of Turkey. Another conspicuous slow anomaly is observed in the north of western Turkey (the eastern Marmara region approx. 42°N–29°E). This circular shaped slow anomaly can be seen below 80 km depth (Fig. 14). Some major features are discussed in detail in the next section.

In the south of western Turkey, a significant slow anomaly is recognized around the IA and Antalya Basin. This slow anomaly is extending down to approximately 70–80 km depth (Figs 15A–A', B–B' and D–D'). A robust fast anomaly is noted in the north of Cyprus (40–140 km) associated with the northward dipping African oceanic lithosphere subducted from the Cyprus trench. This fast speed anomaly can be seen in vertical cross-sections C–C' and E–E' in Fig. 15. The fast anomaly is separated into two parts in the northeast and southwest at around 37°N and 34°E.

In the south of eastern Turkey, we can recognize the Arabian and Eurasian collision zone (Assyrian and Zagros sutures) with a large S -wave speed contrast above 60 km depth (Fig. 14a). We can see a slow anomaly in northeastern Turkey beneath EAAC

(Figs 14 and 16H–H' and I–I'). A fast speed anomaly appears below 80 km depth in the north of Zagros suture (Figs 14, 16F–F' and G–G'). Note that, though the NE–SW smearing effects are seen in the checkerboard test in Fig. 9 with the smaller patterns than 1.5°, the observed anomalies are greater than the scale length of 1.5°, which justifies that these anomalies are not likely to be artefacts.

6 DISCUSSIONS

A 3-D S -wave speed model of the upper mantle beneath Turkey is constructed from phase speeds of the fundamental-mode Rayleigh waves in the period range from 25 to 120 s, which are primarily sensitive to the depth range between 40 and 180 km, in the uppermost mantle. Such intermediate to long-period surface waves may indicate relative variations in the thickness of the crust, but are not very sensitive to the detailed crustal structures and shallow faults, which may be better constrained by ambient noise approach (e.g. Shapiro & Campillo 2004) that is out of our scope. Thus, our following discussions will mainly focus on striking features in the uppermost mantle beneath Turkey using our 3-D model in comparison with some previous studies.

6.1 Western Turkey

In the shallower depth of the upper-mantle structure of Turkey (Fig. 14a), the fastest wave speeds are observed in broad areas in the western part (the Aegean and the Marmara regions). These fastest speeds are likely to represent the mantle structure just

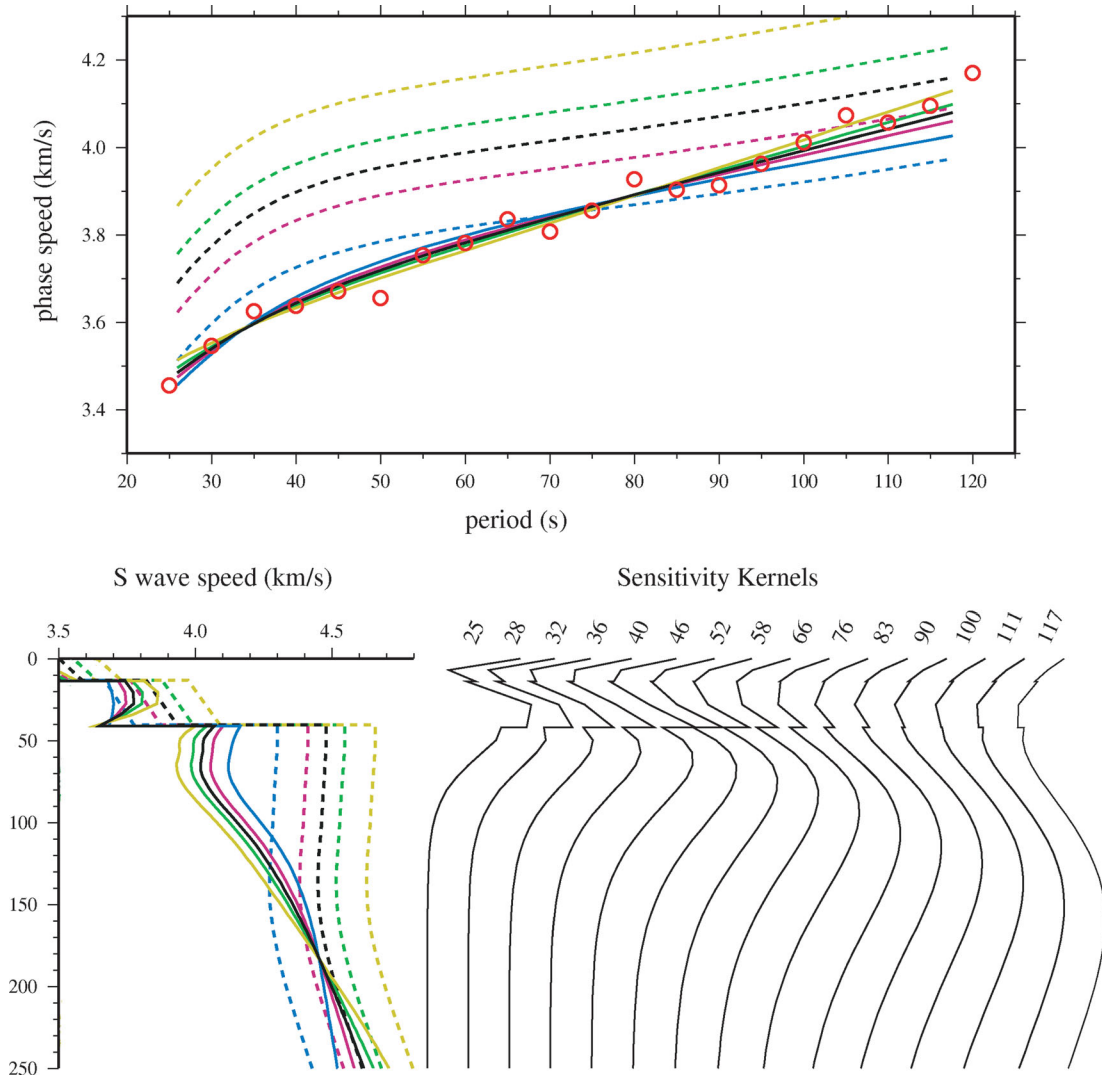


Figure 12. An example of inversion for 1-D S -wave models using different initial models. Phase speed dispersions (top panel), sensitivity kernels (bottom right panel), S -wave models (bottom left panel). Circles: observed dispersions, dashed lines in bottom left panel indicate initial S -wave models; solid lines retrieved S -wave models. Dashed lines in the top panel represent dispersion calculated for the initial S -wave speed models, and solid lines dispersion curves for the retrieved S -wave models. Line styles denote the variation of models; black/plain: a local reference model (PREM + 3SMAC), blue/squares: -4 per cent perturbations from the reference model, magenta/triangles: -1.5 per cent, green/diamonds: $+1.5$ per cent and yellow/stars: $+4$ per cent (the line styles used differ in the print and online versions of the article).

beneath the thin crust of this region whose thickness has been estimated to be around 25–35 km (Saunders *et al.* 1998; Horasan *et al.* 2002; Akyol *et al.* 2006; Zhu *et al.* 2006; Zor *et al.* 2006; Tezel *et al.* 2010), mainly due to the extensional regime of the Aegean region (McKenzie 1978; Le Pichon & Angelier 1979; Jackson 1994; Reilinger *et al.* 1997; McClusky *et al.* 2000; Reilinger *et al.* 2006).

A large-scale fast anomaly is observed beneath the İZ (41.5°N–34°E). This anomaly terminates at intra-Pontide suture zone between the İZ and SZ. This anomaly can be related to the different structure of the İZ, which is a distinctive Pontide zone than the others, consisting of different types of older rocks (Yiğitbaş *et al.* 2004; Şengör *et al.* 2005; Okay 2008). Biryol *et al.* (2011) interpreted that this anomalies reflect the deformation of NAF penetrated into the uppermost mantle, although this interpretation is controversial considering shallow seismicity of NAF (i.e. Honkura

et al. 2000; Ambraseys 2002; Barış *et al.* 2002; Lettis *et al.* 2002; Özalaybey *et al.* 2002; Ben-Zion *et al.* 2003).

6.2 Central Turkey

One of the most significant features of our model is the fast anomaly located in the north of Cyprus. We interpret the anomaly as a segment of the subducted African oceanic lithosphere from the Cyprus trench (Hereafter, referred to as ‘Cyprus slab’). The fast speed anomaly is also visible in P -wave tomography models of Koulakov *et al.* (2002), Piromallo & Morelli (2003) and Biryol *et al.* (2011). We can see a discontinuity of the Cyprus slab between Karaman and Adana cities (approx. 37°N–34°E) at approximately 60–100 km depth (Figs 14c and 15A–A’). The tear of the Cyprus slab also appears in the P -wave tomography model of Biryol *et al.* (2011).

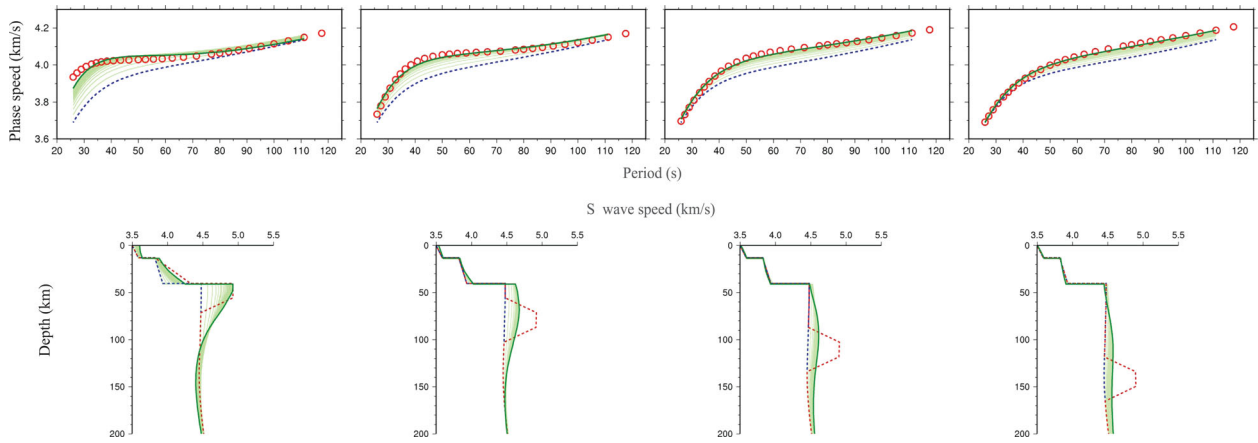


Figure 13. An example of vertical resolution test for 1-D S-wave speed model at $39.5^{\circ}\text{N} \times 36.5^{\circ}\text{E}$ (In the top panels, (blue) dashed line: reference dispersion curve (PREM + 3SMAC); open circles: synthetic dispersion curve; light (green) solid lines: dispersion curves for each iteration's inversion; (green) solid line: dispersion curve of retrieved S-wave model. Dispersion curves are calculated for the corresponding 1-D S-wave speed models. In the Sidebottom panels: (blue) dashed line: reference 1-D S-wave model (PREM + 3SMAC), starred/red dashed line: synthetic S-wave models, light (green) solid lines: inversion results of each iteration for 1-D S-wave model and (green) solid line: retrieved S-wave model).

They suggested that the minor tearing starts above 200 km and terminates around 250 km. In this study, we cannot observe the termination of the minor tearing due to the limited depth resolution of our model.

The eastern part of the Cyprus slab extends N–S and NE–SW direction up to 37°N latitudes beneath Niğde and Karaman cities and terminates beneath the quaternary volcanoes which appear as the slow anomalies (Figs 14 and 15C–C' and E–E'). The western part of the Cyprus slab terminates at the edge of the IA (Figs 15A–A', B–B' and D–D') where a slow anomaly appears. In the south of the IA, the African oceanic lithosphere in the Mediterranean Sea, subducting in the eastern Hellenic trench towards the NW and in the Cyprus trench towards the NE, has been shown by tomographic and seismicity studies (Piromallo & Morelli 2003; Boschi *et al.* 2004). Existence of the major tear between the two slabs also identified by some former studies (Wortel & Spakman 1992; Barka & Reilinger 1997; Wortel & Spakman 2000; Piromallo & Morelli 2003; Agostini *et al.* 2007; Dilek & Sandvol 2009; Biryol *et al.* 2011). The slow speed is very likely related to this major tear in our model at depth between 50 and 100 km. Previous geochemical investigations suggest that the presence of volcanic activities may be related with the major tear (Tokçar *et al.* 2005; Dilek & Altunkaynak 2009; Dilek & Sandvol 2009). This slow anomaly has also been shown in previous studies (Spakman *et al.* 1993; Piromallo & Morelli 2003; Agostini *et al.* 2007; Cambaz & Karabulut 2010; Biryol *et al.* 2011). In the south of the IA, our tomography model indicates that the tear probably ends between the Hellenic and Cyprus slabs and they simply merge forming the still continuous oceanic mantle lithosphere beneath the eastern Mediterranean Sea (Figs 14c and d and 15A–A' and D–D'). The convergence between the two slabs may cause the seismicity in the intermediate depth, appeared in the cross-sections in Figs 15 A–A', B–B' and D–D'.

6.3 Eastern Turkey

Our model indicates a significant contrast in shear wave speed between Arabian and Eurasian collision zone (Assyrian and Zagros sutures) down to 60 km depth (Fig. 14a). This contrast can be related to the differences of structures and ages between the Arabian Plate and Anatolian Plateau. The Arabian Plate is the most stable

portion in this region as a continental shield, whereas the eastern Anatolian Plateau is under compression and has weak structure as a subduction–accretion complex. This sharp speed contrast also appears in some tomographic studies (Al-lazki *et al.* 2003; Zor 2008; Cambaz & Karabulut 2010).

In the cross-sections in Fig. 16, the subducted Arabian mantle lithosphere (hereafter, referred to as 'Arabian slab') from Assyrian and Zagros sutures is clearly observed as fast anomalies, which have a discontinuity at about 70–100 km depth in the north of Zagros Suture. This discontinuity of the Arabian slab may indicate a detachment, which is also visible in some previous studies (Davies & von Blanckenburg 1995; Lei & Zhao 2007; Zor 2008), although Zor (2008) suggests the existence of the detached slab at 600 km depth. Davies & von Blanckenburg (1995) and Lei & Zhao (2007) also suggested the detached slab at a similar depth with our study. Lei & Zhao (2007) displayed cross-sections at almost the same position as H–H' and I–I' (Fig. 16), but down to 400 km depth, showing that the depth of the detached slab is deepened towards the west at 41°E longitude. We cannot image such detached slab in Fig. 16H–H' due to the limit of our resolvable depth. The slab is imaged at about 100 km east of the 41°E longitude in this study (Figs 16F–F', G–G' and I–I').

In the east of Turkey above 100 km depth, we can see an extensive slow anomaly related to widespread young Neogene and Holocene volcanism (<8 Ma) (Fig. 14). In this region, the fissure-fed mantle-derived alkaline volcanism was observed until Pleistocene (0.4 Ma) and the presence of alkali basalt has been observed as an evidence of lithospheric mantle material (Yılmaz *et al.* 1998; Yürür & Chorowiez 1998; Pearce *et al.* 1990). Based on such geochemical evidences, all of these basaltic features may indicate the existence of the lower lithospheric mantle. The slow speed anomaly is consistent with the results from Pn tomography (Al-Lazki *et al.* 2003, 2004; Al-Damegh *et al.* 2004; Gans *et al.* 2009), Sn tomography (Gök *et al.* 2003; Al-Damegh *et al.* 2004), surface waveform tomography (Villasenor *et al.* 2001; Maggi & Priestley 2005; Gök *et al.* 2007), and P-wave tomography (Piromallo & Morelli 2003; Lei & Zhao 2007; Zor 2008). The slow speed anomalies can be caused by changes in both temperature and composition as well as by the presence of partial melts within the upper mantle. Şengör *et al.* (2003) and Keskin (2003) suggest that the subducted slab

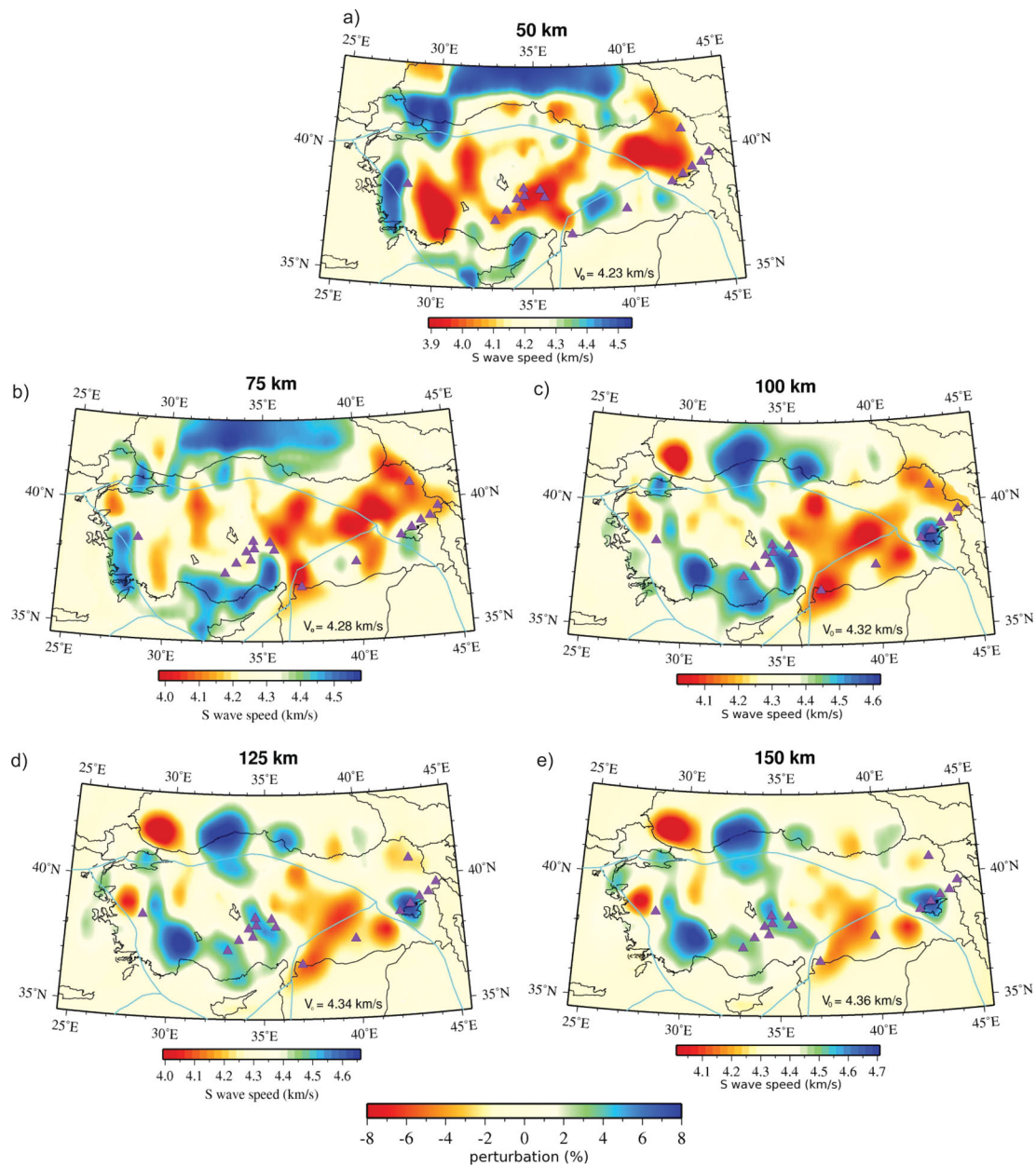


Figure 14. Examples of the horizontal slices of 3-D shear wave speed maps at (a) 50 km, (b) 75 km, (c) 100 km, (d) 125 km and (e) 150 km.

beneath the EAAC has broken off around 11 Ma and the bottom of the crust began to melt because of the direct contact with hot asthenosphere. As a result of sinking slab, the lithospheric mantle is either thinned or completely removed beneath the EAAC and has been started the widespread volcanism in Eastern Anatolia (Keskin 2007; Şengör *et al.* 2008).

7 CONCLUSIONS

We have constructed a new 3-D *S*-wave speed model of the uppermost mantle beneath Turkey using the two-station method for phase speed measurements of the fundamental-mode Rayleigh waves. With a high-density broad-band seismic network deployed throughout Turkey, the large-scale features including three major tectonic plates (Arabian, Eurasian and African plates) have been retrieved.

Our model clearly indicates three main tectonic features (Hellenic slab, Cyprus slab and collision of the Arabian Plate and Anatolian Plateau) and their tectonic and geodynamic implications. Even though the Hellenic trench remains outside of our study area, the influence of the backarc extension regime can be perceived in our model as a fast speed anomaly in the western Turkey. The crustal structure of Aegean Sea and its surroundings is thinned and formed as an extensional backarc basin due to subduction of the African lithosphere from the Hellenic trench. The Cyprus slab is separated into two parts at around 37°N–34°E, approximately 60–100 km depth by a minor tear. The eastern part of the Cyprus slab is extending to 37°N–38°N latitudes beneath Niğde and Karaman cities and terminates beneath the quaternary volcanoes. The western part of the Cyprus slab is extending under the IA and Antalya Basin and is probably closed with the

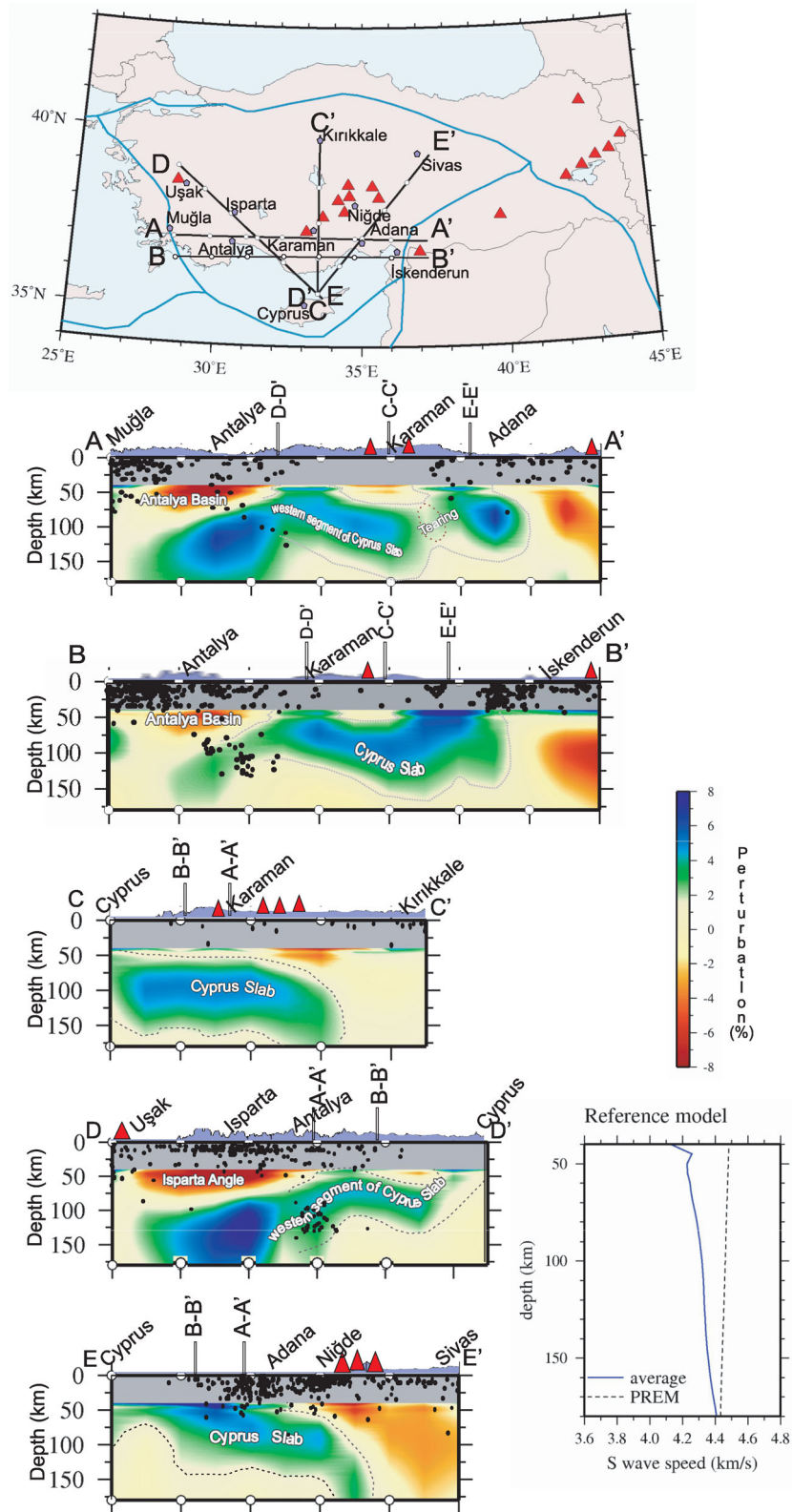


Figure 15. Vertical cross-sections across the Cyprus slab. At the top of each slice, elevations are displayed with 10 times exaggeration. White circles on the frame of each slice are plotted every 1° . Black circles on the slices are focal depth of earthquakes with magnitude ≥ 3.0 . Earthquake parameters are taken from the ISC catalogue in the period between 1900 and 2008. Bottom right panel: PREM and an average 1-D S -wave model which is used as a reference model for all cross-sections. Top panel displays locations of the cross-sections.

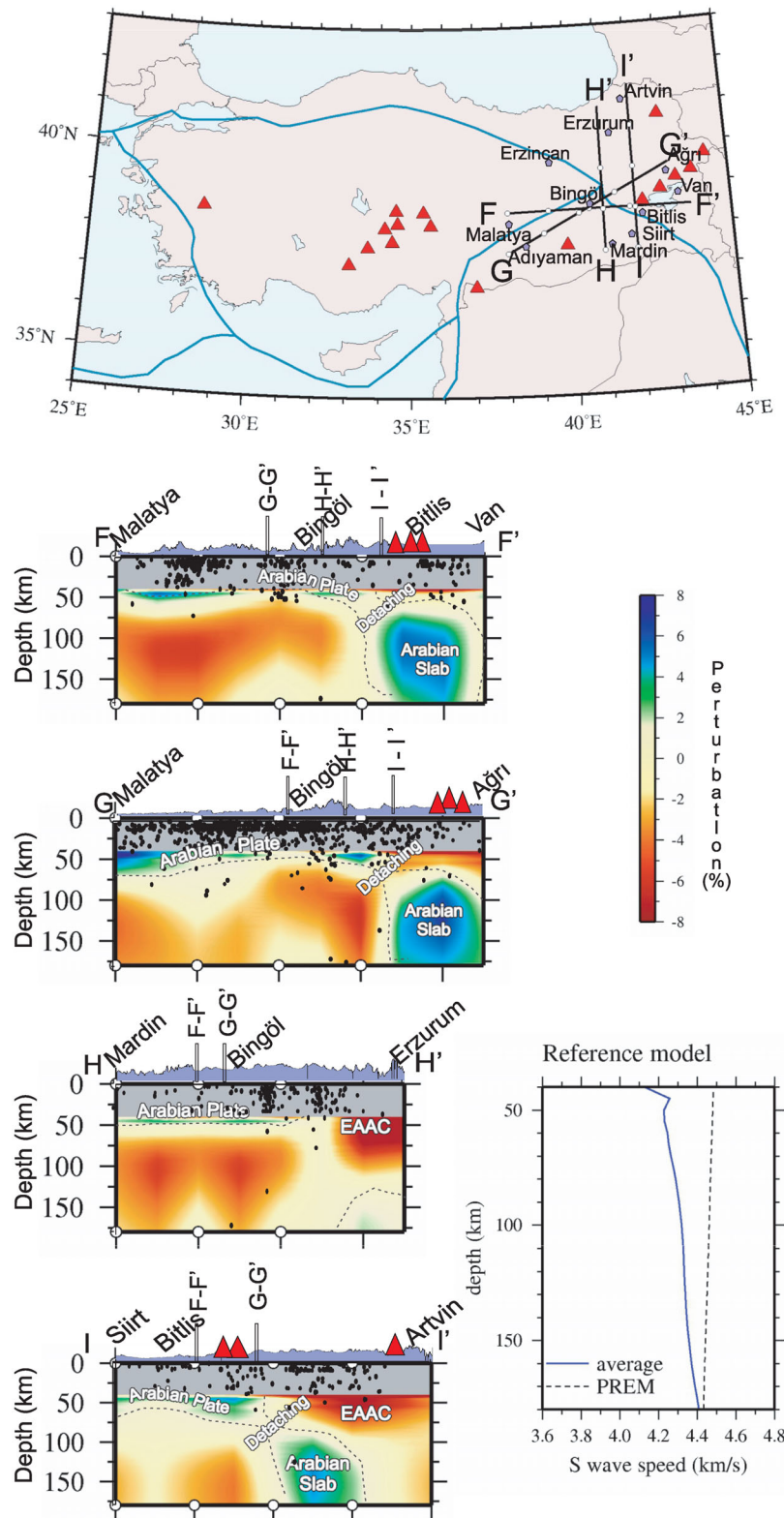


Figure 16. Same as Fig. 15, but for the Arabian slab.

Hellenic slab. We believe the tearing between the Hellenic slab and the Cyprus slab is likely to be closed and the slabs may simply be merged south of the IA. The slow anomalies under the IA and Antalya Basin are likely to be associated with rising hot materials

from asthenosphere due to major tearing between the Hellenic and Cyprus slabs. The Arabian slab appears to be broken-off under the Eastern Anatolia, which may cause the widespread volcanism and uplift in the Eastern Anatolia. We also observe the detached slab

around 39°N–41.5°E at 100 km depth. The detached slab is probably located in the deeper part of the mantle in the west of 41°E longitude.

The smaller scale features of the model have been argued in some regions where we have satisfactory resolution. The fast speed anomaly in the western Black Sea region (in the İZ Zone) can be related to the relatively thicker and older lithosphere of the western Pontides. A clear contrast in shear wave speed is observed along Assyrian–Zagros suture zone, which can be related with the structural difference between Anatolian Plateau and Arabian Plate.

Since the objective of this study is to construct the uppermost mantle structure beneath Turkey using surface waves in the intermediate to long-period range between 25 and 120 s, we have simply employed a reference 3-D crustal structure of the 3SMAC model for crustal correction. To better constrain the crustal structure of Turkey with higher horizontal resolution, it will be better to use alternative information that can be more sensitive to shallow structures; for example, surface wave group speed and ambient noise tomography. The combined use of the two-station phase speed measurements as well as the ambient noise and/or group speed analysis will allow us to better constrain both the crust and uppermost mantle structures.

The current method used in this study is based on the classical two-station method, but now we are able to constrain the local-scale features in the uppermost mantle working with the dense broad-band seismic networks. Although we have only employed the Rayleigh wave information, which tend to be more robust compared to Love waves in the horizontal components, it will be extremely helpful to map anisotropic properties (both azimuthal and radial anisotropy) that should directly reflect the current and historical tectonic processes and mantle dynamics beneath this region. Such an approach will require much denser ray coverage as well as a large number of Love wave measurements, which cannot readily be achieved with only 3 yr of data set as used in this study. With the increasing number of seismic data from the current seismic networks in Turkey, we can envisage that we will be able to gather much larger numbers of reliable surface wave measurements covering the entire region of Turkey in the following years to come, which will eventually allows us to construct higher resolution images of the crust and upper mantle including anisotropic properties beneath this tectonically active and complex region.

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