CONTINENTAL LITHOSPHERE STRUCTURE BENEATH THE IRANIAN PLATEAU, FROM ANALYSIS OF RECEIVER FUNCTIONS AND SURFACE WAVES DISPERSION

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ABSTRACT :
More than one year of teleseismic waveforms recorded by 12 broad band stations of Iranian National Seismic Network (INSN) from 2004 to 2005 were used to study structure and thickness of the crust in different parts of Iran by joint inversion of receiver functions and regional surface wave group velocities. A combined inversion of body wave receiver functions and Rayleigh wave phase velocities increases the uniqueness of the solution over separate inversions and also facilitates explicit parameterization of layer thickness in the model space. However, receiver functions are mostly sensitive to sharp velocity contrasts, and relatively insensitive to the average velocity and to smooth velocity gradients. Group velocity dispersion is sensitive to average shear velocity over a broad range of depths between two seismic stations. While extremely useful for determining the general velocity profile with depth, dispersion techniques are largely insensitive to velocity discontinuities. Combining these complimentary tools in a single inversion allows for more unique analyses of crustal and upper mantle structure and increases the uniqueness of the solution over separate inversions and also facilitates explicit parameterization of layer thickness in the model space. Our results indicate an average crustal thickness which differs from 40 km beneach MAKU station in NW of Iran up to 56 km beneath NASN station. A Moho depth of 40 (+/-2) km beneath the MAKU station located in NW of Iran, consistent with gravity studies (Dehghani and Makris, 1984) and previous results (Tatar, 2001), indicates on existing of a thin crust in this part of Iran.

KEYWORDS: Iran, joint inversion, receiver functions, surface waves.

1.INTRODUCTION

The active deformation in Iran is caused by Arabia-Eurasia convergence (Figure 1) (Tchalenko and Braud, 1974; Berberian, 1995). Nearly all of the shortening is accommodated within the political borders of Iran itself. The seismicity within Iran suggests that much of the deformation is concentrated in the Zagros, Alborz and Kopeh Dagh mountains, and in east Iran, surrounding Central Iran and the Lut desert, which are virtually aseismic and behave as relatively rigid blocks (Jackson & McKenzie 1984; Vernant et al., 2004). Results from a regional GPS network again suggest that the total convergence across Iran is 24 mm/yr (Vernant et al., 2004). The crustal structure of the Iran is rather poorly known.
Crustal structure is a basic and important subject in seismology, because it is often required as priori information for various geological and geophysical research. It can be estimated from various geophysical data acquired on the surface of the Earth. Surface wave velocity dispersion and P-wave receiver function inversion techniques provide complementary information regarding crustal and upper mantle structure. The receiver functions are sensitive to shear wave velocity contrasts in layered structures. The relative character of the receiver function constraints makes the inversion problem implicitly non-unique (Ammon et al., 1990) and this limitation can be greatly overcome by incorporating the constraints on the absolute velocities from the dispersion estimates (Juli`a et al., 2000) with joint inversion these two dataset. Acknowledge of crustal and upper mantle structure seismic velocity set us better perception of lithosphere. The Iranian plateau has been known as one of the seismically active areas of the world which frequently suffers destructive and catastrophic earthquakes that cause heavy loss of human life and widespread damage.

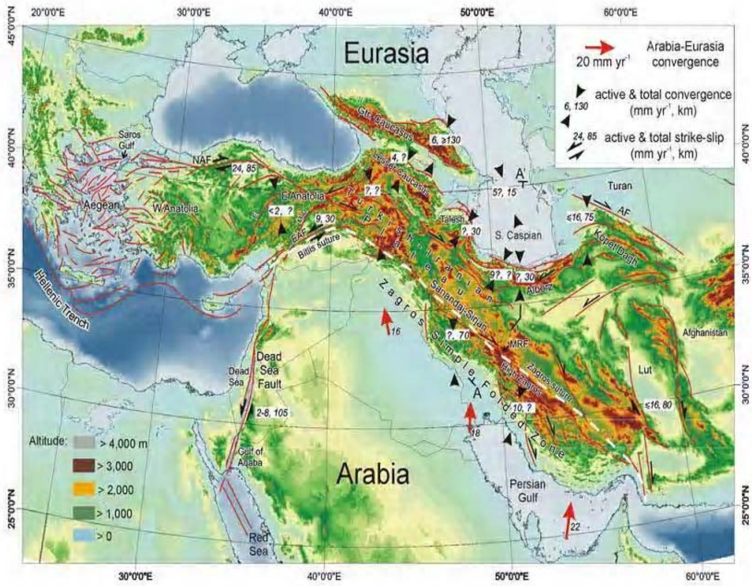

Figure 1: Topography map of Iran and deformation in Arabia-Eurasia convergence (Allen et al., 2004).

2.Previous Studies

The first of crustal thickness variations have been computed from bouguer anomaly modelling by Dehghani & Makris (1984) for whole Iran, and Snyder & Barazangi (1986) for Zagros. Mangino and Priestley (1998), for South Caspian Basin, showed that the crust in the southwestern and southeastern parts of the Caspian basin is 30-33 km thick. Hatzfeld et al. (2003) estimated a crustal thickness of 46 ± 2 km from receiver functions computed at a single station close to the town of Ghir in central Zagros. Javan and Roberts (2003) estimated a crustal thickness of 46 ± 2 km with receiver function method. Sodoudi et al. (2004) used the receiver function method to investigate crustal and upper mantle structure beneath Central Alborz in Northern Iran. They analyzed data from 290 teleseismic events by 12 short-period stations of the Tehran telemetric network. They estimated Average Moho depth about 44-46 km. Paul et al., (2006) computed Crustal receiver functions from the records of 45 temporary seismological stations installed on a 620-km-long profile across central Zagros. Their results displayed the Moho depth variations across the belt with good spatial resolution. From the coast of the Persian Gulf to 25 km southwest of the Main Zagros Thrust (MZT), the Moho is almost horizontal with slight depth variations around 45 km. Crustal thickness then increases abruptly to a maximum of ~70 km beneath the Sanandaj–Sirjan metamorphic zone, between 50 and 90 km northeast of the surface exposure of the MZT. Further northeast, the Moho depth decreases to ~42 km beneath the Urumieh-Dokhtar magmatic.

3.Data and methodology

In this study we investigated the velocity structure of the crust and upper mantle beneath the Iranian plateau from joint inversion of receiver functions and surface waves group velocity dispersion. We utilized 220 teleseismic events recorded by almost all the broadband seismic stations of Iranian National Seismic Network (INSN).The locations of INSN stations are shown on a simple geologic map of Iran in figure (2).Information about the group velocity dispersion comes from tomographic images between 15 and 100 s period produced by a study of regional fundamental mode Rayleigh waves propagating across Iran and surrounding regions (Rham et al., 2006).

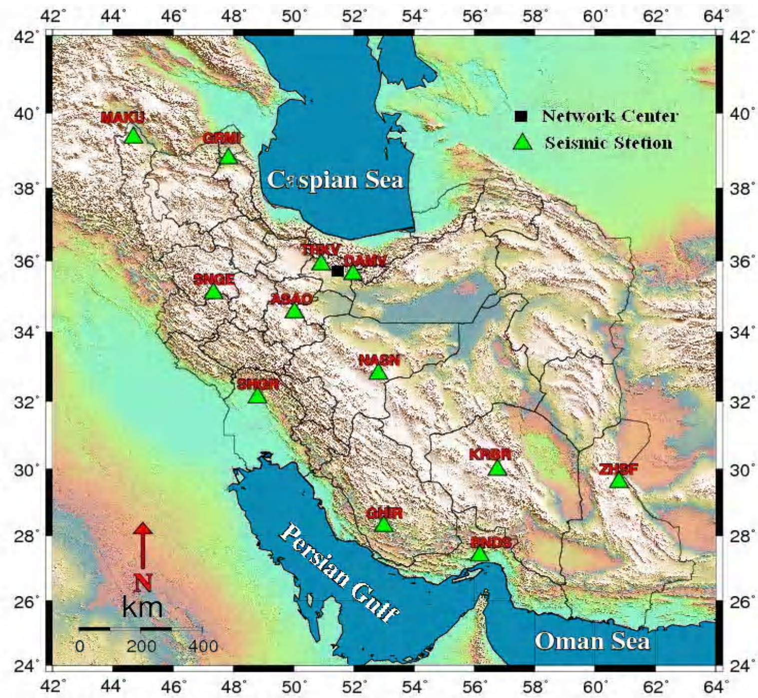

Figure 2: Iranian National Broadband Seismic Network.

4.Receiver function

Receiver functions represent the local earth response to the arrival of nearly vertical P-waves beneath a three-component broadband seismometer (e.g. Langston, 1979). Mathematically, the receiver function is the transfer function between the P-wave with all associated P multiples and reverberations, and the Ps phases with their multiples and reverberations [Ammon, 1991]. The waveform is a composite of P-to-S converted waves that reverberate in the structure beneath the seismometer (Figure 3). Modeling the amplitude and timing of those reverberating waves can supply valuable constraints on the underlying geology. In general, the receiver functions sample the structure over a range of ten's of kilometers from the station in the direction of wave approach (the specific sample width depends on the depth of the deepest contrast). Recent innovations in receiver function analysis include more detailed modeling of receiver function arrivals from sedimentary basin structures (e.g. Clitheroe et al., 2000), anisotropic structures (e.g. Levin and Park, 1997), and estimation of Poisson's ratio (e.g., Zandt and Ammon, 1995; Zhu and Kanamori, 2000).

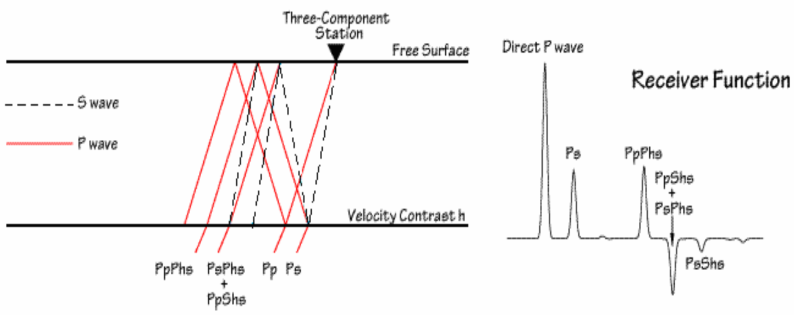

Figure 3. A graphical representation of receiver functions.

We computed receiver functions using the methodology of Ammon, (1991). More than one year of teleseismic waveforms recorded by 11 broad band stations of Iranian National Seismic Network (INSN) in the 25°–90° epicenter distance range were selected in order to obtain receiver function estimates for the INSN broad-band stations. The vertical component was deconvolved from the radial component to remove the signature of the source and instrument responses from the waveforms (Langston, 1979). The instrument response and the gains corrected before proceeding to the receiver function deconvolution. High-frequencies are excluded by using a Gaussian filter, we set the parameter a of the Gaussian filter to 1.00, which gives an effective high-frequency limit of about 0.5 Hz in the P-wave data. At most stations, the structure varies with azimuth and even in the simplest cases, the response can vary with distance from the station. We group the observations by azimuth, then distance. To increase the signal-to-noise ratio of the deconvolved traces, the individual receiver functions were aligned according to the P-wave arrival and point-to-point stacked waveforms from the same back-azimuth (< 20°) and distance range, Δ, (< 10°) for each station. The stacked receiver function was then allocated the average slowness and back-azimuth of every event included in the stack. The ±1 standard deviation (σ) bounds of each stack were used to monitor its quality.

5.Surface wave dispersion

Surface wave in the Earth are observed on seismograms of distant surface earthquake as long trains of dispersed waves with large amplitudes. Dispersion is easily detected, first arrivals corresponding to waves of longer period. Love and Rayleigh wave of short periods (8-12s) in continental trajectories are channeled in the upper crust and for periods between 60 and 300s they travel mainly through the mantle (Udias. 1999). Surface wave velocity dispersion primarily depends on S-velocity, with some dependence on P-velocity, and little dependence on density (Aki and Richards, 1980). The sensitivity of the fundamental mode to velocity at depth varies with period. Inverting Rayleigh wave group velocity dispersion simultaneously for structure should provide greater depth constraints on velocities, as each of them are sensitive to different depth ranges. They have been shown to improve inversions of receiver functions for crustal structure (Julia et al., 2000). They provide valuable information on the absolute seismic shear velocity, but are relatively insensitive to sharp (high wave-number) velocity changes. The dispersion velocity data set utilized in our joint inversion consists of group velocities of fundamental-mode Rayleigh waves. The group velocities were incorporated into our joint-inversion scheme from an independent surface wave tomography study by Rham et al., (2006). Group velocities from regional events recorded at permanent and broad-band stations were measured for fundamental-mode Rayleigh waves within the 15–100s period range.

6.JOINT INVERSION

Both receiver functions and surface wave dispersion velocities are primarily sensitive to S-wave velocity structure. However, receiver functions constrain velocity contrasts and relative vertical travel times beneath the recording station, while dispersion velocities constrain average absolute S-wave velocity values within frequency-dependent depth ranges (e.g. Takeuchi & Saito, 1972). The relative character of the receiver function constraints makes the inversion problem implicitly non-unique (Ammon et al,1990) and this limitation can be greatly overcome by incorporating the constraints on the absolute velocities from the dispersion estimates (Ozalaybey et al, 1997; Julia et al. 2000). The data sets are thus complementary. The simultaneous inversion of two datasets to find a single velocity model was carried out using the Computer Program in Seismology software package (Hermann and Ammon, 2003). Joint inversion of these complementary observations should reap the benefits of each, helping to reduce the imperfections of either. Inversion program relies on the awareness that the data, dispersion curve and receiver function, have different units, magnitudes, noise and number of observations. The inversion package requires that the real velocity structure be represented by a set of flat lying homogeneous isotropic velocity layers and during inversion the vertical extent of each layer remains fixed, whereas the layer velocity is free to change (within user defined damping limits). The starting model comprised velocity layers that were 1 km thick for the top 6 km of the model space, 2 km thick between 6-66 km, 4 km thick between 66-78 km. The starting velocity for every layer in the model was Vp=8.0 km/s.

Weighing parameter, p, is used to increase the importance of one dataset above the other. We tested several weighting parameter and selected which had good fitness for the two dataset at time. The value of receiver function standard error s_r was set at 0.0002 second. For the surface wave information the standard error was quantified based on the magnitude of residuals between observation and predicted phase velocities from the Stevens et al. (2002) model. Typical residuals are less than 1%, hence when the phase velocity is 5 km s^{-1}, the expected error is of the 0.005 km s^{-1}.

7.DISCUSSION

We computed receiver functions using the methodology of Ammon, (1991). More than one year of teleseismic waveforms recorded by 12 broad band stations of Iranian National Seismic Network (INSN) in the 25°–90° epicenter distance range (220 events) were selected in order to obtain receiver function estimates for the INSN broad-band stations. Rham et al., (2006) studied Rayleigh wave group velocity dispersion within Iran by performing a tomographic inversion. Group velocities from regional events recorded at permanent broad-band stations were measured for fundamental-mode Rayleigh waves within the 15–100s period range. We used these results for joint inversion receiver function and dispersion curve. The model parameterization utilized during our inversion procedure consisted of a large number of layers of uniform velocity and increasing thickness with depth.

We present the procedure in detail for one of station (GHIR) which is located southwestern of Iran. At GHIR acceptable receiver functions were calculated for 154 teleseismic events, from which 15 receiver function stacks were made. From the 15 receiver function stacks, a total of 14 were modeled for crustal structure beneath the station. We modeled the crustal velocity structure for all stacked receiver functions of different back azimuth, As can be inferred from Figure (4). The agreement between observed and predicted group velocities dispersion values, is very good. The match of the predicted receiver functions to the corresponding observed curve for almost all back azimuths is good for GHIR, BNDS and … stations .

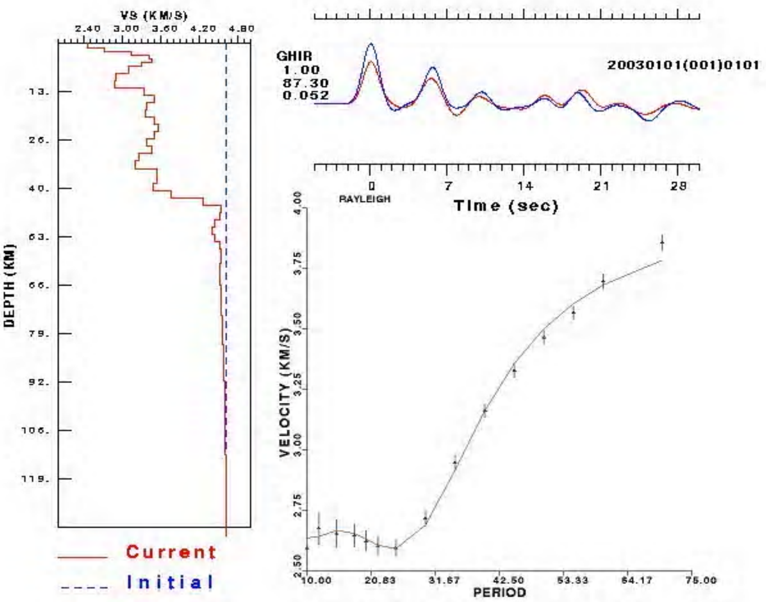

Figure 4: joint inversion results for GHIR station for 41-50 back-azimuth bin receiver function stack. The receiver function is at the upper right, the surface wave dispersion at the lower right, and the model at the left. The blue dashed line denotes the data and the red solid line the predictions for the model at the left.

For GHIR station, all the models contain a rapid velocity increase during the few first kilometers below the surface, a lower crust with a rather constant velocity, and a gradational crust-to-mantle transition. There is a thin (approximately about 2 km) of low-velocity material (Vs < 2.7 km s^{-1}) at the surface and a 10 km thick sediments layer (Vs = 3.1-3.25) indicative of above a 28 crustal thickness (Vs = 3.5-3.8). Moho is located at 42 ± 1 km depth.We estimated a Moho depth of 52 (+/-2) km beneath the DAMV station and 53 (+/-1) km beneath the THKV located in the central Albors, in north of Iran.
The inversion results for all the stations are summarized in Table (1). Our results indicated an average crustal thickness of 40-56 km beneath INSN stations.

Table 1. The inversion results for all the stations.

Station ZA	Lat.N	Long.E	Elevation (m)	Moho Depth (km)	Vs Velocity (km/s)
Ashtian-Arak (ASAO)	34.549	50.024	2217	46	4.2
Bandar-Abbas (BNDS)	27.387	56.174	1500	51	4.3
Damavand (DAMV)	35.63	51.97	2520	52	4.2
Ghir-Karzin (GHIR)	28.285	52.986	1200	43	4.2
Germi-Ardebil (GRMI)	38.79	47.83	1300	51	4.3
Kerman (KRBR)	29.98	56.7610	2576	47	4.3
Maku (MAKU)	39.35	44.68	1730	40	4.2
Naein (NASN)	32.799	52.808	2379	56	4.2
Sanandaj (SNGE)	35.092	47.346	1940	55	4.4
Tehran (THKV)	35.902	50.914	1795	53	4.3
Shooshtar (SHGR)	32.108	48.801	150	42	4.4
Zahedan (ZHSF)	29.614	60.795	1575	40	4.2

8.CONCLUSIONS

The combination of receiver functions and surface waves produces robust earth models when the data are high quality. We tried to study the Moho depth in different parts of Iran, beneath the Iranian National Seismic Network (INSN). The joint inversion of receiver function curves computed from more than one year of teleseismic events at range 25°-90°, with Rayleigh wave group velocity dispersion curve extracted close to each station from 3-D surface wave tomography Image, provide a good seismological constrain for determination of the crustal thickness in different parts of Iran within a reasonable error bars.
Our results indicate that the crustal thickness beneath Iranian plateau varies from 38-40 km in NW of Iran beneath MAKU station, up to 53-56 km beneath the NASN station located in central Iran. The average Moho depth in southern parts of the central Alborz (DAMV and THKV stations) is 52-53 km consistent with recent study in the region by Radjaee (personal communication).
The crustal thickness estimation for stations located in central (KRBR, ASAO) , southwestern (GHIR), and south (BNDS), (Figure 5) parts of Iran is in the range of 43-51 km. These results are in good agreement with Paul et al. (2006) and Hatzfeld et al. (2003), crustal structure studies. A Moho located at 38-40 km depth beneath the MAKU station confirms the results obtained by Mangino and Priestley (1998), for the south Caspian Basin, but 51 km thickness of crust beneath the GRMI station, the only station very close to the Caspian Basin is surprisingly high comparing with their results.

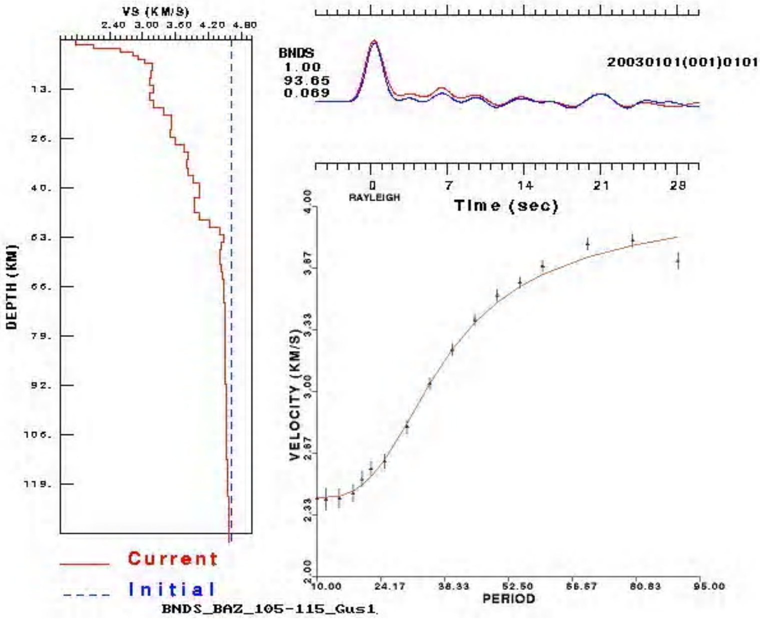

Figure 5: joint inversion results for BNDS station for 105-115 back-azimuth bin receiver function stack.

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