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New seismic imaging of some tectonic zones in the Iranian Plateau

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Abstract

The Iranian Plateau is characterized by diverse tectonic domains, including the continental collisions (e.g. the Zagros and Alborz Mountains) and oceanic plate subduction (e.g. Makransubduction zone). To derive a detailed image of the crust–mantle (Moho) and lithosphere–asthenosphere (LAB) boundaries in some tectonically units of the Iranian Plateau, we used a large number of S receiver functions obtained from teleseismic events recorded at 68 national permanent stations (19 broadband and 49 short period stations). The S receiver functions clearly imaged the base of the crust and lithosphere and their variations within the different tectonic zones of the Iranian Plateau. Our new seismic images show a significant variation of the lithospheric thickness in the different geological features. The most complex structure was detected beneath the Zagros Mountains where the Arabian Plate is believed to underthrust beneath centralIran. We found the thickest crust under the Sanandaj-Sirjan metamorphic zone (SSZ)which proposes the overthrusting of the crust of central Iran into the Zagros crust along the main Zagros thrust (MZT), in agreement with the results of Paul et al., (2010). Furthermore, our results clearly show a shallow LAB at about 80-90 km depth beneath the Alborz, the central domain (CD) and central Iranian micro plate (CIMP) zones. Based on our results, the Arabian LAB, beneath the Zagros fold and thrust belt (ZFTB), SSZ and the Urumieh-Dokhtar magmatic assemblage (UDMA) is 200 km and may contain a dipping structure at depths ranging from 100 beneath the ZFTBto 150 km beneath the SSZ and the UDMA. This dipping structure interpreted as the presence of remnants of the fossil Neo-Tethys subduction. The location of the boundary between the Arabian and central Iranian lithospheres is beneath the UDMA, which is shifted northeastward relative to the surficial expression of the MZT.

Keywords: Iranian Plateau, Lithosphere, Crust, S receiver function

1 Introduction

The Iranian Plateau is extremely complex due to its location at the junction of the Arabian and Eurasian plates. A seismically active continental-continental plate boundary, in southwestern Iran is the most well-known tectonic unit of the Iranian Plateau, which has been resulted from the collision of the Arabian Plate with the central Iranian plate after the subduction of Neo-Tethys Ocean beneath Eurasia. The subduction of the Neo-Tethys ocean started in late Jurassic period (e.g. Berberianand King, 1981), and the onset of the closure of the oceanic domain occurred in late Cretaceous (Agard et al., 2005). The time of collision is controversial among the geologists and varies from late Cretaceous (Berberianand King, 1981; Agard et al., 2005) to Oligocene–Miocene (Koop and Stoneley, 1982). The Zagrose collision zone comprises of three majorsub-parallel elements, which are nominated from SW to NEas the Zagros fold and thrust belt (ZFTB), the Sanandaj-Sirjan metamorphic zone (SSZ) and the Urmieh-Dokhtar magmatic assemblage (UDMA) (Figure 1).

The main Zagros thrust (MZT) along the

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NE edge of the Zagros region, which separates the ZFTB and the SSZ (Stöcklin, 1968; Ricou et al., 1977; Berberian, 1995), is a major geological boundary. Geological (e.g. Agard et al., 2005) and geophysical (e.g. Paul et al., 2006) researches suggest that the MZT is deeply rooted and coincides with the suture zone between the Arabian and Iranian plates. The global positioning system (GPS) studies show that the convergence between the Arabian and Eurasian plates is trending north to north-northeast at velocity ranges from 23 to 25 mmyr^{-1} (Masson et al., 2007, Vernant et al., 2004).

In the north, the E-W trending Alborz Mountains belt with approximately 200 km wide and approximately 600 km long were formed when Gondwana collided with Eurasia in the Late Triassic (Sengor et al., 1988). This region still represents a strong active deformation between the stable central Iranian Plate and the Eurasian Plate to the north, and undergoes extensive crustal deformation and shortening. There are several studies that indicate crustal thickening beneath the Alborz Mountains (Dehghani and Makris, 1984; Sobouti and Arkani-hamed, 1996; Doloei and Roberts, 2003; Rham et al., 2007; Sodoudi et al., 2009; Shad Manaman et al., 2011).

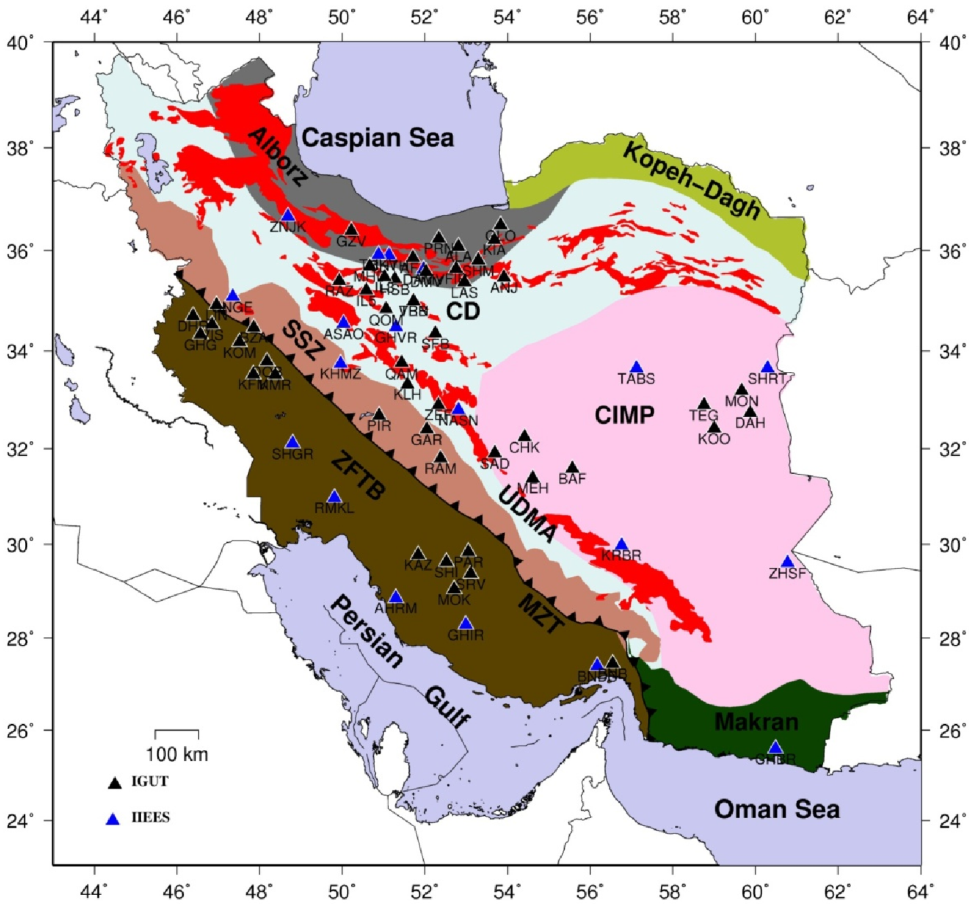

Figure 1. The main tectonic features of the Iranian Plateau and seismic stations which are used in this study. The black triangles are the Institute of Geophysic, University of Tehran (IGUT) stations and the blue triangles are the International Institute of Earthquake Engineering and Seismology (IIEES) stations.The following abbreviations are used in the figure: central domain (CD); the Urumieh–Dokhtar magmatic assemblage (UDMA);Sanandaj–Sirjan metamorphic zone (SSZ); central Iranian micro plate (CIMP); Zagros fold andthrust belt (ZFTB) andmain Zagros thrust (MZT).

Central domain (CD) of Iran, entrapped between the Zagros and Alborz regions and central Iranian micro plate (CIMP) are relatively aseismic and rigid blocks. Central Iran was a part of the Arabian Plate during the Precambrian–Paleozoic times. Late Paleozoic to Triassic rifting resulted in separation and northward movement of central Iran away from the Arabian Plate. Central Iran was assembled back with the Arabian Plate after the closure of Neo-Tethys and its collision with Arabia (Nabavi, 1976; Stöcklin, 1968).

The deformation in the Iranian collision zone is accommodated on relatively narrow areas, providing a natural laboratory for the study of continental collisions and processes related to the early phases of mountain building (Walker and Jackson, 2004). Many questions are still remain about this collision: contribution of continental subduction in the transition from oceanic subduction to collision; Tethyan slab being attached to the Arabian margin or detached from it; the response of the lithospheric mantle to shortening; and the role of delamination in the formation of lithosphere beneath the Iranian Plateau. Previous studies show differences in upper-mantle structure beneath the Arabian and Iranian plates, but the resolution of documented depth for Moho and lithospheric discontinuity is too low. To improve understanding of mantle processes beneath this region, it is necessary to improve the resolution of Moho and LAB depths and indicate clear crustal and lithospheric structures.

2 Previous studies of the Iranian Plateau structure

The crustal thickness of the Iranian Plateau has been studied by Dehghani and Makris (1984). They prepared the first Moho map of the Iranian Plateau using gravity and seismic data. Their results indicate that crustal thickness is 55 km beneath the MZT, 35 km below the Alborz mountains and less than 40 km beneath the Lut block.

Snyder and Barazangi (1986) were focused on the Zagros belt and estimated the maximum Moho depth of 65 km beneath the MZT. Hatzfeld et al., (2003) inverted arrival times of microearthquakes and receiver functions beneath the Ghir station and measured the Moho depth of 46 ± 2 km under this station in central Zagros. Pn velocity studies by Al-Lazki et al., (2004) revealed high velocities in the Arabia Plate, including the Zagros (8.1–8.4 km s^{-1}) and beneath most of the Iranian Plateau (with normal velocities of 7.9–8.1 km s^{-1}). They believe that transition from high to normal velocities occurs at the MZT in the northwestern part of Zagros, while in the southern Zagros, the high-velocity area extends 100 km north of the MZT. Paul et al., (2006) determined the Moho depth beneath a profile across the Zagros belt using receiver functions method. They estimated a crustal thickness of approximately 45 km beneath the High Zagros, rapidly reaching to approximately 70 km beneath the SSZ before thinning to approximately 42 km beneath the UDMA and central Iran. They proposed this thickening results from the overthrusting of the crust of central Iran into the Zagros crust along the MZT. Based on S and Preceiver function methods, Sodoudi et al., (2009) presented clear images from the Moho and LAB under the Alborz region. Their results show variation of crustal thickness from 47 km beneath central Iran to about 54 km beneath central Alborz and a relatively thin lithosphere of 90 km beneath the high central Alborz zone.

S-wave velocities have been estimated using surface wave analysis beneath the Iranian Plateau (e.g. Curtis et al., 1998; Maggi and Priestley, 2005; Kaviani et al., 2007; McKenzie and Priestley, 2007; Shad Manaman and Shomali, 2010). Most recently, Priestley et al., (2012) mapped lateral variations of shear wave speed in upper mantle beneath the Zagros region using a large multi-mode surface wave data set. They showed that upper mantle is slow for most of the Iranian Plateau, but a high shear wave speed lid extending to approximately 225 km depth beneath the Zagros. They converted the shear wave speed profiles to temperature profiles to identify the base of the lithosphere. Their results showed that lithosphere is less than approximately 120 km thick over the region, except for a thick lithospheric root beneath the Zagros, implying that shortening of the mantle is accommodated by thickening of the lithospheric.

In this paper we use S receiver function method to image the high resolution crustal and lithospheric structures of the region. We show the existence of a thickened lithosphere beneath the Zagros Mountains, but there is no evidence for such thickness to the north and east beneath central Iran and Alborz Mountains.

3 Data processing

We analyze the data from 49 permanent short-period seismic stations operated by the Institute of Geophysics, University of Tehran (IGUT) and 19 permanent broadband seismic stations belong to the International Institute of Earthquake Engineering and Seismology (IIEES). They are located in different tectonic zones of Iran (Figure 1). The IGUT and IIEES stations are equipped with SS-1 Kinemetrics, and CMG-3T Guralpsensors, respectively. Teleseismic data, which have been recorded between 2005 and 2011 at seismic stations, were used in this study. More than 900 events with magnitudes greater than 5.7 (m_b) at epicentral distances between 60° and 85° were utilized for calculating the S receiver functions (Figure2). These epicentral distances were suggested by Faber andMüller (1980) as the best distance range to detect the converted S-to-P phases from the Moho and LAB. We adopted the method used by Sodoudi et al., (2006a, 2006b, 2009).

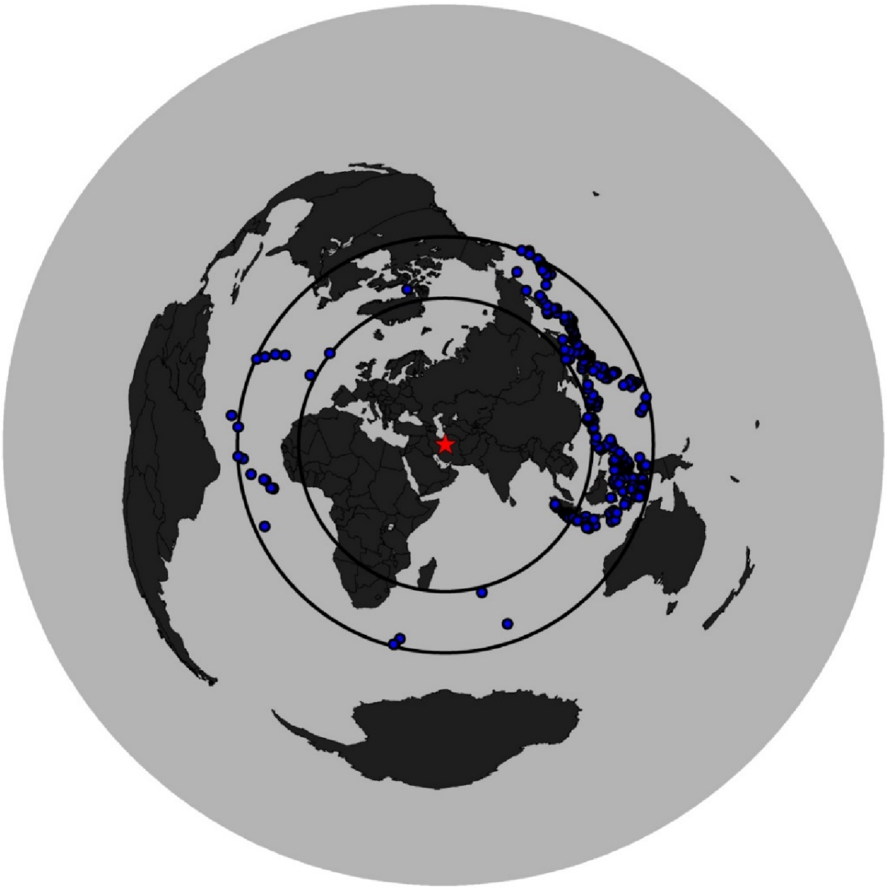

Figure 2. Azimuthal distribution of the teleseismic events recorded by the IGUT and IIEES networks between 2005 and 2011 used in this study in relationship to the studied area (red star). The black circles mark the 60° and 85°epicentral distances. Events used for this study are shown by the blue circles.

4 Results

S receiver function method was applied for mapping the crust–mantle and lithosphere–asthenosphere boundaries (e.g. Farra and Vinnik 2000; Vinnik et al., 2004, Kumar et al., 2005a, 2005b; Yuan et al., 2006; Sodoudi et al., 2006a, 2006b, 2009, 2011; Kumar et al., 2006, 2007; Heit et al., 2007; Kawakatsu et al., 2009; Abt et al., 2010; Geissler et al., 2010). The LAB boundary is often invisible in the PRFs due to the crustal multiples arriving at the same time, and heavily disturbing the time window of the LAB arrival. The S receiver function technique uses the teleseismic S waves to extract the S-to-P converted phases. This technique can resolve the Moho and LAB boundaries because the S-to-P converted phases arrive earlier than the direct S wave; therefore they are free from crustal multiples. We selected a time window of 200 s in length (100 s before the S onset) and calculated the S receiver functions (SRFs). A low-pass filter of 4 s was applied to the data. Calculation of SRFs was performed in different steps including restitution, coordination rotation, and deconvolution. We rotated the ZNE components into the local LQT ray-based coordinate system using theoretical back azimuth and observed incidence angle (see Kumar et al., 2006). The incidence angle was determined considering the minimum energy in the L component at arrival time of the S phase. Deconvolving the Q component from the L component equalizes the different sources. To compare the SRF directly with the PRF, we reversed the polarity of the S receiver function amplitudes and the time axis. In the resulted SRF, positive amplitudes indicate velocity contrasts, with velocity increases downward and vice versa. As the S-to-P conversions are generally weak, a number of records must be summed to obtain a good signal-to-noise ratio. Individual and stacked SRFs for each station were determined. Before stacking, the SRFs were moved out and corrected to the reference slowness of 6.4 sdeg^{-1}. An example of computed SRFs for ZNJK station in the Alborz region is shown in Figure 3.

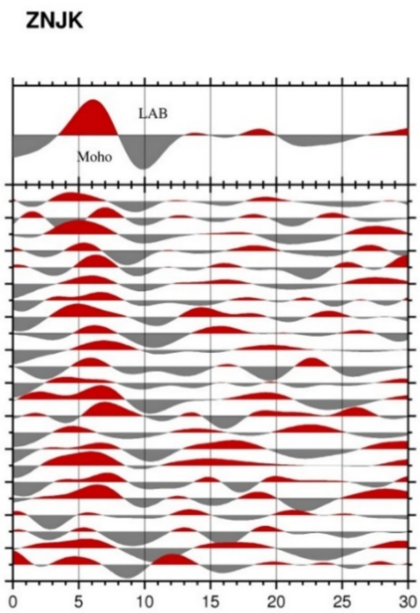

Figure 3. Individual and stacked SRFs for ZNJK station located in the Alborz. The traces are arranged with increasing observed back azimuth and filtered with a low-pass filter of 4 s. Stacked and individual traces showclear S-to-P conversions, corresponding to the Mohoand LAB boundariesat approximately 6and 10 s, respectively.

To present the variety of SRF results in different tectonic zones, the SRFs arranged in each tectonic zone by the latitude of their piercing points at 100 km depth (see Figure 4). Two phases are visible in these data. The first phase (black) indicates the Moho boundary, while the second stable and coherent phase (grey) most probably stems from the LAB. As Figures 4a and 4b show the arrival time value of the second phase in the Zagros (approximately 13s) and SSZ (approximately 15s) is large in comparison with approximately 9 s for other tectonically zones such as CIMP, CD and Alborz. As Mohammadi et al., (2013b) showed the LAB depth in SSZ zone is deeper than other tectonic zones.

The SRFs presented in Figure 4 show that the difference between the delay time of the converted phases in Moho and LAB boundaries (in the ZFTB and SSZ units) is large in comparison with other tectonic zones. The LAB converted phase seems to be much deeper beneath the ZFTB (approximately13 s) and SSZ (approximately 15 s) compared to those that are observed beneath the Alborz, CD and CIMP tectonic zones (approximately 9 s). Therefore, the observed large delay time for the second phase (grey) of the SRFs in the ZFTB and SSZ regions proposes a thick-lithosphere or deeper LAB. Later, we will discuss about other clear grey phases that labeled by the red question marks beneath the ZFTB, SSZ, UDMA zones (Figures 4a, b, f).

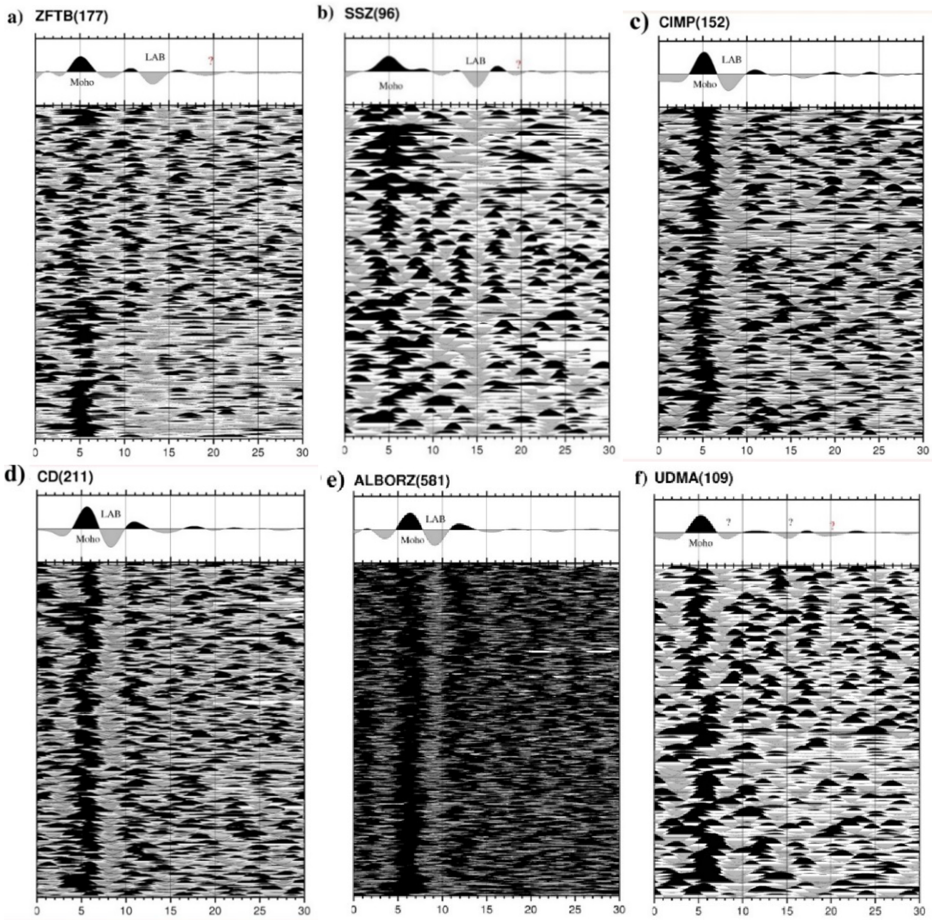

Figure 4. Individual and stacked SRFs for group of stations which are located in different structural zones. (a) ZFTB zone, (b) SSZ unit, (c) CIMP zone, (d) Central domain (CD), (e) Alborz region, (f) UDMA zone. In each graph the horizontal axis is in second, representing the delay time of the converted phases. Figures b, d, and e extracted from Mohammadi et al., (2013b).

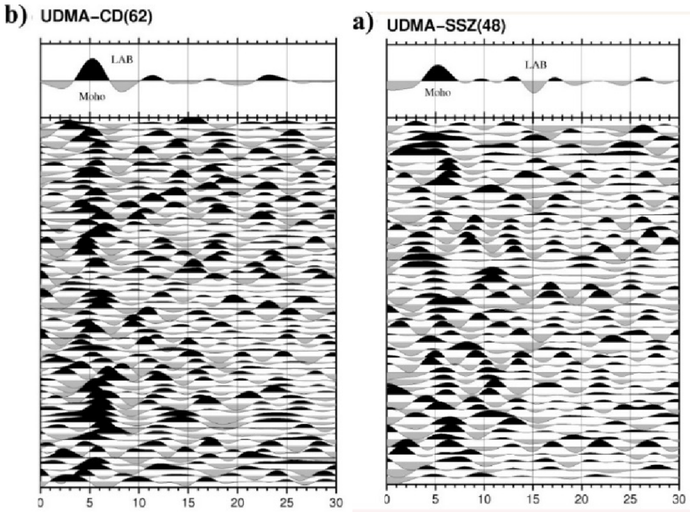

Figure 5. (a) Individual and stacked SRFs in the UDMA zone. (a) SRFs for which piercing pointsare close to the SSZ unit, (b) SRFs for which piercing pointsare close to the CD region.

Contrary to the other zones, in the UDMA zone instead of a clear LAB phase, two phases are observed on the stacked SRF (Figure 4f). These two phases which labeled with black question mark suggest that the LAB depth probably is different in various parts of the UDMA zone. For more detail, we grouped the stations according to their distance from the SSZ and CD zones. Considering this criterion in piercing points, the SRFs were sorted and plotted in Figure 5. Two weak LAB phases which are observed in Figure 4f, clearly separated in Figure 5. Figure 5 shows that the LAB depth in the western part of the UDMA zone (Figure 5a) is deeper than the LAB depth in the eastern part (Figure 5b).

5 Discussions

Presented receiver functions here propose an average crustal thickness of approximately 40km for the ZFTB, CD and CIMP zones and approximately 55 km for the Alborz Mountains. In the Zagros Mountains, the Moho depth increases from the ZFTB to the SSZ unit. To transform time into depth, we used the IASP91 reference model (Kennett and Engdahl, 1991). Results of our S receiver functions show a relatively shallow LAB at about 90 km depth beneath the whole Alborz, CD and CIMP zones. Beneath the Zagros, a more complex structure is detected. Considering the S receiver functions, the LAB (strong grey phase) seems to be located at approximately130 km depth beneath the ZFTB zone, while beneath the SSZ and the UDMA zones the LAB depth is increased to approximately150 km.

Using three cross sections perpendicular to the Zagros collision zone, Mohammadi et al., (2013a), suggest a 200 km thick lithosphere beneath the Zagros collision zone and a thin lithosphere of about 80–90 km beneath central Iran and the Alborz Mountains. Their results show the presence of remnants of the fossil Neo-Tethys subduction at depths ranging between approximately 100 km and 150 km within the Arabian lithosphere. Their results suggest that dipping structure can be seen beneath the Zagros, SSZ and UDMA. In our results the observed phases at approximately 13 s in the ZFTB which increases to approximately 15 s beneath the SSZ and UDMA interpreted as LAB. These phases may present the remnants of the fossil Neo-Tethys subduction at depths ranging between approximately100 km and 150 km within the Arabian lithosphere. Other later phases beneath these zones (indicated by the red question marks in the ZFTB, SSZ, and UDMA in Figures 5a, b, f) may show deeper LAB for these regions. The location of the boundary between the Arabian and Iranian plates is estimated to be beneath the UDMA, which is shifted northeastward relative to the MZT.

Mohammadi et al., (2013a) believed that the boundary between the Arabian and Iranian plates seems to be located northeast of the UDMA in the northwest of Zagros and southwest of the UDMA in the central Zagros region. The idea of a break-off of the oceanic Neo-Tethyan slab beneath central Iran results in an asthenospheric upwelling and thinning of the Iranian lithosphere beneath CD, CIMP, and the Alborz zones. Our results are in good agreement with recent study of mantle beneath Iran (Priestley et al., 2012). Using surface wave data, Priestley et al., (2012) indicated that a high shear wave velocity lid is extended to about 225 km depth beneath the Zagros, while a slow upper mantle is extended down to the depth of about 120 km beneath the whole Iranian Plateau. They believed that lithospheric thickening beneath the Zagros is taking place in response to the shortening that occurred in the mantle.

The large crustal thickness beneath the Zagros suture zone indicates the influence of the crustal thickening and shortening beneath the SSZ in the Arabian–Eurasian Plate boundary. Beneath the SSZ, the crust is significantly thickened and this thickening clearly seen in the S receiver functions beneath this zone, the feature that was interpreted by Snyder and Barazangi (1986) as a result of the plastic deformation of the flexible lower crust due to horizontal compression during the collision. Also, this conclusion agrees with the results of Mohammadi et al., (2013a, b). Paul et al. (2006, 2010) showed that the crustal thickness increases to a maximum value of approximately 70 km beneath the SSZresulted from underthrusting of the Zagros crust with the crust of central Iran along the MZT.

The CD and CIMP is undeformed and aseismic regions have the crust with 40 km thick. This is supported by the results derived by Paul et al. (2006, 2010) and Afsari et al. (2011), confirming that these regions (where topography is the lowest) are characterized by relatively lower P- and S-wave velocities compared to Arabia and Eurasia (e.g. Al-Damegh et al., 2004; Al-Lazki et al., 2004; Mooney and Detweiler 2005; Alinaghi et al., 2007; Kaviani et al. 2007; Shad Manamanet al., 2011). Our results in the Alborz region shows a strong Moho deepening in response to the shortening related to the collision of the Arabian–Eurasian plates (Sodoudi et al., 2009). Previous studies in the Alborz region (e.g. Dehghani and Makris 1984; Jackson et al., 2002; Sodoudi et al., 2009) found that there is no deep crustal root beneath the high-elevated central Alborz. Sodoudi et al. (2009) showed significant local crustal thickening (approximately 67.5 km) under the Damavand volcano which is interpreted as the magmatic addition at the base of the crust beneath the volcanic region. Our results show a thick crust of approximately55 km beneath the high topography part of central Alborz, which is in good agreement with those values obtained by Sodoudi et al. (2009) and Radjaee et al., (2010). The S receiver functions exhibit the existence of a sharp velocity discontinuity in the upper mantle at a depth of approximately 90 km, which we interpret as the base of the lithosphere in this region. New seismic images presented in this study show different values for the lithospheric depth beneath the Zagros and central Iran. We believe that the MZT does not separate these two different lithospheres. This result is in contrast to the previous studies (e.g. Paul et al., 2006; Shad Manaman and Shomali 2010). The location of the boundary between the Arabian and Iranian lithospheres is estimated to be beneath the UDMA, which is shifted northeastward relative to the MZT.

The Paleocene and Neocene volcanic activity in central Iran and Quaternary volcanism in northern Iran (Berberian and King, 1981) suggest a hot mantle (thin lithosphere). Kaviani et al. (2007) used surface wave dispersion data and showed the lack of high shear wave velocities at depths greater than 150 km and suggest that the slab could be detached. However, they believed that the absolute S-wave velocities measured in their study are too high to support the hypothesis of mantle lid delamination in this transition zone between Arabia and central Iran.

6 Conclusion

Applying the S receiver function method to the data recorded by 68 short period and broad-band permanent stations within some tectonic zones in the Iranian Plateau provide new images of lithosphere structure and crustal thickness. The SRFs indicate a coherent Sp conversion of the Moho and LAB. The results are used to deduce valuable information about the tectonics of the study area. Analysis of the S receiver functions documents significant increase in crustal thickness beneath the SSZ which is interpreted by overthrusting of the crust of central Iran to crust of the Zagros along the MZT. This result agrees with the model of the crustal scale by Paul et al. (2006 and 2010). We present a high resolution lithospheric structure beneath the region. These images propose that the Arabian lithosphere, beneath the ZFTB, the SSZ, and the UDMA zones, is 200 km thick and may contain a dipping structure at depths ranging between 100 km beneath ZFTB and 150 km beneath the SSZ and the UDMA. The boundary between the passive margin of the Arabian platform and the microblocks of central Iran is in the UDMA. This hypothesis implies that the Arabian lithosphere extends to approximately150 km northeast of the MZT (Kaviani et. al., 2007). This may presents at depths below the crust, the Arabian-Eurasian continental collision results in extensive subducting of the Arabian lithosphere beneath the UDMA. Even though, the presence of the thin lithosphere below central Iran and the Mountains Alborz would support the asthenospheric upwelling to subcrustal levels due to the slab detachment beneath the Zagros collision zone.

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