Crustal Velocity and Poisson’s Ratio Structures beneath Northwest Anatolia Imaged by Seismic Tomography

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Abstract
The northwestern part of Anatolia is a part of an orogenic belt within the Mesozoic-Cenozoic tectonic units and is located westward relative to the North Anatolian Fault Zone (NAFZ), which is a major tectonic feature in Anatolia. It is a seismically active zone as a result of the tectonic deformation prevailing in the region which has been accompanied by a high-grade magmatism in the latest Oligocene time. In this study, we determine 3-D velocity and Poisson’s ratio images of the crust beneath northwestern Anatolia by inverting a large number of arrival time data of P- and S-waves. Our tomographic images elucidate the major tectonic features of the area and are consistent with the previous geological and geophysical observations in the region. At shallow depths, high P- and S-wave velocity anomalies are revealed, which disappear progressively downward and are replaced by low-velocity zones at depths of 25 and 40 km. The elastic parameter Poisson’s ratio exhibits significant structural heterogeneities compared to the imaged velocity structure with generally high values at most crustal layers. The seismic activity is intense along the highly heterogeneous zones above a depth of 10 km and is closely associated with the westward extension of the NAFZ as well as other local faults in the study region. Results of the checkerboard resolution test, hit count maps and the ray-path coverage, indicate that the imaged velocity and Poisson’s ratio anomalies are reliable features down to a depth of 40 km. The wide distribution of the low-velocity/high Poisson’s ratio zones in the middle and lower crust are consistent with many geophysical observations such as strong Lg attenuation, high heat flow, and the absence of mantle lid, indicating the presence of partial melt in the uppermost mantle.

Keywords: Northwestern Anatolia, Marmara Sea, Seismic Tomography, Seismic Wave Velocity, Poisson’s Ratio, Crustal Structure.

INTRODUCTION
The convergent plate boundary between the African/Arabian plates to the south and the Anatolian Block to the north (Fig. 1) is the main tectonic element in the eastern Mediterranean region. Recent models of plate tectonics (e.g., DeMets et al., 1990; Jestin et al., 1994; McClusky et al., 2000) based on analysis of global seafloor spreading, fault systems, and earthquake slip vectors in the eastern Mediterranean indicate that both
the Arabian and African plates are moving in a north-northeast direction relative to Eurasia at rates of 18-25 and 10 mm/yr, averaged over about 3 Myr, respectively. This differential motion between Africa and Arabia (~ 10 – 15 mm/yr) is accommodated predominantly by left-lateral motion along the Dead Sea Fault Zone (DFZ). The northward motion of Africa and Arabia results in continental collision along the Bitlis-Zagros fold and thrust belt, intense earthquake activity (Fig. 2), high topography in eastern Turkey and the Caucasus Mountains, and the westward extrusion of the Anatolian plate. GPS studies have shown that west Anatolia is extending southwestward at a rate of 30–40 mm/year (Straub and Kahle, 1994, 1995; Oral et al., 1995; Reilinger et al., 1997). In contrast, central Anatolia is moving to the west more slowly, at ~15–20 mm/year (Barka and Reilinger, 1997). The Fethiye Burdur Fault Zone (FBFZ), is thought to be the border between the extending part of western Anatolia and stable central Anatolia (e.g., Barka and Reilinger, 1997). The FBFZ has itself been interpreted as a wide left-lateral zone, with a large component of extension and well-defined seismicity (Taymaz et al., 1991; Taymaz and Price, 1992; Temiz et al., 2001).

The Anatolian block consists of major paleotectonic belts separated by sutures. The tectonic belts were formed by the closure of the Tethyan Ocean and basins connected with it (Elmas and Gürer, 2004). The Tethyan evolution of Anatolia can be divided into two main stages: the Paleotethyan and Neotethyan. The tectonic setting of the Marmara Sea region in northwest Anatolia is controlled mainly by the NAFZ, which is one of the best-known strike-slip faults in the world. The NAFZ is an approximately 1500 km-long, broad arc-shaped, dextral strike-slip fault system that extends from eastern Turkey in the east to Greece in the west (Fig. 1). The age and cause of dextral motion along NAFZ is controversial and there are basically: 1) the right-lateral motion commenced by the Middle Miocene and it is the result of the westward motion of Anatolia away from the collision zone in eastern Turkey when the Arabian and Eurasian plates collided (McKenzie, 1970; Şengör, 1979). 2) others claimed that the NAFZ did not initiate until the latest Miocene or Early Pliocene (Barka and Gülen, 1989), and 3) there are also claims that the NAFZ was initiated in eastern
Anatolia during the Late Miocene and propagated westwards reaching the Sea of Marmara region during the Pliocene (Barka, 1992; Şengör, 1979). The analysis of geological data suggests that the rate of motion on the NAFZ is about 5-10 mm/yr (Barka, 1992). On the other hand, recent GPS data indicate present-day higher rates of about 15-25 mm/yr (Reilinger et al., 1997; McClusky et al., 2000). Just to the east of the Sea of Marmara, the NAFZ splays into two major strands, which are the northern strand that traverses in part the Sea of Marmara and constitutes the most active section of the NAFZ including the segment that slipped in the 17 August, 1999 earthquake, and the southern strand, which bounds the southern margin of the Sea of Marmara to the east; then bends southward and runs in a SW direction into the Aegean Sea (e.g., Bozkurt, 2001; Tur, 2007; Fig. 1). Several depressions are aligned along the NAFZ and its major splays (Neugebauer, 1995; İmren et al., 2001). The NAFZ in the Sea of Marmara (Marmara Fault) is a relatively young active strike-slip fault zone in the floor of a sea of moderate depth. Its tectonics is poorly known in spite of its high potential for producing large earthquakes.

The analysis and distribution of historical earthquakes reveal that among the two westernmost branches of the NAFZ in the Marmara region (Fig. 1), it is the northern strand that is the most active one and has accommodated more large earthquakes. Recent seismological observations indicate that the Anatolian plate and the surrounding regions are seismically active because of intense tectonic deformation with most earthquakes occurring mainly in the upper crust (Fig. 2). To the west, the stress was possibly accumulated on the branches of the NAFZ in the Marmara region, and there are claims that these faults are close to failure (Gürbüz et al., 2000; Parson et al., 2000). Historical records, on the other hand, indicate the occurrence of major destructive earthquakes in the Sea of Marmara, affecting Turkey's most populous and richest region with 80% of her industry and her largest city (population > 10,000,000) Istanbul (İmren et al., 2001; Le Pichon et al., 1999, 2001). Apart from the stress transformed during the Kocaeli earthquake, the Sea of Marmara to the southwest of Istanbul is a region where the strain has been aseismically accumulating since the last earthquake in 1719 (Ambraseys and Finkel, 1987). The close proximity of these faults to Istanbul; a major city with huge number of inhabitants, makes it even more important.

Topographically, the Sea of Marmara comprises several deep marine asymmetric strike-slip basins (e.g., the Çınarcık, Central Marmara, and Tekirdag basins), separated by NE-trending submarine ridges that rise several hundreds of meters above the seafloor. The basins consist of Plio-Quaternary sediments reaching over 3 km in depth and constitute part of the Sea of Marmara. There are two distinct, steep, continuous bathymetric features (submarine escarpments) that bound the Sea of Marmara, both in the north and in the south, and mark the location of major active faults (Okay et al., 2000). These structures are named as the North and South Boundary Faults (Wong et al., 1995; Okay et al., 2000).

The region around the Sea of Marmara has been the locus of many geological and geophysical observations. However, the area under study lacks a detailed model crustal and upper mantle velocity structure. Bariş et al. (2005) determined three-dimensional structures of \( V_p \), \( V_s \) and \( V_p/V_s \) in the upper crust of Marmara region using about 4,000
Figure 2. Epicentral distribution of NEIC (US Geological Survey) seismicity from 1973 to 2011 around the Aegean Sea and western Anatolia. Circles vary in their color according to the depth of the hypocenter. The black rectangle shows the present study area.

events, which generate around 59,000 and 33,000 P- and S-wave arrival times, respectively. Thus, the obtained tomographic images, especially those of S-wave, suffer from poor resolution along most parts of the study area. In this study, we invert a larger data set of arrival times from local earthquakes in northwest Anatolia using a seismic tomography method. The implications of the obtained velocity and Poisson’s ratio structures and their relation to the distribution of other geophysical parameters, the current seismic activity and the present-day tectonics are then discussed for a better understanding of the seismotectonics of northwest Anatolia and its surrounding regions.

DATA AND METHODS
In the present work, we used an initial number of 10,640 local crustal events that occurred between latitudes 39.0 and 41.3° N and longitudes 26.0 and 31.2° E during the period from 2007 up to 2010. Few events are recorded at few stations and have been excluded before inversion, giving a final number of 8,590 events (Fig. 3a). The selected events are thus, recorded by at least 8 stations. These events are recorded by the Turkish National seismological observation network, which is operated by the Earthquake Department of the Turkish Republic Disaster and Emergency Management Presidency. A total of 41 seismic stations have recorded the selected events in the present study area (Fig. 3a). These stations comprise 33 broadband (BB) and 8 short-period seismic stations. Only two seismic stations have analog recording, while the remaining
record the data in a digital format. The sampling frequency of the stations is 50 or 100 Hz. The dynamic range is 140 and 164 - 184 dB for the broadband and the short-period seismic stations, respectively. The crustal model of Herrin (1968) and the HYPO71 source code (Lee and Lahr, 1972) are used for the routine determination of the hypocentral parameters. The estimated errors in the hypocentral locations do not exceed 2.0 km for the majority of the events. With the exception of the eastern and the northern parts of the study area; all other parts have a dense and uniform distribution of seismicity. The seismic stations, on the other hand, are mainly concentrated in the central parts especially along the western segments of the NAFZ. This particular geometry of the recording stations and events controls the spatial resolution and reliability of the obtained velocity and Poisson’s ratio anomalies.

The travel time picking is done automatically first; then checked manually. Seismograms with low signal/noise ratio are routinely excluded from analysis. The selected 8,590 events give rise to 76,130 and 71,960 P- and S-wave arrivals, respectively.

In order to study the relation between the nucleation zones of moderately large earthquakes ($M_b$ or $M_w \geq 5.0$) and the obtained seismic velocity and Poisson’s ratio anomalies, we collected 32 events that occurred in the study area since 1975 up to 2011 from the earthquake catalogs of the National Earthquake Information Center (NEIC), U. S. Geological Survey (Fig. 3b). A close inspection of the distribution of the epicenters of these large events shows that they are distributed mainly along the northern strand of the NAFZ.

We examined the travel time residuals of both P- and S-wave data with respect to the assumed initial velocity model to remove the outliers in the data set. Hence, the numbers of P- and S-wave arrivals are reduced to 69,880 and 64,050, respectively (Fig. 4). It is clear that S-wave residuals show larger spread compared with P-wave residuals. The accuracy of arrival times is estimated to be lesser than 0.15 s for P-wave data and slightly larger (< 0.25 s) for the S-wave data. More than 95% of the travel time residuals used in the tomography inversion have values between ±5.0 s (Fig. 4). This approximately equal number of P- and S-wave arrivals and the good ray criss-crossing of both data sets in most parts of the study area ensure the reliability of the obtained velocity and Poisson’s ratio structures (e.g., Widiyantoro et al., 1999; Gorbakov and Kennett, 2003). Moreover, the large number of the events used in the tomographic inversion along with the relatively uniform distribution of the 41 seismic stations in northwest Anatolia implies good ray path coverage for both data sets (Fig. 5), which is sufficient to accurately locate the anomalous velocity and Poisson’s ratio zones in the study area.

In order to invert the arrival time data around the Marmara Sea region in northwest Anatolia for a 3-D velocity structure, we used the tomographic method of Zhao et al. (1992, 1994) to conduct two separate inversions: one for P-wave and the other for S-wave velocity structures. This method has been adopted for many parts around the globe.
Figure 3. (a) Epicentral distribution of the 10590 earthquakes used in this study shown as circles, which vary in color according to the focal depth (scale at the bottom). The black triangles show the 41 seismic stations around the Marmara Sea region. Black lines denote active faults (Şaroğlu et al., 1992). (b) Distribution of 32 moderately large crustal earthquakes (stars) that occurred in northwestern Anatolia ($7.6 \geq M \geq 5.0$) since 1975 (see text for details). Stars vary in color according to the focal depth. Thick black lines denote active faults in the study region; whereas the thin gray lines denote political borders.
Figure 4. Input travel time residuals relative to the assumed initial velocity model vs. epicentral distance in kilometers for P-wave (a), and S-wave (b), data sets. Note the larger spread in S-wave residuals.

having different tectonic settings (e.g., Zhao and Kanamori, 1995; Zhao et al., 1996, 1997, 2001; Serrano et al., 1998, 2002a, b; Kayal et al., 2002; Salah and Zhao, 2003; Salah et al., 2007; 2011). It can be applied to regions characterized by a general velocity structure, which includes several complex-shaped velocity discontinuities and allows for 3-D velocity variations everywhere in the model. The seismic discontinuities represent known geological boundaries, like the Conrad and Moho boundaries. A 3-D grid net is set up in the model to express the 3-D velocity structure in the modeling space. Velocity perturbations at the grid nodes are taken as unknown parameters. The velocity perturbation at any point in the model is calculated by linearly interpolating the velocity perturbations at the eight grid nodes surrounding that point. An efficient 3-D ray-tracing scheme (Zhao et al., 1992); that iteratively uses the pseudo-bending technique (Um and
Thurber, 1987) and Snell’s law is employed to calculate travel times and ray paths accurately and rapidly. Station elevations and the consequent corrections are taken into account in the calculation of travel times and the ray tracing scheme. The LSQR algorithm (Paige and Saunders, 1982) with a damping regularization is used to solve the large and sparse system of observation equations, allowing a great number of data to be used to solve a large tomographic problem. The nonlinear tomographic problem is solved by iteratively conducting linear inversions. In each iteration, perturbations to hypocentral parameters and velocity structure are determined simultaneously. A detailed description of the method is given by Zhao et al. (1992, 1994) and Zhao (2001). A grid spacing of 0.33° in horizontal directions is adopted for the present study. Vertically, five layers of grid nodes are set up at 4, 12, 25, 40, and 55 km depths (Fig. 6). We first tried slightly different grid spacing and finally adopted the previous grid spacing to get an optimum resolution with respect to the present data set.

Figure 5. Horizontal ray path coverage of P-wave (a) and S-wave (b) data sets. Every path between an event and a recording station is drawn as one straight line. Small white circles and triangles denote events and recording stations, respectively.
After we determined the $P$- and $S$-wave velocity models as described before, we used the equation: 

$$\left(\frac{V_p}{V_s}\right)^2 = 2(1-\sigma)/(1-2\sigma),$$

to determine the spatial distribution of the elastic parameter Poisson’s ratio ($\sigma$) (see Utsu, 1984). By definition, Poisson’s ratio is the ratio of radial contraction to axial elongation, and is a key parameter in studying the physical properties of crustal rocks and can provide more constraints on the crustal composition and fluid distribution than either $P$- or $S$-wave velocity alone (Zhao et al., 2004; Salah et al., 2011). Its value in common crustal rocks ranges from 0.20 to 0.35 (Christensen, 1996). Poisson’s ratio has proved to be very effective for the clarification of the seismogenic behavior of the crust, especially the role of crustal fluids in the nucleation and growth of earthquake rupture (e.g., Zhao and Negishi, 1998; Kayal et al., 2002; Zhao et al., 2002; Salah et al., 2007).

Seismic tomography applied to the solid Earth is a nonlinear process (Pavlis and Booker, 1983). In general, solutions are obtained by linearization with respect to a reference Earth model (e.g., Aki and Lee, 1976; Nolet, 1978). The tomographic images resulting from such linearized inversion are dependent on the initial reference models and hypocentral locations (Michael, 1988; VanderHilst and Spakman, 1989; VanderHilst et al., 1991; Kissling et al., 1994). Thus, selecting the initial velocity model is an important step in any tomographic inversion since it affects the amplitude and distribution of the obtained velocity anomalies. Moreover, since the iterative inversion scheme is linear, the only carefully-chosen model assures to find the globally best-fitting model. After trying many models, which have been obtained for northwest Anatolia, we used the model of Özalaybey et al. (2002) since it gives the minimum travel time residuals for both data sets (Table 1, Fig. 7). In this model, the transition between the upper crust and lower crust occurs at a depth of 20 km; while the Moho is encountered at a depth of 33 km, being close to the global average (Bilim, 2011). The $V_p/V_s$ ratio is generally around 1.73. As stated before, we first checked a number of slightly different initial $P$-wave velocity models with different $V_p/V_s$ ratios (varying gradually from 1.65 to 1.80) and applied it to different sub-data sets and found that the overall seismic structure has no substantial variations with only slight changes in some portions. Finally, the model shown in Table 1, with a $V_p/V_s$ ratio of 1.73 are selected as they give the minimum RMS travel time residuals.

**RESOLUTION AND RESULTS**

In any tomographic inversion, it is important to check the reliability and resolution scale of the obtained velocity anomalies. The checkerboard resolution test (e.g., Inoue et al., 1990; Zhao et al., 1992) is useful in this case as it gives a straightforward way to discriminate areas of good ray criss-crossing from those of poor ray coverage and consequently poor resolution. To make a checkerboard, positive and negative velocity anomalies of 3% are assigned to the 3-D grid nodes as in Fig. 8. Synthetic arrival times are calculated for this input checkerboard model. Numbers of stations, events and ray paths in the synthetic data are the same as those in the real data set. Random errors of Table 1. Initial $P$- and $S$-wave velocity, and $V_p/V_s$ ratio used in the tomographic inversion (Özalaybey et al., 2002).

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>$V_p$ (km/s)</th>
<th>$V_s$ (km/s)</th>
<th>$V_p/V_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>2.9</td>
<td>1.68</td>
<td>1.726</td>
</tr>
<tr>
<td>1</td>
<td>5.7</td>
<td>3.29</td>
<td>1.733</td>
</tr>
<tr>
<td>6</td>
<td>6.1</td>
<td>3.53</td>
<td>1.728</td>
</tr>
<tr>
<td>20</td>
<td>6.8</td>
<td>3.93</td>
<td>1.730</td>
</tr>
<tr>
<td>33</td>
<td>8.05</td>
<td>4.65</td>
<td>1.731</td>
</tr>
</tbody>
</table>
Figure 6. Configuration of the grid net adopted for the present study in horizontal (a) and depth (b), directions. Grid spacing is 0.33° and 8 - 15 km in horizontal and depth directions, respectively. Straight lines in (a) running E-W and N-S denote the locations of two vertical cross sections shown in Figs. 16 and 17 (see later sections).
Figure 7. The initial $P$- (solid line) and $S$-wave (dashed line) velocity models adopted for the present tomographic inversion. The Moho discontinuity is set at a depth of 33 km.

Figure 8. An example of the input checkerboard synthetic model for both $P$- and $S$-wave data (see text for details). Black and white symbols denote positive and negative synthetic velocity anomalies ($\pm 3\%$), respectively, which are assigned to the grid nodes.

0.1–0.2 s similar in magnitude to those of the real data are added to the synthetic data, which are then inverted with the same algorithm used for the real data. The inverted image of the checkerboard shows areas of good and poor resolution. Figs. 9 and 10 show the resolution of both $V_p$ and $V_s$ structures, respectively. The checkerboard resolution test indicates a good and uniform resolution of about 33 km horizontally for both $V_p$ and $V_s$ structures beneath northwest Anatolia especially at 12 and 25 km depths. This is because of the more uniform distribution of many horizontal and vertical ray paths passing at these depths. However,
the edge portions at 4 and 40 km depths have a relatively poor resolution because of insufficient ray paths criss-crossing at these regions.

We applied the tomographic method described in the previous section to the northwestern Anatolia data set (Fig. 3a). We found that the sum of squared travel-time residuals has reduced by more than 45% of its initial value after the inversion. The final root-mean-square travel time residuals are 0.287 s for P-wave and 0.423 s for S-wave data. The study area has enough ray coverage at four depth layers (4, 12, 25, and 40 km) in which the number of P and S rays passing through each grid node (hit count) is sufficient to retrieve the velocity anomalies at such depths (Figs. 11 and 12). Large areas in the study region have good hit counts and many nodes are hit by more than 5000 rays at the first three layers. Grid nodes with hit counts <8 are not included in the inversion.

Inversion results for $V_p$, $V_s$, and $\sigma$ distributions at four depth layers are shown, respectively, in Figs. 13, 14, and 15. Figs. 16 and 17 show the $V_p$, $V_s$, and $\sigma$ images along two vertical cross-sections in northwest Anatolia (see Fig. 6 for the location of cross-sections). These images show the velocity and Poisson’s ratio perturbations in percentage from the initial velocity model at each depth. We first tried a number of inversions by adopting slightly different initial models and using different sub-data sets. It was found that the overall pattern of the velocity and $\sigma$ structures as shown in Figs. 13–15, and Figs. 16–17 is stable and the change in the amplitude of the velocity anomalies is generally less than 1%.

Significant lateral and vertical velocity variations amounting to ±8% and up to ±10% for Poisson’s ratio are clearly revealed beneath northwest Anatolia. Higher-than average velocity anomalies are seen at a depth of 4 km, which change gradually downward to average velocity at 12 km depth and low-velocity at 25 km depth (Figs. 13 and 14). The background seismicity (usually with depths < 8 km) and the moderately large crustal earthquakes occur throughout the study region with some concentrations near fault zones. Poisson’s ratio ($\sigma$) shows higher structural heterogeneity at all crustal layers (Fig. 15), and is generally higher than the average at both shallow and deeper layers (4, 12, and 40 km depths) indicating a general low S-wave velocity compared to the P-wave velocity. The $\sigma$ image at a depth of 25 km is clearly lower than the average; although small portions of high $\sigma$ zones are also detected (Fig. 15-c). The majority of the moderate and large earthquakes are closely related to the active fault zones which are characterized generally by low-velocity/high Poisson’s ratio (Figs. 13-15). Along the E-W and N-S cross sections (Figs. 16 and 17); shallow high-velocity/high Poisson’s ratio zones are clearly visible, which change to low-velocity/low Poisson’s ratio at a depth of 25 km. We can notice that the crustal earthquakes are concentrated at 7-8 km depths, which may partly be related to the routine location of earthquake hypocenters. It is clear that the moderately large crustal earthquakes (shown as big white circles) occur in both average velocity/average to high Poisson’s ratio zones (Figs. 16 and 17). The implications of these velocity and $\sigma$ anomalies and their relation to other geophysical studies conducted in northwestern Anatolia are discussed in the following paragraphs.

DISCUSSION

4.1. Previous Geophysical Observations in Northwest Anatolia

We estimated 3-D seismic velocity and Poisson’s ratio structures for the crust beneath northwest Anatolia. Shallow high-velocity zones with intense seismic activity are detected in this study (Figs. 13, 14, 16, and 17), which are consistent with shallow resistive structures obtained by Tank et al. (2003) through the two-dimensional inversion of magnetotelluric data in the present study area. The aftershocks of the 17 August 1999 İzmit earthquake occur mostly in this resistive zone which is underlain by a moderately conductive zone. On the other hand, swarm activities tend to occur in the conductive zone below the highly resistive zone (Tank et al., 2003). Çağlar and İseven (2004) mapped three electrically-conductive zones (~4-6 Ω m) beneath northwest Anatolia that extend in depth from 5 to 15 km below the surface. These zones
are significant for the geothermal system and the ascending of hydrothermal fluids containing altered minerals. Elmas and Gürer (2004) detected low resistivity values (10 $\Omega$ m) beneath the northern Sakarya Low and the Sakarya Zone (located at the northeastern part of the present study region) at a depth of 26 km. This low resistivity zone may provide an indication of thinning in the total crustal thickness and the presence of a visco-elastic crust with partial melting. They also detected a five-layer conductive sequence through magnetotelluric profiling along the northern part of their profile corresponding to fragmentation of the İstanbul Zone. Although no surface geologic observations indicate the presence of underlying magma except some geothermal activities; the stratigraphic, sedimentologic and structural features of the northern Sakarya Low indicate that the region has been affected by an extensional regime. Inversion of magnetotelluric data by Gürer et al. (2004) in southwest Anatolia reveals two subzones of crust with varying thicknesses: conductive lower crust (< 75 $\Omega$ m), overlain by resistive (> 350 $\Omega$ m) upper crust, with four resistive cores (> 2000 $\Omega$ m) separated by three relatively conductive vertical zones. The moderate- to high-velocity zones in the upper crust at depths of 4 and 12 km from the present results are in good agreement with the resistive upper crust detected by Gürer et al. (2004). Although the area covered by Gürer et al. (2004) is located slightly to the south of our study area, we think that similar conditions prevail throughout western Anatolia. Results of 2-D resistivity-depth cross sections obtained by Özürlan et al. (2006) for western Anatolia show very low resistivity values near hot springs. The 2-D inversion results also indicate the presence of fault zones striking nearly N-S and E-W, and fault-bounded graben-horst structures that show promising potential geothermal field resources; which are consistent with low-velocity zones in middle/lower crust.

A mean value of heat flow of $107 \pm 45$ mW/m$^2$ has been obtained for the western part of Anatolia, which is about 60% above the world average (Ilkışık, 1995). Moreover, a close association exists between areas of high heat-flow values (above 100 mW/m$^2$), and areas of Tertiary and younger volcanism. On the other hand, estimations of terrestrial heat flow density by Pfister et al. (1998) for the Marmara Sea region varies regionally from 35 to 115 mW/m$^2$, with a mean value of 60 mW/m$^2$. The heat flow density map shows that the extensional tectonics of the region south of the Marmara Sea is characterized by higher heat flow density values. The abundance and distribution of hot springs and geothermal fields in the region can be explained by local zones of strongly elevated vertical hydraulic permeability due to active transtensional faulting of the crust. These observations are consistent with the low-velocity/high Poisson’s ratio detected in the present study. Zor et al. (2006) estimated an average crustal thickness of 33 km and high heat flow value (101±11 mW/m$^2$) for the eastern Marmara Sea region, with remarkable extensional features, implying that the region has a Basin and Range-type characteristics. Results of Zor et al. (2006) also indicate that the eastern Marmara region seems to be a transition zone between the Marmara Sea extensional domain and the continental Anatolian inland region. In addition, the Galatia volcanic complex, which is located in the eastern part of the present study area, exhibits low Curie-point-depth and high heat flow values (Bilim, 2011).

4.2. Seismological Studies beneath Northwest Anatolia

According to Nakamura et al. (2002), the $P$-wave velocity structure west of the mainshock hypocenter of the 1999 İzmit earthquake ($M_w=7.4$) has a distinct low-velocity area. Al-Lazki et al. (2004) mapped broad-scale (~500 km) zones of low (< 8 km/s) $Pn$ velocity anomalies beneath the Anatolian plate and the Anatolian plateau and smaller scale (~200 km) very low (< 7.8 km/s) $Pn$ velocity zones beneath many parts in the eastern Mediterranean including central Turkey and the northern Aegean Sea. Most parts of the present study area have low $Pn$ velocity according to the results of Al-Lazki et al. (2004). The broad-scale low-velocity regions are interpreted to be hot and unstable mantle lid zones, whereas very low $Pn$ velocity zones are interpreted to be regions of no mantle lid. The low $Pn$ velocity zones beneath the Anatolian plate, eastern Turkey and northwestern Iran may in part be a result of the subducted Tethyan oceanic lithosphere.
beneath Eurasia. These results are in agreement with the low-velocity zones in the middle/lower crust detected in this study. The heterogeneous structures that are clear in Figs. 13-17, are consistent with the strong lateral heterogeneity along the western part of the NAFZ as detected by Bariş et al. (2005). High $V_p/V_s$ ratios and low-velocity zones correspond to elevated fluid content, which may be enhanced by postseismic fluid migration processes (Husen et al., 2000). Beneath oceans, they indicate either hydrated oceanic crust with fluid-saturated sediments or serpentinitized uppermost oceanic mantle (Haberland et al., 2009). According to Innocenti et al. (2005), western Anatolia is a region where Miocene-to-present day magmatism is well established, which may interpret the low-velocity and high Poisson’s ratio zones in the intermediate to lower crust. Akyol et al. (2006) obtained a 1-D crustal $P$-wave velocity model for western Anatolia including a significant part from the present study area. Their velocity model is characterized by crustal velocities that are significantly lower than average continental values. The low velocities may be associated with high crustal temperatures, a high degree of fracture, or the presence of fluids at high pore pressure in the crust. Moreover, they found large standard deviation values for the 1-D velocity model indicating significant lateral variations in velocity structure (Akyol et al., 2006). Seismic imaging of the Çınarık Basin located along the northern branch of the NAFZ in the Sea of Marmara by Carton et al. (2007) shows deep-penetrating faults, hence long-lived features, which have accommodated a large amount of extension. These faults serve as conduits for the ascending of the hot uppermost mantle materials. Erduran et al. (2007) analyzed surface waves from shallow regional earthquakes (many of them occur in the present study area) recorded at the GEOFON Isparta station, which is located slightly to the south of our study region. They estimated shear wave velocities which are considerably lower than indicated by PREM in almost all depth ranges starting from the upper crust downward to the uppermost mantle. Other similar studies (e.g., Meier et al., 2004; Karagianni et al., 2005; Maggi and Priestly, 2005; Pasyanos, 2005) detected also low shear wave velocity at the same depth ranges. Earlier studies of Sandvol et al. (2001); and Al-Damegh et al. (2004) have shown that $L_g$ propagation in the Turkish-Iranian plateau is usually blocked or highly attenuated. Similarly, recent observations of Zor et al. (2007) show moderate to high $L_g$-attenuation values ($Q_0 \sim 100-200$) beneath the Turkish plateau, which probably originates from both scattering and intrinsic attenuation due to the tectonic complexity and the wide-spread young volcanics in the region.
Figure 9. The results of the checkerboard resolution test for $P$-wave velocity at four depths (see text for details). Black and white symbols denote high and low velocities, respectively. The perturbation scale is shown at the bottom. The depth of each layer is shown below the map.

Figure 10. The results of the checkerboard resolution test for $S$-wave velocity at four depths. Other details are similar to those of Fig. 9.
Figure 11. Number of rays passing through each grid node (hitcount) for $P$-wave data at four depth slices. Scale is shown at the bottom. Grid nodes with less than 8 rays passing through are not included in the inversion.

Figure 12. Number of rays passing through each grid node for $S$-wave data at four depth slices. Other details are similar to those of Fig. 11.
Figure 13. $P$-wave velocity structures (in %) at depths of 4 (a), 12 (b), 25 (c), and 40 (d) km beneath northwestern Anatolia. Red and blue colors denote low- and high-velocities, respectively. Numbers between brackets show the depth range of the microseismic activity plotted as crosses. Moderately large earthquakes ($M \geq 5.0$) occurring in the same depth range of the background seismicity are plotted as open circles. Thin solid lines denote active faults in northwestern Anatolia. The perturbation scale (±8%) is shown to the right.

Figure 14. $S$-wave velocity structures at the four depth slices. Other details are similar to those of Fig. 13.

Figure 15. Distribution of Poisson’s ratio ($\sigma$) structures at four depth slices. Red and blue colors denote high and low $\sigma$, respectively. The perturbation scale (±10%) is shown to the right. Other details are similar to those of Fig. 13.
FIG. 16. Vertical cross sections of $V_p$, $V_s$ and $\sigma$ structures along the line running E-W (see Fig. 6 for the locations of the cross sections). The red color denotes the low-velocity and high Poisson’s ratio, whereas high-velocity and low Poisson’s ratio are shown in blue. Large white circles and crosses show, respectively, the location of moderately large earthquakes ($M \geq 5.0$) and the microseismic activity in a 40 km wide-zone around the profile. The perturbation scale ($\pm 8\%$ for velocity and $\pm 10\%$ for Poisson’s ratio) is shown to the right.
CONCLUSIONS

We estimated the 3-D velocity and Poisson’s ratio structures beneath the region surrounding the Marmara Sea in northwest Anatolia by inverting a large number of P- and S-wave arrival times generated from local earthquakes, which are recorded at 41 seismic stations in the study area. Ray path coverage for both P- and S-wave data sets and the results of the checkerboard resolution test as well as hit count maps indicate that the obtained structures are reliable features down to a depth of about 40 km. The following conclusions can be drawn from the obtained results:

1. Shallow moderate to high-velocity zones are clearly revealed beneath most parts of the study area, which change downward to lower velocity at the middle – lower crustal depths.
2. High Poisson’s ratio zones are visible at most crustal layers down to a depth of 40 km, which are consistent with the possibility of the existence of ascending fluids at the base of the upper crust along deep penetrating faults.

3. No clear relationship between the seismic activity (including both large and small events) and the velocity anomalies is established from the present results. This conclusion is consistent with the results of Bariş et al. (2005). On the other hand, most of the seismic events occur near moderate to high Poisson’s ratio zones implying the probable existence of over-pressurized fluids in the source areas of the crustal events.

4. The shallow higher-than-average velocity structures are consistent with the presence of highly resistive zones at shallow depths. On the other hand, the obtained low-velocity/moderate to high Poisson’s ratio zones in the middle – lower crust are consistent with many geophysical evidences such as low resistivity, high heat flow, low $Pn$ velocity and strong $Lg$ attenuation, etc.

More detailed deep seismic imaging of the region combined with geological and other geophysical observations are, however, necessary for a better understanding of the seismotectonic setting of the western segment of the Anatolian plateau including the seismically active North Anatolian Fault Zone.

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