

# **Seismic Tomography of the Arabian-Eurasian Collision Zone and Surrounding Areas**

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<b>14. ABSTRACT</b>  We developed P- and S- wave velocity structures in the crust and upper mantle in the Arabian-Eurasian collision zone and surrounding areas, including Iran, Arabia, Eastern Turkey, and the Caucasus. We first obtain P <sub>n</sub> and S <sub>n</sub> velocities using local and regional arrival time data. Second, we obtain the 3-D crustal P and S velocity models. Third, we extend the model into the upper mantle and transition zone by combining local, regional and teleseismic data and crust model constraints. To extend the model deeper into the mantle through the upper mantle transition zone, ISC/EHB data for P and PP phases are combined with the ABCE data. To counteract the "smearing effect," the crust and upper mantle velocity structure, derived from regional travel-times, is used. An adaptive grid method based on ray density is used in the inversion. A P-wave velocity model extending down to a depth of 2000 km is obtained. regional-teleseismic tomography provides a high-resolution, 3-D P-wave velocity model for the crust, upper mantle, and the transition zone. The crustal models correlate well with geologic and tectonic features. The upper mantle tomograms show the images of the subducted Neotethys slab. The slab geometry is quite complex, reflecting the history of the changes in the plate motions and collision processes. V <sub>p</sub> /V <sub>s</sub> tomography was also determined in the study region, and an attenuation tomography was obtained in Iran.						
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## 1. Abstract

The objectives of this study are to determine P- and S- wave velocity structures in the crust and upper mantle, and to characterize seismic wave propagation in the Arabian-Eurasian collision zone and surrounding areas, including Iran, Arabia, Eastern Turkey, and the Caucasus.

The travel-time tomography is carried out in three steps. First, obtain Pn and Sn velocities using local and regional arrival time data. Second, obtain the 3-D crustal P and S velocity models. Third, extend the model into the upper mantle and transition zone by combining local, regional and teleseismic data and crust model constraints. We obtained the regional Pn velocity tomograms using 160,000 arrival times (including data from Iran) from 850 stations and 18,000 earthquakes. For Sn tomography, 75,000 phase readings were used. The Pn and Sn velocity variations agree quite well with some local differences. Pn velocities are very low under eastern Anatolia, northwest Iran, and the Lesser Caucasus. Localized low velocity anomalies are found under the Iranian plateau. Pn velocities are high under the Arabian Platform, the Gulf, and the Zagros. In the north, high Pn velocities extend from the Black Sea to the Caspian and east to the Kara Kum Basin.

To extend the model deeper into the mantle through the upper mantle transition zone, ISC/EHB data for P and PP phases are combined with the ABCE data. To counteract the “smearing effect,” the crust and upper mantle velocity structure, derived from regional travel-times, is used. An adaptive grid method based on ray density is used in the inversion. A P-wave velocity model extending down to a depth of 2000 km is obtained. The combined regional-teleseismic tomography provides a high-resolution, 3-D P-wave velocity model for the crust, upper mantle, and the transition zone. The crustal models correlate well with geologic and tectonic features. The upper mantle tomograms show the images of the subducted Neotethys slab. The slab geometry is quite complex, reflecting the history of the changes in the plate motions and collision processes.

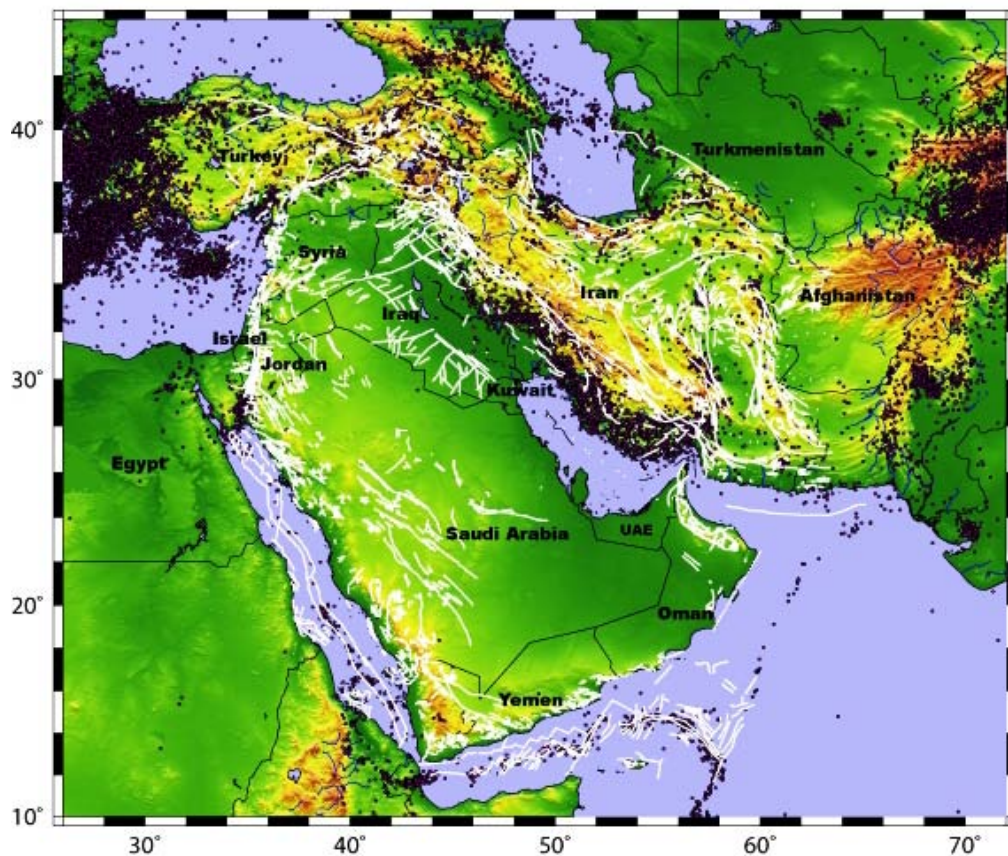
Vp/Vs tomography was also determined in the study region, and an attenuation tomography was obtained in Iran.

## 2. Introduction

The Arabian-Eurasian plate boundary is extremely complex, and it is an ideal region to study a young (geologically) continent-continent collision belt. The area of the study, shown in Figure 1, extends east-west from the Mediterranean to Central Asia, and north-south from the Caspian to the Gulf of Aden and the Arabian Sea. The current tectonics of the region are controlled by the collision and continuing convergence of the Arabian and Eurasian plates. The Arabian and Eurasian plates collided in the early Miocene, after the Neotethys Sea was subducted beneath Eurasia (Bird, 1978; Şengör and Yılmaz, 1981; Jackson and McKenzie, 1984; Dewey et al., 1986). Pre-, syn-, and post-collision tectonics produced very complex structures in the region. Over the last decade, a number of seismic studies have examined the crust and upper mantle structure beneath the Middle East to constrain the nature of the Arabian-Eurasian collision zone. Large-scale surface wave tomography studies have shown variable crustal thickness and upper mantle velocities (Ritzwoller and Levshin, 1998; Pasyanos et al., 2001; Villasenor et al.,

2001; Pasyanos and Walter, 2002; Shapiro and Ritzwoller, 2002; Alinaghi et al., 2007; Reiter and Rodi, 2006). Regional-scale surface wave tomographic studies further highlight the complexity of the collision zone (Mindevalli and Mitchell, 1989; Rodgers et al., 1999; Mokhtar et al., 2001; Maggi and Priestley, 2005) showing a thickened crust under the Caucasus and Zagros, and low shear velocity beneath the Turkish and Iranian plateaus.

Hearn and Ni (1994), Ritzwoller et al. (1998), Al-Lazki et al. (2003; 2004), and Phillips et al. (2007) found slow Pn velocities ( $\leq 8$  km/s) beneath the Anatolian plateau, northwestern Iran, the Greater Caucasus, and southwestern Arabia. The Pn velocities beneath northern Arabia and the Caspian region are faster than average (Al Lazki et al., 2004; Ritzwoller et al., 2002). This high degree of variability suggests that the Earth structure may be extremely complicated in the region.



**Figure 1. Topographic map of the Middle East and surrounding regions. White lines denote the location of known faults, while the black dots represent the epicenter locations of earthquakes in the region.**

Studies of the propagation and attenuation characteristics of regional waves (e.g., Pn, Sn, and Lg) provide additional evidence for strong heterogeneities. Surface wave studies show high shear wave attenuation beneath Iran, Anatolia and the western part of the Arabian plate, and relatively low attenuation in central and eastern Arabia (Seber and Mitchell, 1992; Sandvol et al., 2001; Cong and Mitchell, 1998; Jamberie and Mitchell, 2004; Molinaro et al., 2005;

Bergman et al., 2008; Priestley et al., 2008; Pasyanos et al., 2009). Sn and Lg waves are attenuated through much of the collision zone between the Arabian and Eurasian plates (Kadinsky-Cade et al., 1981; Rodgers et al., 1987; Mitchell et al., 1997; Cong and Mitchell, 1998; Gök et al., 2000; Sandvol et al., 2001; Al-Damegh et al., 2004; Zor et al., 2007). An Lg blockage exists across the Bitlis suture zone and across the Zagros fold and thrust belt. The studies mentioned above present consistent results for the crust and uppermost mantle seismic properties on a regional scale. Significant variation in waveforms, observed particularly in short-period seismograms, over propagation paths that are close to each other, suggests structural variations over short distances at regional boundaries. Delineating these features requires seismic data from dense local and regional seismic networks.

In this study, using both travel-time tomography and attenuation tomography, we produce multi-scale, 3-D P- and S-wave velocity models for the crust, upper mantle, and transition zone. Vp/Vs models and attenuation models in Iran are also provided. In the following sections we describe regional models and models for Iran, respectively.

### **3. Regional models**

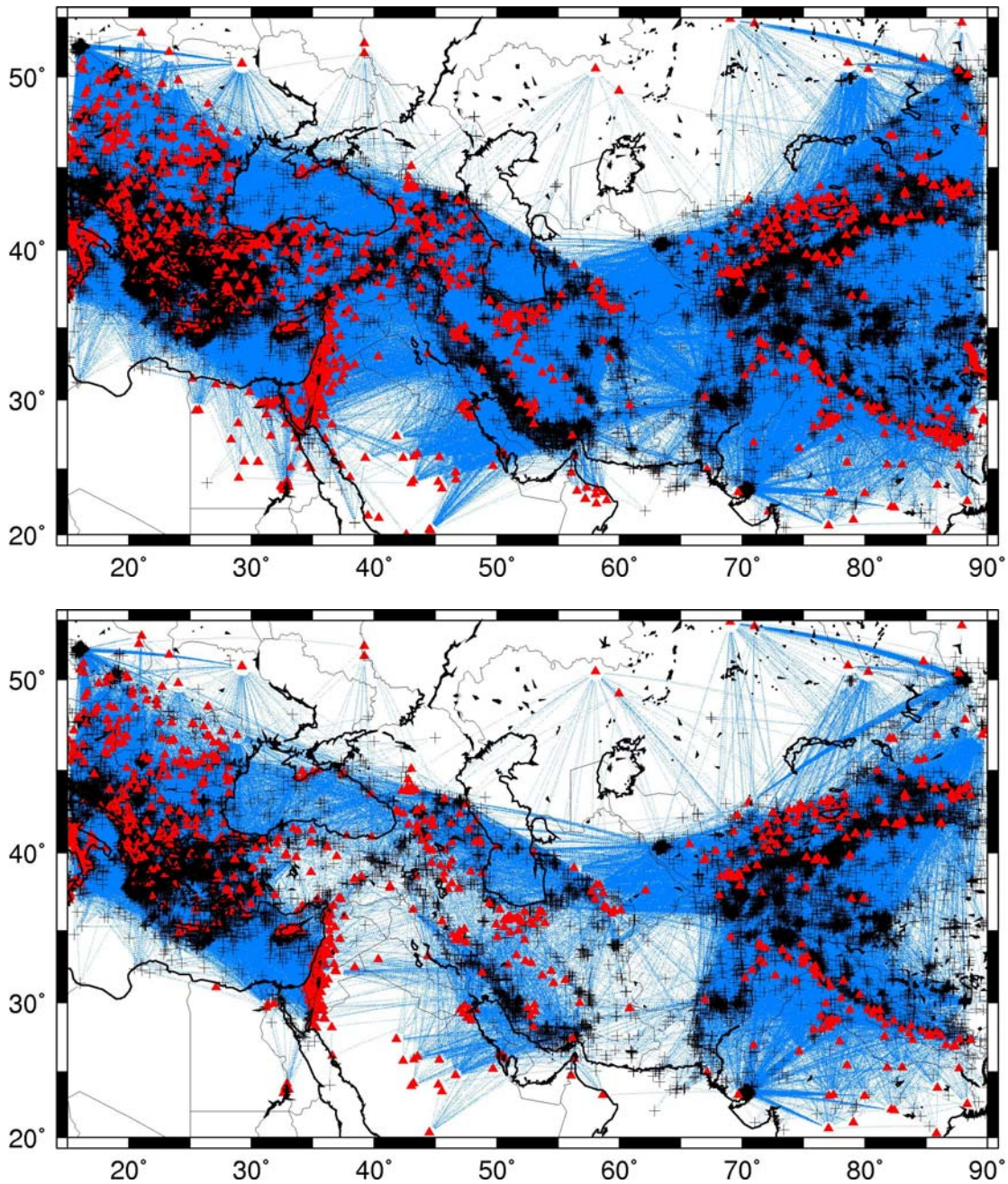
#### ***3.1 Crust and uppermost Mantle Models***

The P- and S-wave arrival times from local, regional and global stations come from four sources: (1) phase readings from many countries in the study region; (2) EHB (Engdahl, van der Hilst, and Buland, 1998) data reprocessed from the International Seismological Centre (ISC) bulletins (1964-2004); (3) PP-P differential times, used to image the upper mantle beneath regions with few earthquakes and stations; and (4) broadband waveform data, used for model validation using synthetic seismograms.

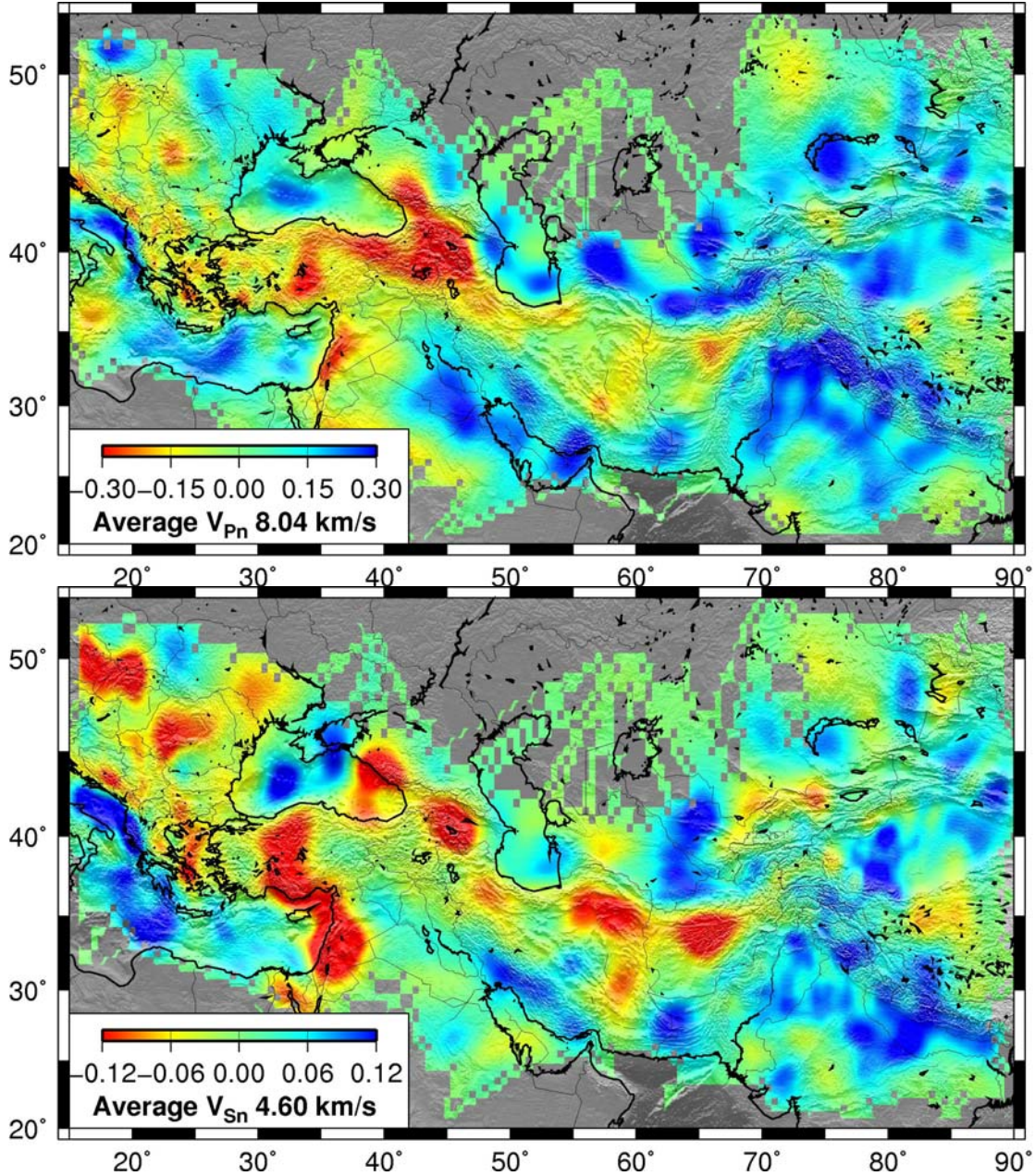
Combining the global, regional and local seismic network data from various sources requires careful processing in order to avoid internal inconsistencies. The travel time picks from the above source 1 have been integrated and processed along with data from source 2 (i.e., the existing EHB data) using a non-linear process of earthquake relocation and phase re-identification (Engdahl et al., 1998). Most baseline inconsistencies can be removed, however, by recalculating all new picks relative to EHB hypocenters using the ak135 travel time.

We have obtained Vp and Vs velocity images of the uppermost mantle beneath the Middle East region by performing tomographic inversion of both Pn and Sn travel times. 654,999 Pn arrivals and 121,838 Sn arrivals were selected from the combined database. Average Pn and Sn velocities are 8.04 km/s and 4.60 km/s, respectively, and maximum velocity perturbations are about 6%. Pn velocity has a close relation with topography. In general, mountains with high elevation show low velocity, while seas, basins and plains with low elevation show high velocity. The large tectonic lines are boundaries between high and low Pn velocity, such as the Hellenic arc, Cyprean arc, Zagros suture, Charman fault, Hari Rod fault. The Sn velocity shows a similar pattern to Pn velocity.

Figures 2 and 3 show the results of Pn and Sn tomography in the study region.



**Figure 2. Ray paths for Pn (upper) and Sn (bottom) travel times. 654,999 Pn rays were obtained from 42,984 events (black crosses) recorded by 1263 stations (red triangles), and 121,838 Sn rays were obtained from 22,283 events recorded by 887 stations. The data were selected for epicentral distances between 2 and 12 degrees with residuals less than 7.0s for Pn, and between 2 and 15 degree with residuals less than 6.0s for Sn. Each event was recorded by at least 3 stations for Pn and 2 stations for Sn.**



**Figure 3. Imaged Pn (upper) and Sn (bottom) velocity lateral variations. Average Pn velocity is 8.04 km/s and average Sn velocity is 4.60 km/s. Red represents slow velocities and blue denotes high velocities.**

### **3.2 *V<sub>p</sub>/V<sub>s</sub> tomography***

Pn and Sn travel-time residuals are inverted for the lateral velocity variation and anisotropy within the mantle lid. Within the epicentral distances considered in this study, the Pn and Sn ray paths can be modeled as refracted rays traveling along the Moho

discontinuity. The variation of seismic velocity within the uppermost mantle is parameterized by subdividing the surface of the uppermost mantle into a 2-D grid of square cells with dimensions of 30 km by 30 km. The Pn and Sn travel-time residuals are described as the sum of three time terms (Pei et al., 2007; Hearn, 1996):

$$t_{pn_{ij}} = a_{pn_i} + b_{pn_j} + \sum d_{ijk} \cdot s_{pn_k} \quad (1)$$

$$t_{sn_{ij}} = a_{sn_i} + b_{sn_j} + \sum d_{ijk} \cdot s_{sn_k} \quad (2)$$

where  $t_{pn_{ij}}$  and  $t_{sn_{ij}}$  are the Pn and Sn travel time residual for the ray from event  $j$  to station  $i$ ;  $a_{pn_i}$  and  $a_{sn_i}$  are the static delays for station  $i$ , depending on the crustal thickness and velocity beneath the station;  $b_{pn_j}$  and  $b_{sn_j}$  are the static delays for event  $j$ , not only depending on the crustal thickness and velocity beneath the event, but also on the event focal depth;  $d_{ijk}$  is the distance traveled by ray  $ij$  in mantle cell  $k$ ;  $s_{pn_k}$  and  $s_{sn_k}$  are the Pn and Sn slowness perturbation for cell  $k$ .

Equation (1) subtracted from equation (2), and using  $t_{ij} = t_{sn_{ij}} - t_{pn_{ij}}$ ,  $a_i = a_{sn_i} - a_{pn_i}$ ,  $b_j = b_{sn_j} - b_{pn_j}$ ,  $s_k = s_{sn_k} - s_{pn_k}$ , then gives:

$$t_{ij} = a_i + b_j + \sum d_{ijk} \cdot s_k \quad (3)$$

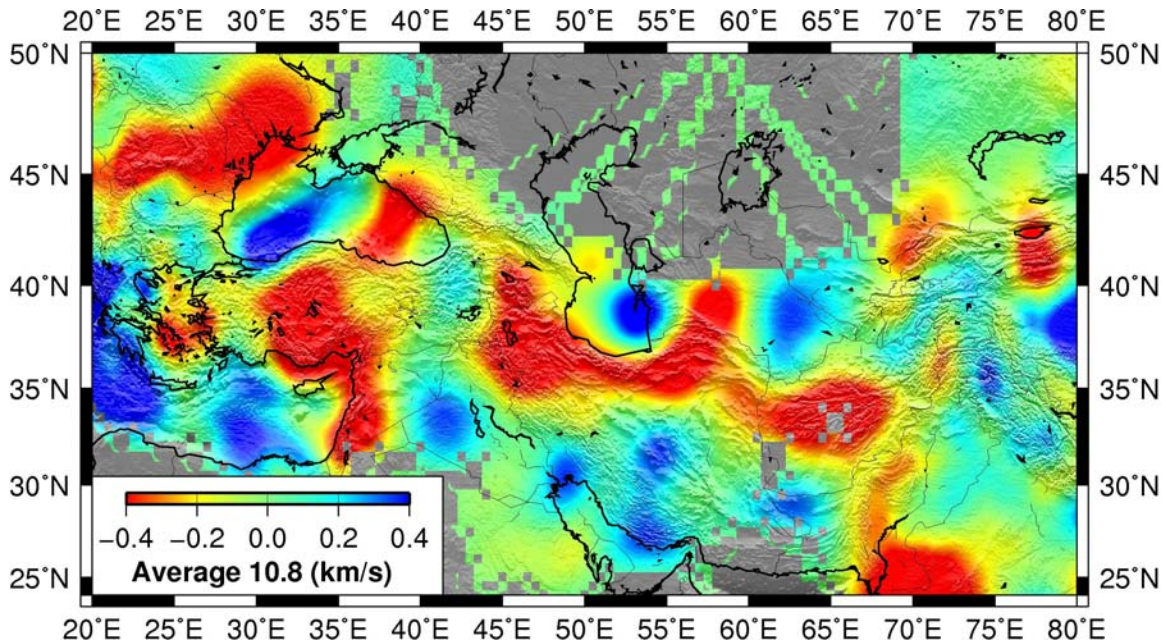
where  $t_{ij}$  is the difference travel time residual;  $a_i$ ,  $b_j$  and  $s_k$  are the static delays for station  $i$ , static delays for event  $j$ , and slowness for pseudowave. Following Hearn's approach and computation method (Hearn, 1996), we invert difference travel time residuals between Pn and Sn for lateral velocity variation of pseudowave within the mantle lid. Equation (3) can be recast as  $\mathbf{A} \mathbf{x} = \mathbf{d}$ , where  $\mathbf{d}$  is the data vector,  $\mathbf{x}$  is the vector of unknowns and  $\mathbf{A}$  represents the model parameters. We apply damped LSQR (Paige and Saunders, 1982) with preconditioning to find the least-squares solution.

The velocity ratio between Pn and Sn can be obtained by multiplying Pn velocity in the two sides of the equation  $s_k = s_{sn_k} - s_{pn_k}$ , namely  $R = 1 + s_k \cdot V_{pn_k}$ , where  $V_{pn_k}$  is the Pn velocity in cell  $k$ . The Poisson's ratio can be obtained by  $\rho = (R^2 - 2) / (R^2 - 1) / 2$ .

In the time period between January 2006 and March 2008, our database for the Pn tomography in the whole study region includes 160,000 arrival times from 850 stations and 18,000 earthquakes. The source-station ray paths are shown in Figure 2. For Sn the data are fewer, with 75,000 total phase readings. The ray paths are also shown in Figure 2. Pseudowave lateral velocity variations (Figure 4) were obtained by tomographically inverting the difference travel time residuals. Relative to the mean value of 10.8 km/s, the pseudowave maximum velocity perturbation is about 0.4 km/s. The average pseudowave velocity is the same as that found from the average Pn and Sn velocity reported last quarter.

The general features of pseudowave velocity patterns are similar to Pn and Sn velocities shown in Figure 3A and 3B, respectively, and similar to the Pn velocities obtained by

Philips et al. (2007), Al-Damegh et al. (2004), and Al-Lazki et al. (2004). Low velocities are under eastern Anatolia, NW Iran, the Southern Caucasus and along the Levant Fracture (Dead Sea Fault) Zone. A prominent low velocity feature is observed just to the south of the Caspian Sea. The  $V_p/V_s$  ratios are shown in Figure 4. Low  $V_p/V_s$  ratios correlate well with high velocity zones and high  $V_p/V_s$  ratios correlate well with low velocity zones.

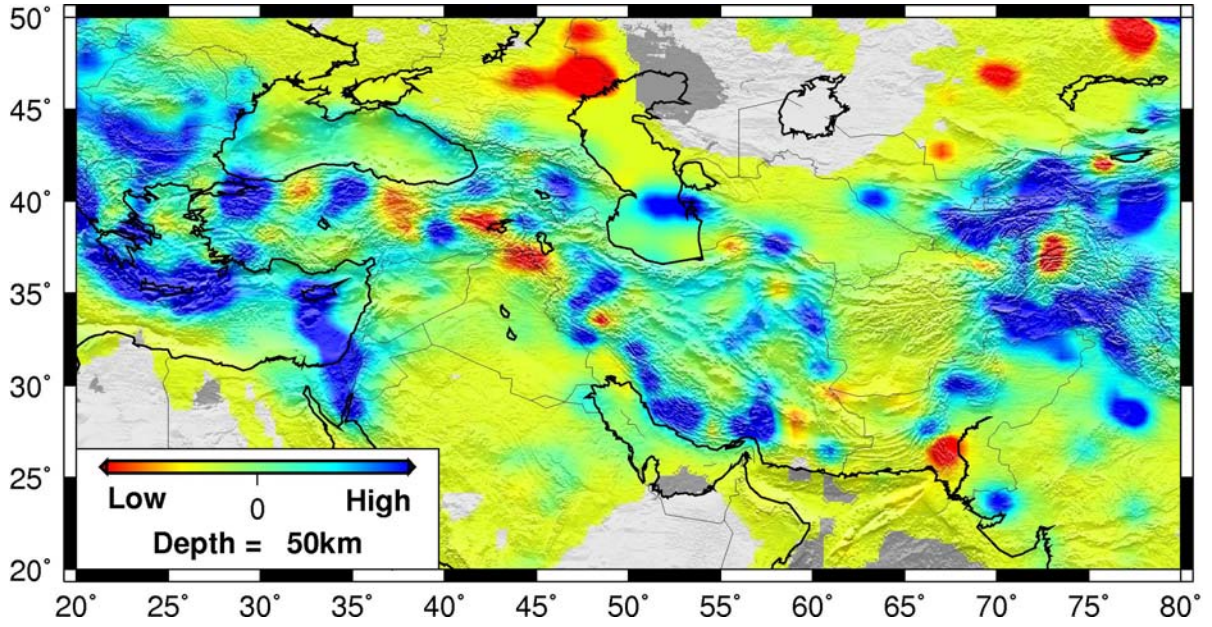


**Figure 4. Pseudowave lateral velocity variations obtained by tomographically inverting the difference travel-time residuals. Relative to the mean value of 10.8 km/s, the pseudowave maximum velocity perturbations is about 0.4 km/s.**

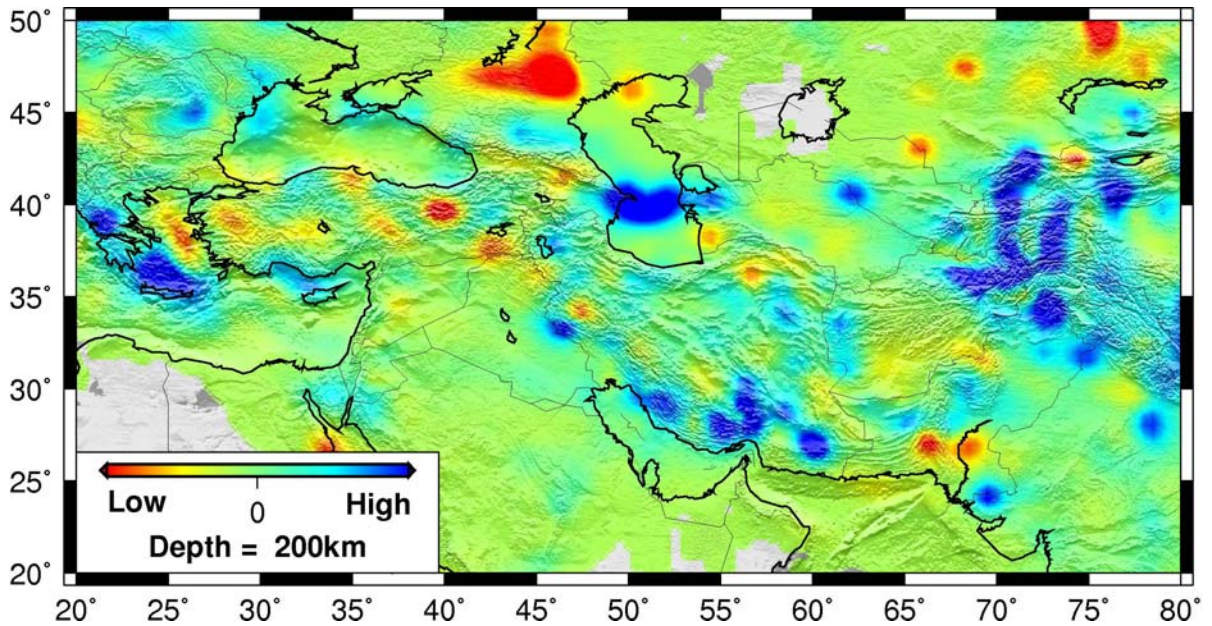
### 3.3 Upper Mantle tomography

We use a modified version of Zhao's tomographic method (Zhao et al., 1992, 1994; Zhao, 2001; Sun and Toksöz, 2006), which allows for 3-D velocity variations everywhere in the model, and can accommodate velocity discontinuities, to construct the upper mantle structure of the Mideast regions. The velocity structure is discretized using a 3-D grid. The velocity perturbation at each point is calculated by linear interpolation of the velocity perturbations at surrounding (adjacent) grid nodes. The velocity perturbations at grid nodes are the unknown parameters for the inversion procedure. To calculate travel-times and ray paths accurately and rapidly, the pseudo-bending technique (Um and Thurber, 1987) is used iteratively. We correct for station elevations by including station correction terms in the inversion. The nonlinear tomographic problem is solved by repeated linear inversions. At each iteration, perturbations to hypocentral parameters and velocity structure are determined simultaneously.

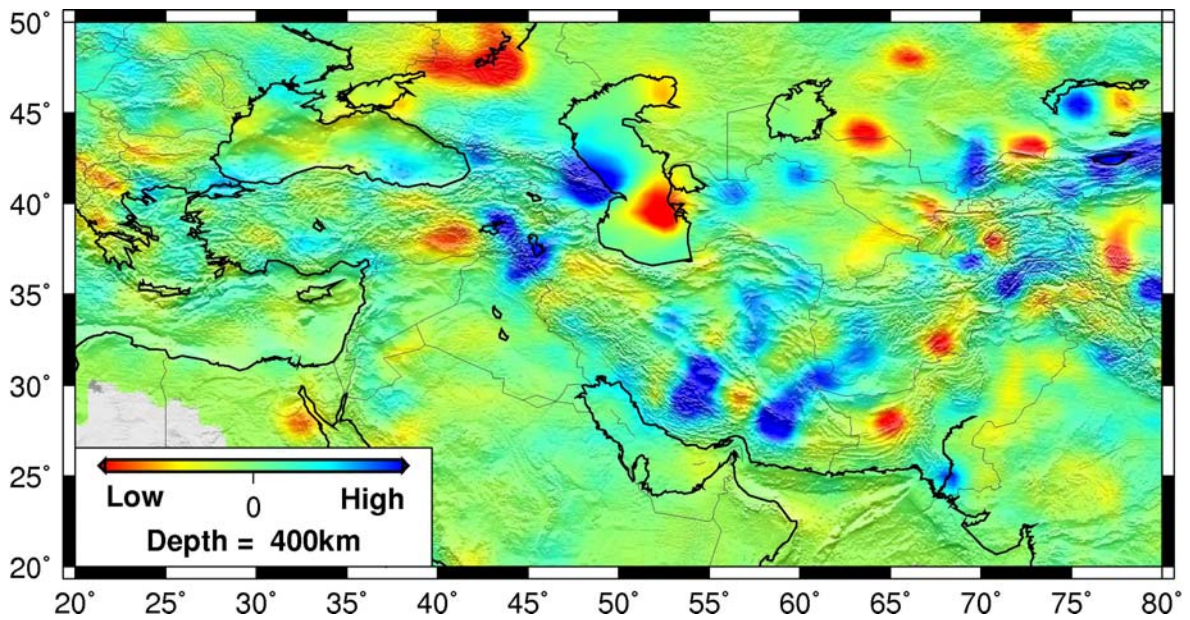
Figures 5 to 8 show the P-wave velocity profiles at depths of 50, 200, 400 and 600 km.



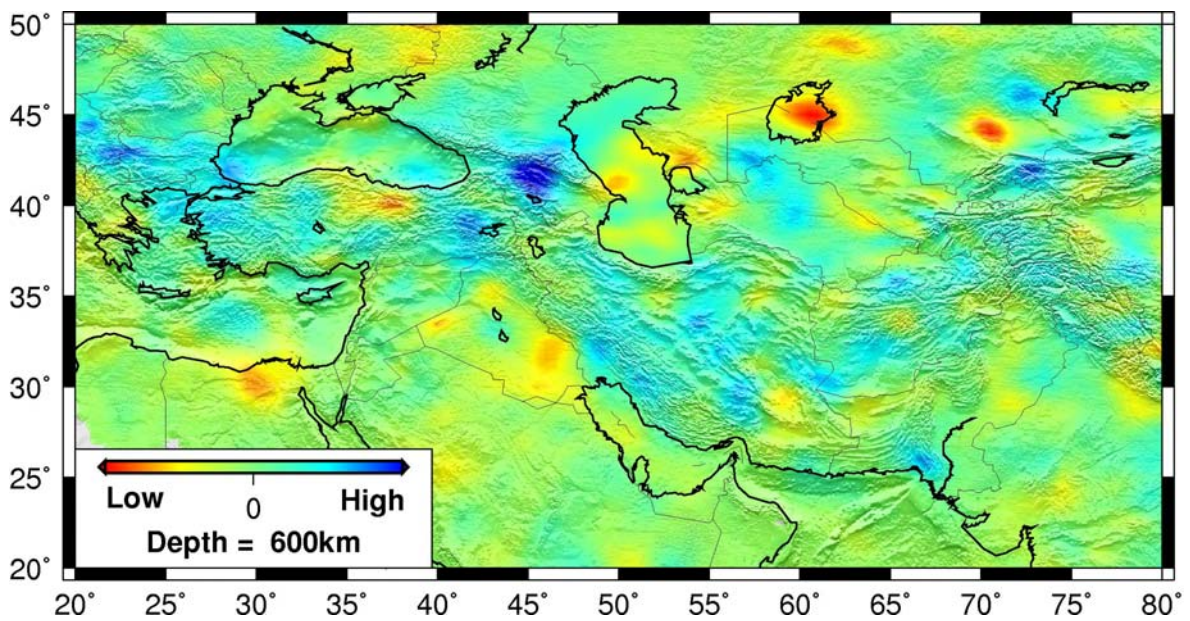
**Figure 5. Velocity profile at depth of 50 km.**



**Figure 6. Velocity profile at depth of 200 km.**



**Figure 7. Velocity profile at depth of 400 km.**

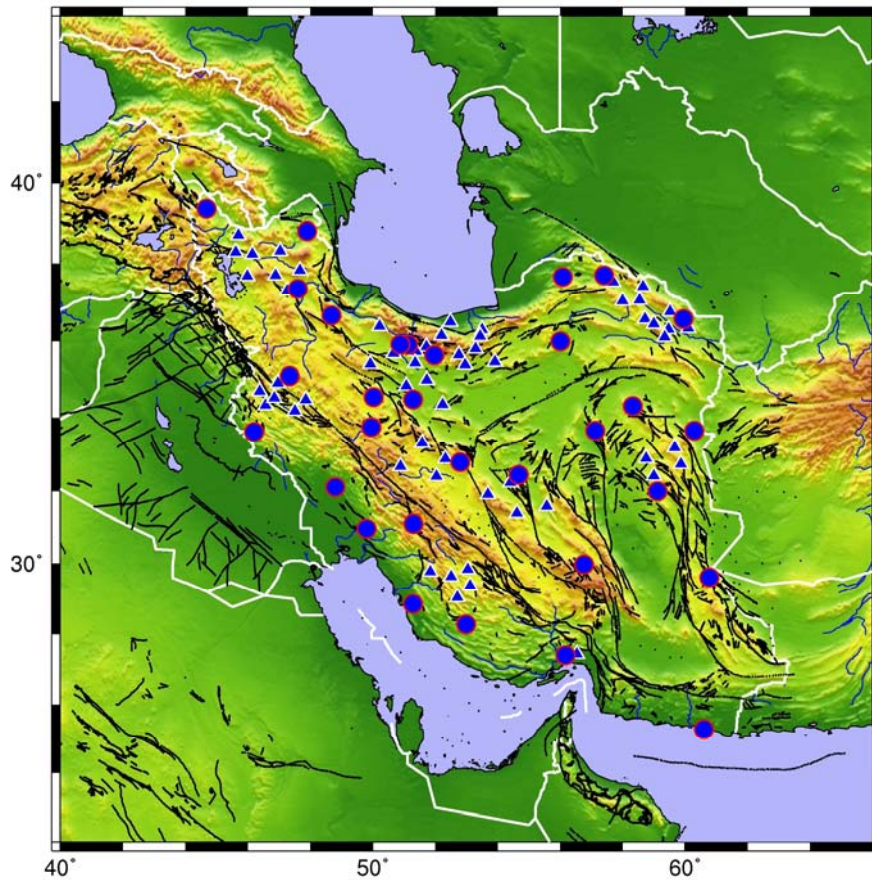


**Figure 8. Velocity profile at depth of 600 km.**

## 4. Models for Iran

### 4.1 Data description

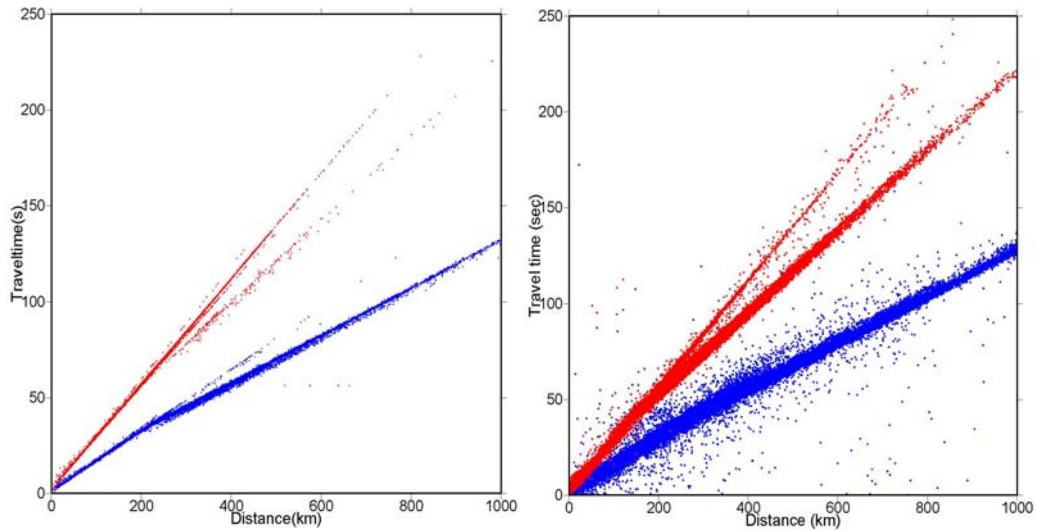
Most important new data for this study comes from two seismic networks in Iran: Iranian Seismic Telemetry Network (ISTN) and Iran National Seismic Network (INSN) (Figure 9). The ISTN was founded in 1995. In total 73 stations are grouped into ten sub-networks distributed in most part of Iran. There are 67 three-component short period stations and six analogue stations. The International Institute of Earthquake Engineering and Seismology operates the other national wide seismic network: INSN. All the 30 INSN stations use broadband seismographs. Figure 9 shows the station distributions in both networks.



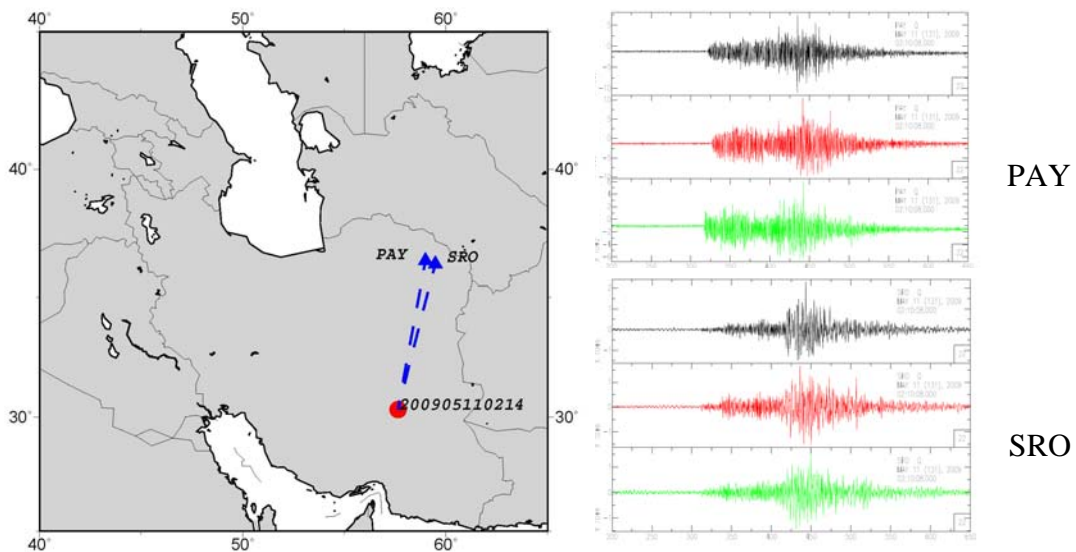
**Figure 9. Seismic stations in Iran. INSN stations are in circles and ISTN stations are in triangles.**

The Institute of Geophysics at Tehran University serves as the central processing facility for ISTN. Under their management, the data quality has improved significantly in the past few years. They make travel time, amplitude, and, for selected events, waveform data available through their website. Figure 10 shows examples of travel time data from

both ISTN and INSN networks. Figure 11 shows the seismograms of a magnitude 4.9 event that occurred on 05/11/2009 at two stations: PAY and SRO. PAY is a short period station. SRO is a borehole station with a broad band sensor. The data from Iran is the newest and most important part of this study.



**Figure 10. Available P and S travel times from INSN (left) and ISTN (right) networks.**



**Figure 11. Seismograms recorded at short period station (PAY) and in a borehole at SRO.**

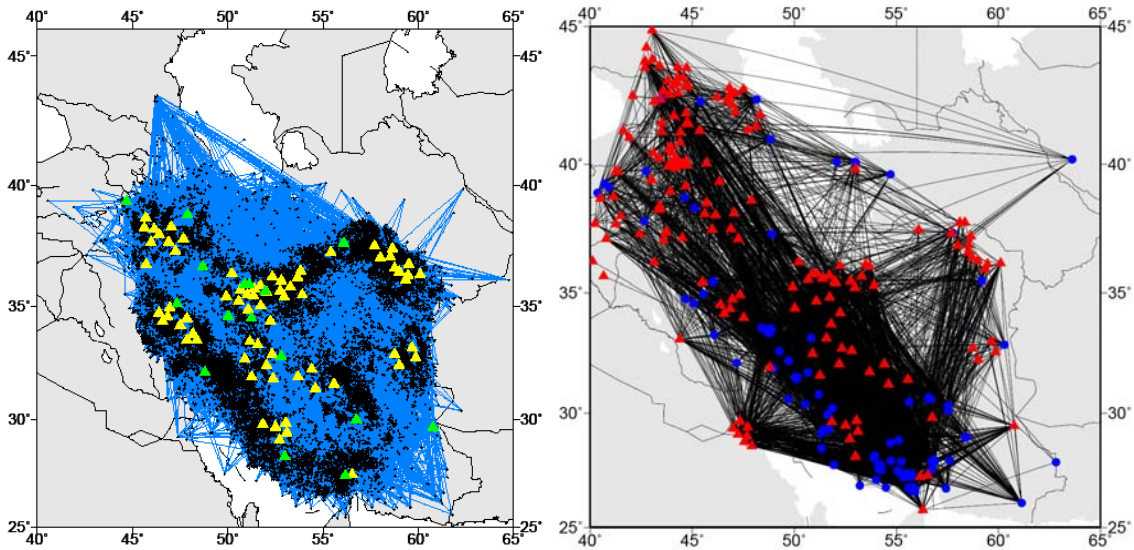
## 4.2 Seismic tomography of the Iran and surrounding regions

The tomography research was completed based on our team's extensive experience in travel-time tomography for the crust and mantle (Sun et al., 2004; Sun, 2005; Sun and Toksöz, 2006; Sun et al., 2008a, b; Zhang and Thurber, 2003, 2005, 2006). The increased number of seismic stations, well-located earthquakes, and high-quality P and S wave travel-times have been used for generating a tomographic model of Iran and surrounding areas. The travel-time tomography was carried out in two steps. First, using the regional travel-time data, the 3-D crust uppermost mantle P and S wave velocity structures were determined. Second, the models were extended into the mantle transition zone by adding teleseismic travel-time data and constraining the crust and uppermost mantle velocities.

### 4.2.1 Data for seismic tomography

For the crust and upper mantle tomography we have used travel-time data from the 1964 -- 2007 ISC/EHB database (30,000 P and S arrivals and 9,000 phases, Figure 10) regional (e.g. Eastern Mediterranean, Gulf Seismic Forum) databases and from national centers (e.g., KOERI in Turkey, National Seismological Center in Oman), and data from Iran.

The two main seismic networks in Iran: Iranian Seismic Telemetry Network (ISTN) and Iran National Seismic Network (INSN) (Figure 9) have more than 100 digital 3-component stations. We collected phase arrivals and maximum amplitudes reported by both two networks from July, 2004 to May, 2009. This dataset provides abundant local and regional phase arrivals which are useful to extract crustal seismic structure. ISTN catalogue includes 25,471 earthquakes between Feb, 2006 and Apr, 2009. More than 240,000 arrivals were reported. There are about 160,000 P arrivals and 80,000 S arrivals. The INSN reported 3342 events before 2008 and used 15,112 P wave arrivals, 4927 S wave arrivals to locate events. All ray paths in this dataset are plotted in Figure 12.

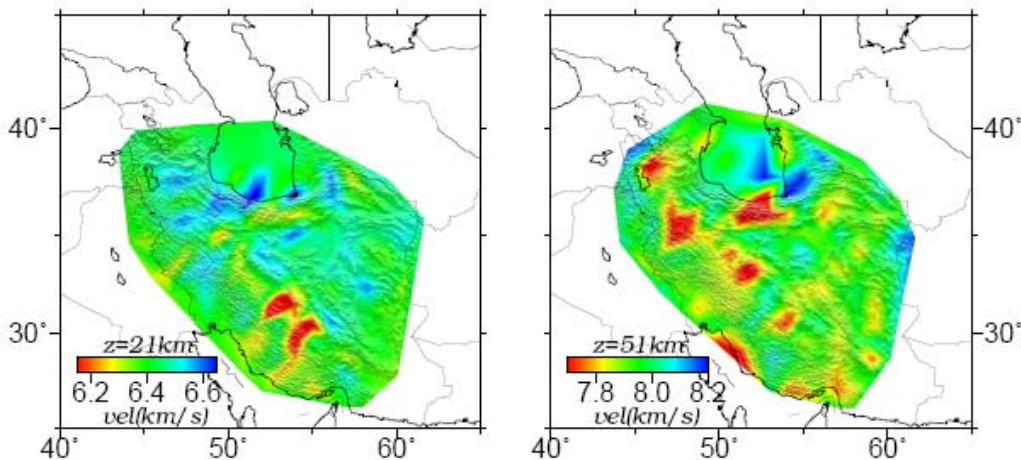


**Figure 12.** Available ray paths from the Iranian networks (left) and EHB database (right).

## 4.2.2 P- and S-Wave Travel-Time Tomography of the Crust and Upper Mantle

A high-resolution tomographic model for the heterogeneous crust is constructed by iterative, non-linear tomography. To generate adequate starting models for the nonlinear inversion we combine pertinent information from global (Mooney et al., 1998; Stevens et al., 2001; Ritzwoller et al., 2002), regional, and local crust and uppermost mantle models. Next, we use the adaptive moving window (AMW) approach (Sun et al., 2004, 2008b) to obtain crustal velocities and Pn and Sn models from a 1-D Monte Carlo inversion of local ( $\leq 20^\circ$ ) arrival time data in the whole region, building these into the next model (Model #2). The third step is a tomographic inversion of the local and regional arrival time data for 3-D variations in the P and S wave speed, using Model #2 as the initial input model.

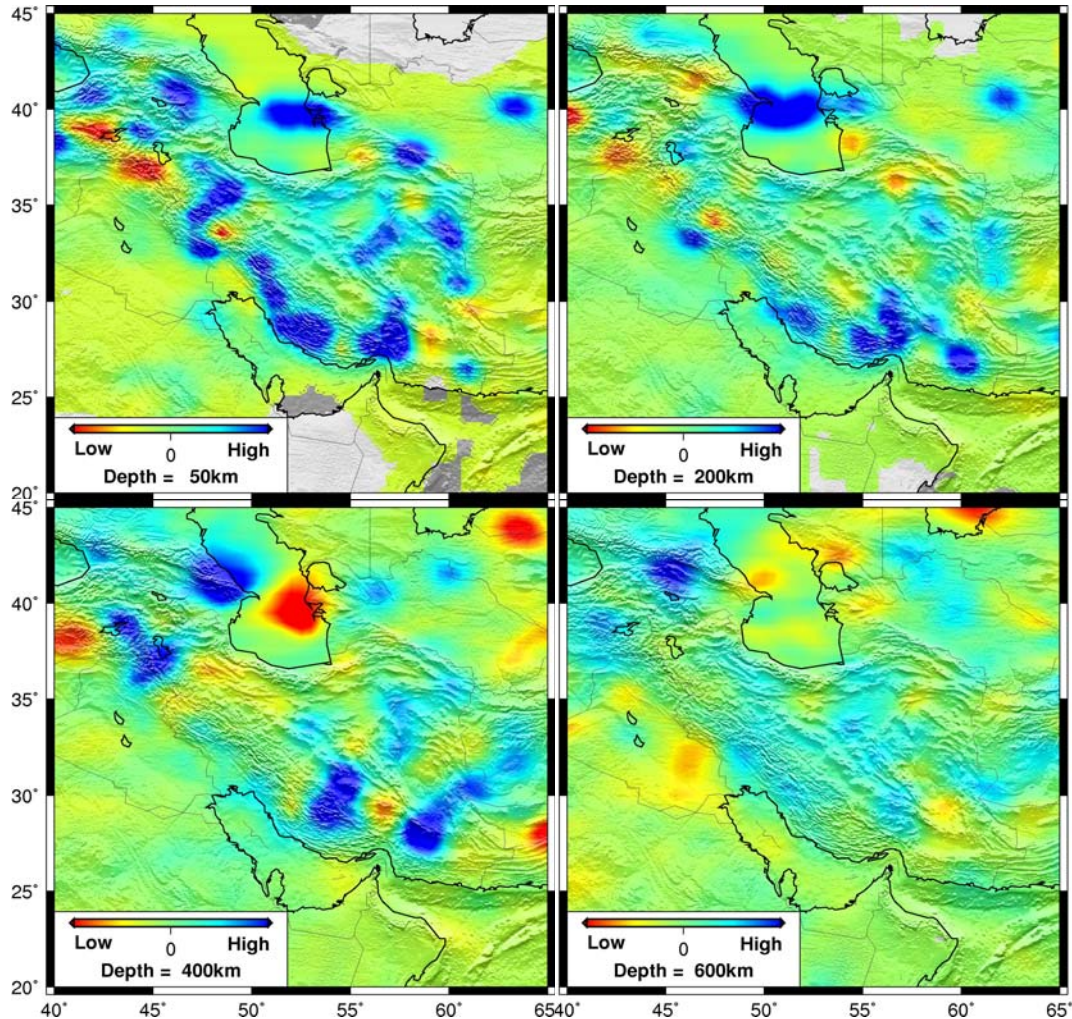
For this purpose we use a modified version of Zhao's tomographic method (Zhao et al., 1992, 1994; Zhao, 2001; Sun and Toksöz, 2006), which allows for 3-D velocity variations everywhere in the model and can accommodate velocity discontinuities. The velocity structure is discretized using a 3-D grid. The velocity perturbation at each point is calculated by linear interpolation of the velocity perturbations at surrounding (adjacent) grid nodes. The velocity perturbations at grid nodes are the unknown parameters for the inversion procedure. To calculate travel-times and ray paths accurately and rapidly, the pseudo-bending technique (Um and Thurber, 1987) is used iteratively. We correct for station elevations by including station correction terms in the inversion. Finally, we apply an adaptive-mesh double-difference tomography method (Zhang and Thurber, 2005) to the newly assembled data set to further improve event locations and velocity models. Figure 13 shows Vp models at depths of 21 and 51 km obtained using the double-difference tomography method.



**Figure 13. P-wave models at depths of 21 and 51 km obtained using the double-difference tomography method.**

The extension of the velocity models into the mantle transition zone are accomplished by adding the teleseismic data from the 1964–2007 ISC/EHB database, which contains over

10 million P-wave travel-times, and from data sets obtained by the various national data centers. Figure 14 shows a tomographic model for the region.

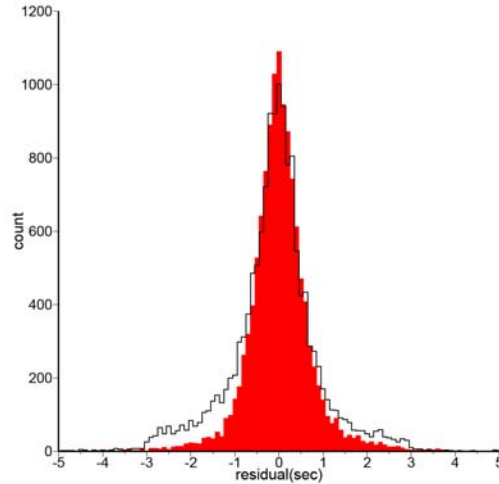


**Figure 14. Horizontal cross sections through the mantle from preliminary results of P-wave travel-time tomography. Blue colors represent regions with faster than average velocities, and red color show regions with slower than average velocities.**

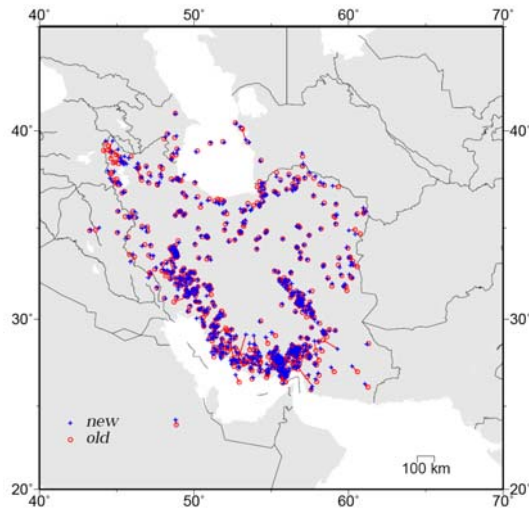
### ***4.3 Relocation of Earthquakes in Iran***

The local and regional networks (shown in Figure 9) provide good data coverage (Figure 10) to relocate earthquakes using high-quality P and S wave travel-times. We apply an adaptive-mesh, double-difference tomography method to the Iranian data set to improve event locations and to obtain preliminary velocity models (Zhang and Thurber, 2003; 2005). Figure 15 shows the comparison of travel-time residuals using 1D and 3D models for event relocation. The 1D P and S models used for relocation were adopted from Crust 2.0 and the 3D models were obtained while simultaneously locating events. The overall residuals reduce from 1.2 s for 3D models to 0.9 s for 1D models. Figure 16

shows the relocation of a group of events in the year of 2006 and an example of event relocation using 1D and 3D models. The average difference between the original locations and the 3D relocations is 25 km.



**Figure 15. Comparison of residuals using 1D model (solid line) and 3D model (red).**



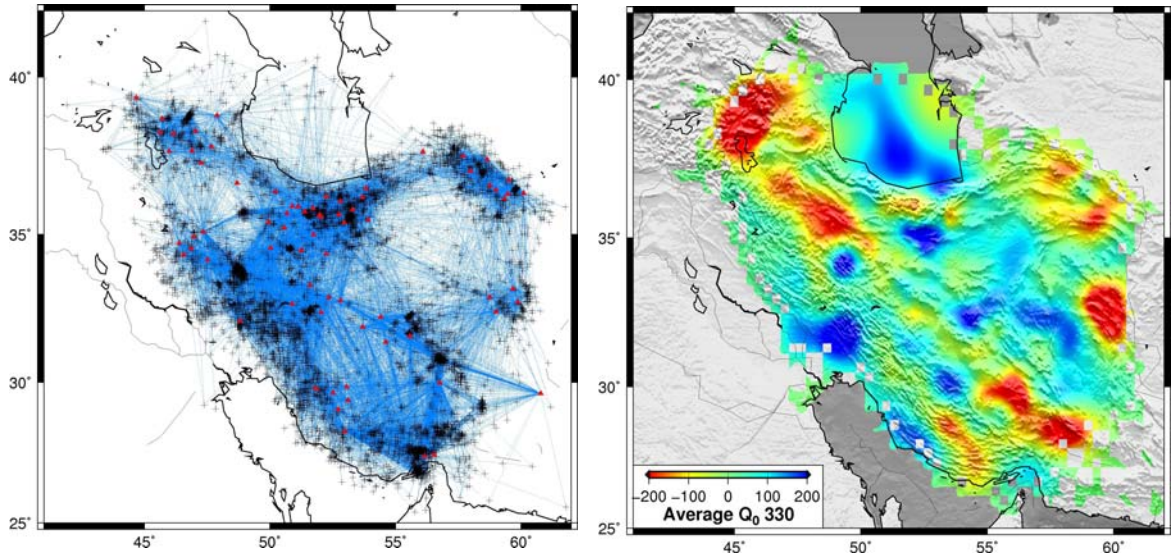
**Figure 16. Map of an earthquake relocation in Iran, where the USGS locations are shown as red circles, and the locations found using 3D tomographic models are shown in blue crosses.**

#### ***4.4 Attenuation tomography***

To obtain the attenuation structure in Iran and the surrounding region, we used the amplitude data of the Iranian National Seismic Network between 2004 and 2007 and the data of the ISTN network from 2006 to 2009. These ML amplitudes are the maximum amplitudes ( $A_{\mu}$ ) measured on the horizontal components of short-period instruments and are used for calculating ML magnitudes. At short epicentral distances, the amplitudes correspond to the short-period  $S_g$  wave. At longer distances, these are coincident with the

Lg wavetrain. These waves travel within the crust and are thus sensitive to crustal variations in attenuation.

In our test modeling we selected data using the following criteria: each event was recorded by at least 2 stations; each station recorded at least 5 events; the reported period of the Sg wave at maximum amplitude was between 0.1 s and 2.0 s; epicentral distance was between 20 km and 800 km and event depth was less than 20 km. We obtained 26,183 amplitude readings from 7,118 events recorded by 75 stations, as shown in Figure 17.



**Figure 17. Left: Distribution of ray paths, stations (triangles) and events (crosses). Right: Preliminary imaged  $Q_0$  lateral variations. Blue corresponds to a high value and red to low. The average  $Q_0$  is 330.**

We use the attenuation tomography method of Pei et al. (2006, 2009) using ML amplitude data to estimate  $Q$  of the crust in Iran. The attenuation of seismic waves between station  $i$  and source  $j$  can be estimated from the amplitude spectrum  $A_{ij}(f)$ , where  $f$  is the frequency at which the amplitude is measured. Amplitude is influenced by several factors: the size of the event, station site and instrument gain, geometrical spreading and seismic attenuation. With these, the amplitude can be written as (e.g., Sanders, 1993; Haberland and Rietbrock, 2001):

$$A_{ij}(f) = O_j(f)S_i(f)G_{ij}(R)B_{ij}(f, R) \quad (1)$$

where  $O_j(f)$  represents the event amplitude at frequency  $f$ ,  $S_i(f)$  is the site and instrument gain,  $G_{ij}(R)$  is the geometrical spreading factor at distance  $R$ , and  $B_{ij}(f)$  is seismic attenuation. The event radiation pattern has not been included. Since we average over many stations around the focal sphere, this is supported by the relative consistency of ML magnitude estimates made at different seismic stations. We include frequency dependence in all terms except geometrical spreading.

We use a frequency dependent source spectrum  $O(f)$  (Brune, 1970; Hanks and Wyss, 1972) that can be described by a long-period spectral level  $\Omega_0$  and the corner frequency  $f_c$  as  $O(f) = \Omega_0 / (1 + (f^2 / f_c^2))$ . The  $\Omega_0$  is proportional to the seismic moment  $M_0$ . Frequency  $f$  and corner frequency  $f_c$  can be considered as constants for one event, and frequency  $f$  corresponds to the frequency of the recorded waveform at maximum amplitude of Sg. We assume source spectrum  $O(f)$  is frequency independent for one event, although it is variable for different events.

The site response,  $S_i(f)$ , includes the effect of instrument gain and site amplification. The spectral dependence of  $S_i(f)$  is unknown; however, the ML bandwidth is small (0.5 Hz to 2 Hz) and the instrument response is reasonably flat within this range. We therefore assume  $S_i(f)$  is also independent of frequency.

The geometrical spreading,  $G(R)$ , is considered to be proportional to  $(1/R)^k$ , where  $k$  is the geometrical spreading coefficient, which chiefly depends on wave type.  $k = 1.0$  implies spherical spreading of body waves and  $k = 0.5$  represents cylindrical spreading of Lg and surface waves.

The attenuation of S waves,  $B_{ij}(f)$ , is attributed to the quality factor  $Q$ . We assume a frequency dependent  $Q$  and model it as an exponential decay (e.g. Sanders, 1993; Haberland and Rietbrock, 2001):

$$B_{ij}(f) = \exp\left(-\frac{\pi f R}{v Q}\right) = \exp(-c Q_0^{-1} R) \quad (2)$$

where  $c = \pi f^{1-\eta} / v$ ,  $Q = Q_0 f^\eta$ ,  $Q_0$  and  $\eta$  are the S-wave  $Q$  at 1 Hz and its frequency exponent, respectively.

Using equation (2) and taking the natural logarithm of equation (1) yields

$$Y_{ij} = \ln A_{ij}(f) - \ln G_{ij}(R) = a_i + b_j - Q_0^{-1} c R \quad (3)$$

where the station term is  $a_i = \ln S_i(f)$  and event term is  $b_j = \ln O_j(f)$ . Equation (3) is very similar to the Pn travel time equation (Pei et al., 2007), thereby allowing us to resolve lateral variation of  $Q_0$  using the Pn tomography method. We first estimate the average  $Q_0$  by fitting the line  $Y - cR$ . Then we discretize the crust into small 2-D pixels of  $X$  km by  $Y$  km, then equation (3) can be rewritten as

$$Y_{ij} = a_i + b_j - \sum_k Q_{0k}^{-1} c R_{ijk} .$$

By subtracting  $Y_{ij}$  from the average model (fitted line), a perturbation equation can be determined as follows

$$\Delta Y_{ij} = \Delta a_i + \Delta b_j - \sum_k \Delta Q_{0k}^{-1} c R_{ijk} . \quad (4)$$

Equation (4) can be recast as  $A x = d$ , where  $d$  is the data vector,  $x$  is the vector of unknowns and  $A$  represents the model parameters. We apply damped LSQR (Paige and Saunders, 1982) with preconditioning to find the least-squares solution. Laplacian smoothing was applied in the inversion (Lees and Crosson, 1989). We apply this approach to data from Iran as shown in Figure 17.

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## **Appendices**

### **Appendix I**

#### **Deliverables:**

- 1) Event catalogue in the Mideast region; (Mideast\_pha.txt)
- 2) Crust velocity model in the Mideast; (Mideast\_crust\_vel.txt)
- 3) Mantle velocity model in the Mideast;  
(Mideast\_mantle\_vel.txt, at 50, 200, 400 600 km)
- 4) Pn and Sn velocities in the Mideast; (Mideast\_Pn\_vel.txt; Mideast\_Sn\_vel.txt)
- 5) Relocated events in Iran. (iran\_relocated\_catalog.txt)

**Appendix II**

Tomographic Pn and Sn Velocity beneath  
Continental Collision Zone in Middle East and Middle Asia

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**Abstract** We have obtained Vp and Vs velocity images of the uppermost mantle beneath the Middle East region by performing tomographic inversion of both Pn and Sn travel times. 654,999 Pn arrivals and 121,838 Sn arrivals were selected from the ISC/EHB database and the Annual Bulletin of Chinese Earthquakes. Average Pn and Sn velocities are 8.04 km/s and 4.60 km/s, respectively, and maximum velocity perturbations are about 6%. Pn velocity has a close relationship with topography. In general, the mountains with high elevation have low velocity, while the seas, basins and plains with low elevation have high velocity because the mountains are collision zones with strong tectonic activity and the stable plates have low elevation. The large tectonic lines are boundaries between high and low Pn velocity, such as the Hellenic arc, Cyprean arc, Zagros suture, the Charman fault, and the Hari Rod fault. The Sn velocity has a very similar pattern to Pn velocity.

Key Words: Tomography, Pn and Sn velocity, Middle East  
Primary Index Term: 8180 Tomography; 7218 Lithosphere

## 1. INTRODUCTION

The Arabian-Eurasian plate boundary is extremely complex, and it is an ideal region to study a young (geologically) continent-continent collision belt. The current tectonics of the region are controlled by the collision and continuing convergence of the Arabian and Eurasian plates (Figure 1). The Arabian and Eurasian plates collided in the early Miocene, after the Neotethys Sea was subducted beneath Eurasia (Bird, 1978; Şengör and Yılmaz, 1981; Jackson and McKenzie, 1984; Dewey et al., 1986). Pre-, syn-, and postcollision tectonics produced very complex structures in the region. Over the last decade, a number of seismic studies have examined the crust and upper mantle structure beneath the Middle East to constrain the nature of the Arabian-Eurasian collision zone. Large scale surface wave tomography studies have shown variable crustal thickness and upper mantle velocities (Ritzwoller and Levshin, 1998; Pasyanos et al., 2001; Villasenor et al., 2001; Pasyanos and Walter, 2002; Shapiro and Ritzwoller, 2002; Alinaghi et al., 2007; Reiter et al., 2005; Reiter and Rodi, 2006, 2008). Regional-scale surface wave tomographic studies further highlight the complexity of the collision zone (Mindevalli and Mitchell, 1989; Rodgers et al., 1999; Mohktar et al., 2001; Maggi and Priestley, 2005) showing a thickened crust under the Caucasus and Zagros, and low shear velocity beneath the Turkish and Iranian plateaus. Hearn and Ni (1994), Ritzwoller et al. (1998), Al-Lazki et al. (2003; 2004), and Phillips et al. (2007) found slow Pn velocities ( $\leq 8$  km/s) beneath the Anatolian plateau, northwestern Iran, the Greater Caucasus, and southwestern Arabia. The Pn velocities beneath northern Arabia and the Caspian region are faster than average (Al Lazki et al., 2004; Ritzwoller et al., 2002; Angus et al., 2006). This high degree of variability suggests that the Earth structure may be extremely complicated in the region. The studies mentioned above present consistent results for the crust and uppermost mantle seismic properties on a regional scale. Significant variation in waveforms, observed particularly in short-period seismograms, over propagation paths that are close to each other, suggests structural variations over short distances at regional boundaries. Delineating these features requires seismic data from dense local and regional seismic networks. During this year we were able to increase the phase (arrival time) data significantly by adding readings from Iran and other regional networks.

In this work, we use many more Pn and Sn arrivals reported in the ISC/EHB (International Seismological Centre) database and many national seismic networks in the Middle East to construct a tomographic image of Pn and Sn seismic velocity beneath the Middle East and surrounding regions.

## 2. DATA

An important task under this project is to collect arrival time data from seismic stations situated in more than 20 countries in the region. A significant number of these stations are in networks that are relatively new and whose data are not available from global data centers such as IRIS or ISC. In the past year, Iran started to make phase data available from its networks. In Iran there are seven regional networks, deployed around major cities, with a total of 50 stations. Recently, Tehran University was designated as the central location that would collect and integrate data from the regional networks into a central database. The use of data from stations in Iran required deciphering of nomenclature and resolution of some apparent discrepancies between data from individual networks and the central database.

Our study area is defined by 20°N - 53°N and 15°E - 90°E. The epicenter distance range of the travel times used in this study is 2°-12° for first P and 2°-15° for first S. We chose the lower limit to remove Pg and Sg phases, and the upper limit to include only the observations along the linear portion of the travel time curve. Within this distance range the Pn phases are first arrivals and the Sn phases are first S-wave arrivals. We also remove the data point with smaller Pg/Sg residuals than Pn/Sn. Starting models were obtained by a linear fit to the travel time – epicentral distance curve (Figure 2. ). To ensure the data quality, we impose further conditions: each event has at least 3 Pn and 2 Sn readings. In addition, the travel time residuals to the fitted line used in this paper are limited to the range of +/-6 s for Pn and +/-7 s for Sn because the error in picking the Sn phase is usually larger than Pn. The data set included 654,999 Pn times from 42,984 earthquakes recorded by 1263 stations (Figure 3. ), and 121,838 Sn times from 22,283 earthquakes recorded by 887 stations (Figure 3. ). The depth of each event is less than 40 km. The standard deviation of the travel time residuals is 2.20 s for Pn and 4.40 s for Sn for original data, 1.68s and 2.87s for selected data.

### 3. METHOD

Following Hearn’s approach and computation method (Hearn and Ni, 1994; Hearn, 1996), modified by Pei et al. (2007), we invert Pn and Sn travel time residuals for lateral velocity variation and anisotropy within the mantle lid. The Pn and Sn are modeled as refracted rays traveling along the Moho discontinuity (head waves). The variation of seismic velocity within the uppermost mantle is parameterized by subdividing the surface of the uppermost mantle into a 2-D grid of 30'×30' cells. The Pn/Sn travel-time residuals are described as the sum of three time terms:  $t_{ij} = a_i + b_j + \sum d_{ijk} \cdot s_k$ , where  $t_{ij}$  is the travel time residual for the ray from event  $j$  to station  $i$ ;  $a_i$  is the static delay for station  $i$ , depending on the crustal thickness and velocity beneath the station;  $b_j$  is the static delay for event  $j$ , not only depending on the crustal thickness and velocity beneath the event, but also on the event focal depth;  $d_{ijk}$  is the distance traveled by ray  $ij$  in mantle cell  $k$ ;  $s_k$  is the slowness (inverse velocity) perturbation for cell  $k$ . The unknown quantities in the equation are  $a_i$ ,  $b_j$ , and  $s_k$ . The sum is calculated over all cells through which the ray travels in the uppermost mantle. Further details of the tomography technique used here can be found in Hearn and Ni (1994) and Hearn (1996).

We have performed a series of tests by choosing various cell sizes and different damping constants for Pn and Sn. The adopted cell size is 30'×30', as stated before, and the damping constants for slowness perturbation were set to 1300 for Pn and 800 for Sn. The total number of model parameters is 7162 for Pn and 6278 for Sn.

### 4. RESULT and DISCUSSION

Fitting a straight line to the data set of Pn and Sn travel time versus epicentral distance, we obtained an average Pn velocity of 8.0 km/s (Figure 2. ) and average Sn velocity of 4.58 km/s (Figure 2. ). The average P and S velocities within the crust were set to 6.3 km/s and 3.6 km/s, respectively.

#### 4.1 Pn velocity

Pn lateral velocity variations (Figure 4. ) were obtained by tomographically inverting the Pn travel time residuals. Relative to the mean value of 8.04 km/s, the Pn velocity

perturbations range from  $-0.50$  km/s ( $-6.2\%$ ) to  $0.38$  km/s ( $4.7\%$ ). The standard deviation of travel time residuals decreased from  $1.38$  s to  $2.20$  s, after the inversion. The general features of Pn velocities are similar to those of previous work (Al-Lazki et al. 2004; Reiter et al. 2005) except that Figure 4 has finer spatial resolution.

Pn velocity has close relation with topography. The mountains with high elevation have low velocity from Anatolia plateau, Iran plateau to Tibet plateau, while the seas, basins and plains with low elevation have high velocity such as the Mediterranean Sea, Black Sea, Caspian Sea, Arabian Gulf and Amu-Darya Basin, Tarim basin and Indian plain. In general, these mountains not only consist of many subplates or blocks, but also are the collision zone between the Eurasian Plate and African Plate, Arabian Plate and Indian Plate, and have strong tectonic activity, accompanied with many volcanoes and volcanic rocks. In contrast, the plates are old cratons, and tectonically stable, so have high Pn velocity.

Pn velocities are low under eastern Anatolia, NW Iran, and the Southern Caucasus. Isolated low velocity anomalies exist along the Dead Sea Fault Zone. A prominent low velocity feature is observed just to the south of the Caspian Sea. The Iranian plateau is characterized by lower than average Pn velocity. Pn velocities are high under the Arabian platform, the Persian Gulf, and under the Zagros fold belt. These velocities are most likely associated with the top of the Neotethys lithosphere. In the north, the higher velocities are under the Black Sea, Rioni and Kura Basins between the Greater and Lesser Caucasus, under the southern Caspian, and Kyzyl and Kara Kum Basins of Turkmenistan. These are most likely the remnants of the Paleotethys (Gülen, 1989). In the east, prominent high velocities exist under the Amu-Darya Basin, North Afghan Platform and Indian Plate. Low velocities exist in the Afghan Accreted Terranes and Hindu Kush. Central Tibet also has low velocity between two high velocity regions under Tarim basin and Indian Plate.

The large tectonic lines are boundaries between high and low Pn velocity. The Hellenic arc and Cyprean arc in the Mediterranean Sea are boundaries between the African Plate, with high  $V_{pn}$ , and Anatolian plate, with low  $V_{pn}$ ; the Zagros suture is the boundary between Arabian Plate, with high  $V_{pn}$ , and North West Iranian Plate, Central Iranian Microplate and Lut block, with low  $V_{pn}$ ; the Charman fault and Hari Rod fault are the boundaries between Afghan Accreted Terranes, with low  $V_{pn}$ , and North Afghan Platform and Indian Plate, with high  $V_{pn}$ .

## 4.2 Sn velocity

Figure 5. shows lateral variation of Sn velocity. Relative to the mean value of  $4.60$  km/s, the Sn velocity perturbations range from  $-0.28$  km/s ( $-6.0\%$ ) to  $0.19$  km/s ( $4.1\%$ ). The standard deviation of travel time residuals for Sn decreased from  $4.40$  s to  $2.25$  s after inversion. Sn velocities shown in Figure 5. are similar to Pn (Figure 4. ). In general, the regional features correlate quite well. Some isolated anomalies (e.g. Dead Sea Fault Zone) are very similar between Pn and Sn. In west Tibet, a similar pattern is comparable to previous Sn study (Pei et al., 2007). A low velocity anomaly exists under central Tibet between high Sn velocity regions under the Tarim basin and Indian Plate. This study shows more detailed structure in the Indian plate with many more ray paths. Compared with Pn velocity, the resolution is a little lower for Sn because there are few Sn ray paths.

## **5. CONCLUSION**

We have imaged both the Pn and Sn velocity structure of the Middle East using a large amount of data. We discuss in detail the relationship between velocity variations and tectonic activity, and we have arrived at the following conclusions:

- 1) The average Pn and Sn velocities are 8.04 km/s and 4.60 km/s, respectively. The maximum velocity perturbations are about 6%.
- 2) Pn velocity has a close relation with topography. In general, the mountains with high elevation have low velocity, while the seas, basins and plains with low elevation have high velocity because the mountains are collision zones with strong tectonic activity and the stable plates have low elevation.
- 3) The large tectonic lines are boundaries between high and low Pn velocity, such as the Hellenic arc, Cyprean arc, Zagros suture, Charman fault, Hari Rod fault.
- 4) The Sn velocity has a very similar pattern to Pn velocity.

## **ACKNOWLEDGMENTS**

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## Figure captions

Figure 1. Topographic map of the Middle East and surrounding regions. White lines denote the location of known faults, while the black dots represent the epicenter locations of earthquakes in the region.

Figure 2. Travel times for Pn (bottom) and Sn (upper)

The data shown within the box outlining the lower fitted line were selected to invert Pn lateral velocity. The average Pn velocity of 8.04 km/s was determined from the slope of this line. The data within the box outlining the upper fitted line represent the Sn arrivals selected to invert for Sn lateral velocity perturbations. The average Sn velocity of 4.60 km/s was determined from the slope of this line. The white straight lines are fitted using the data only within the boxes of Pn and Sn, respectively.

Figure 3. Ray paths for Pn (upper) and Sn (bottom) travel times

654,999 Pn rays were obtained from 42,984 events (black crosses) recorded by 1,263 stations (red triangles), and 121,838 Sn rays were obtained from 22,283 events recorded by 887 stations. The data were selected for 2-12 degrees of epicentral distance and residuals of less than 7.0s for Pn, and 2-15 degrees and residuals of less than 6.0s for Sn, and each event was recorded by at least 3 stations for Pn and 2 stations for Sn.

Figure 4. Imaged Pn velocity lateral variations

The average Pn velocity is 8.04 km/s and variation corresponds with color. Red represents lower velocity than average and blue denotes higher velocity.

Figure 5. Imaged Sn velocity lateral variations

The average Sn velocity is 4.60 km/s and variation corresponds with color. Red represents lower velocity than average and blue denotes higher velocity.

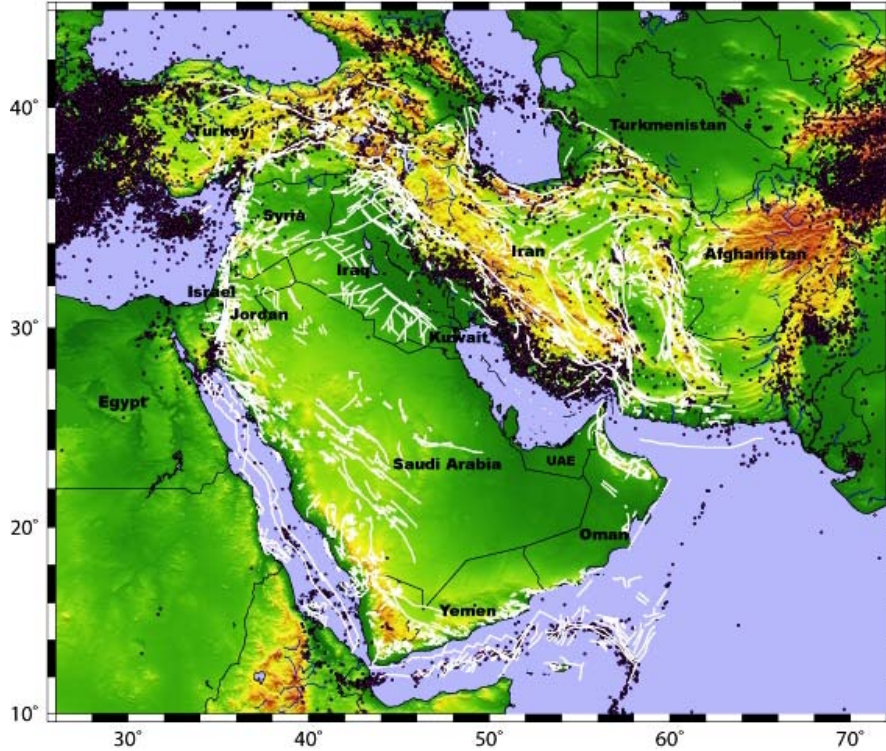


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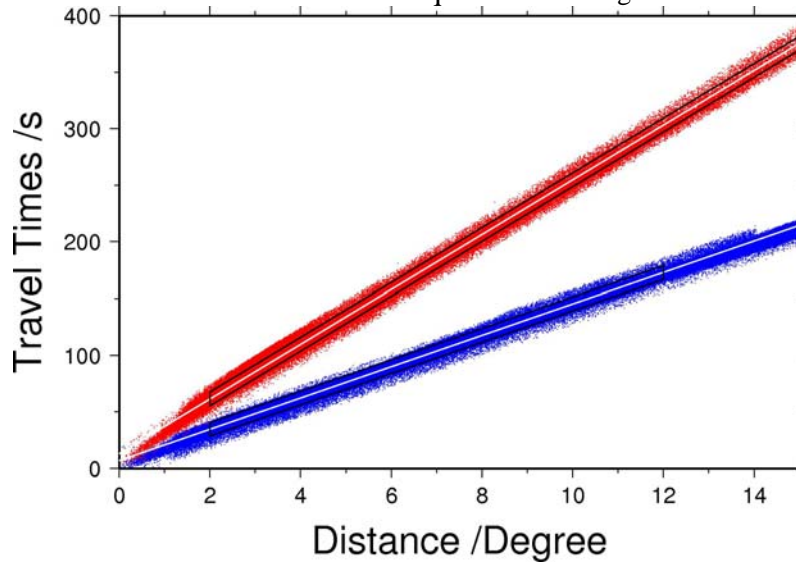


Figure 2. Travel times for Pn (bottom) and Sn (upper)

The data shown within the box outlining the lower fitted line were selected to invert Pn lateral velocity. The average Pn velocity of 8.04 km/s was determined from the slope of this line. The data within the box outlining the upper fitted line represent the Sn arrivals selected to invert for Sn lateral velocity perturbations. The average Sn velocity of 4.60 km/s was determined from the slope of this line. The white straight lines are fitted using the data only within the boxes of Pn and Sn, respectively.

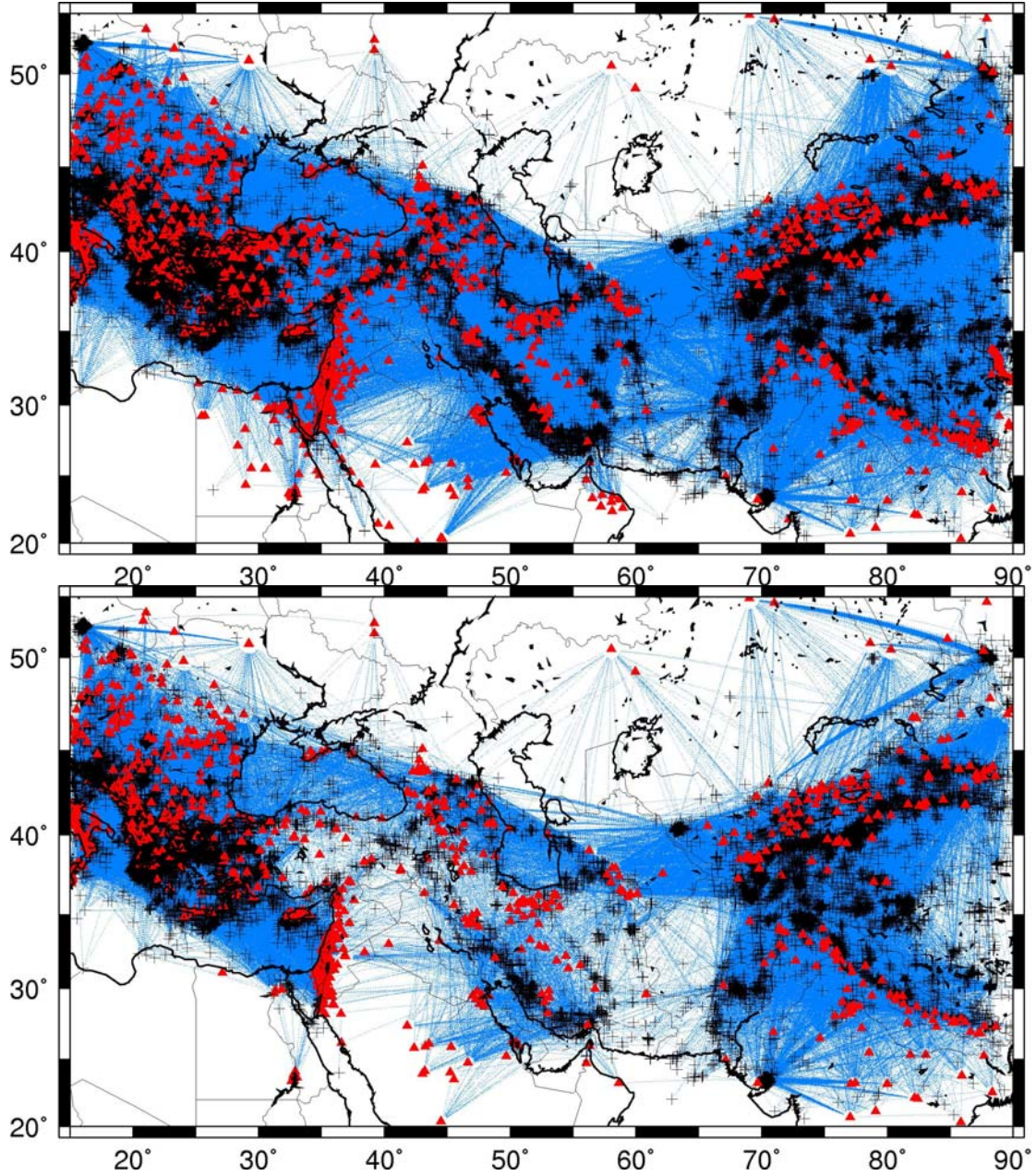


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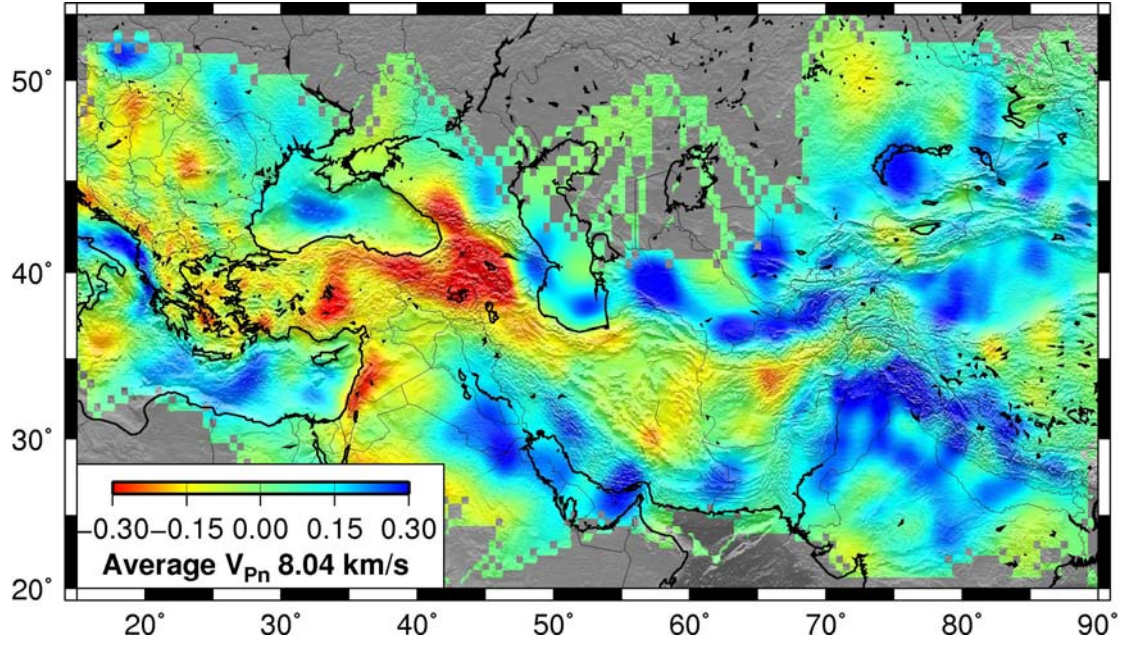


Figure 4. Imaged Pn velocity lateral variations  
The average Pn velocity is 8.04 km/s, and variation corresponds with color. Red represents lower velocity than average and blue denotes higher velocity.

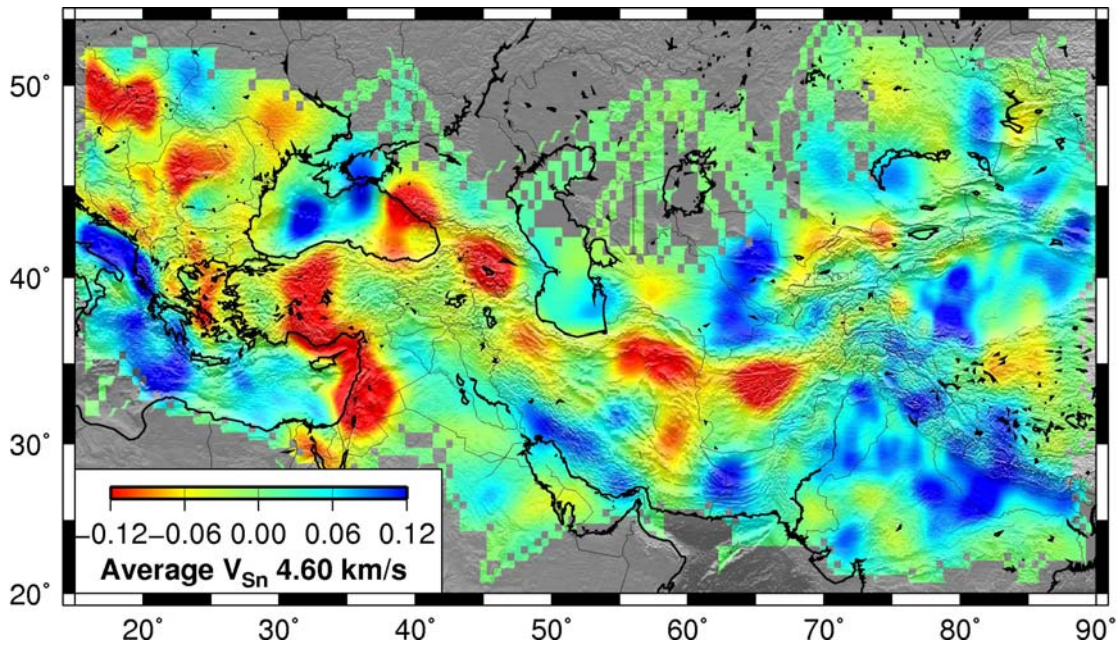


Figure 5. Imaged Sn velocity lateral variations  
The average Sn velocity is 4.60 km/s and variation corresponds with color. Red represents lower velocity than average and blue denotes higher velocity.