Ambient noise surface wave tomography of the Iranian Plateau

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SUMMARY

Ambient noise tomography is used to retrieve Rayleigh wave group and phase velocity variations in the period range of 8–40 s based on the vertical component of cross-correlation functions from permanent broad-band and mid-band seismometers across the Iranian Plateau. The iterative, non-linear inversion method of fast marching surface tomography (FMST) is employed to produce 2-D group and phase velocity maps. Shear wave velocities are also estimated using a linear least-square method.

Unlike most previous largescale tomographic results, our group, phase and shear wave velocity estimations, emphasize low velocity crustal structure (up to 50 km depth) beneath Zagros Fold and Thrust Belt (ZFTB) and Sanandaj-Sirjan metaporphic Zone (SSZ). The suture zone resulting from the subduction of the Arabian plate under the Central Iran is inferred along the boundary of SSZ and Urmieh-Dokhtar Magmatic Arc (UDMA). The velocity patterns show the main sedimentary basins, and reveal lateral velocity changes indicating the crustal thickening beneath ZFTB, SSZ and Lut Desert (LD), and the crustal thinning beneath Kavir Desert (KD) and UDMA are well inferred. A prominent low velocity is persistent in the whole crust beneath the central part of Alborz mountain range with high topography, and we suggest that it is likely due to elevated crustal temperatures within thin lithosphere.

Key words: Surface waves and free oscillations; Tomography; Crustal structure.

1 INTRODUCTION

Earlier tomographic studies across the Iranian Plateau including surface wave tomography (e.g. Shad & Shomali 2010), traveltime tomography (e.g. Alinaghi et al. 2007) and surface wave anisotropy (e.g. Kavian et al. 2009) come from the analysis of earthquake data at regional and teleseismic distances. Some shortcomings in traditional earthquake-based tomography such as the effect of irregular distribution of sources (or stations) and potential errors in seismic source parameters, affect the accuracy of the results. Furthermore, because observations at periods below ~20 s are weak or absent, the related tomographic results do not well correlate with upper crustal features.

Ambient noise tomography (ANT) provides a new tool for crustal imaging with horizontal resolution depending on the interstation spacing, by making use of group and phase velocity measurements between all available station pairs (as two deterministic points) and at periods much below 20 s. ANT has been performed around the world (e.g. Shapiro et al. 2005; Gerstoff et al. 2006; Yao et al. 2006; Li et al. 2009). The main idea behind seismic ANT is based on the fact that the empirical Green’s functions (EGFs) between two receivers can be extracted from the average of cross-correlation functions (CCFs) over a long time period as the negative time derivative of CCFs (Lobkis & Weaver 2001; Roux et al. 2005; Gouedard et al. 2008).

In this work, a permanent deployment of seismometers across the Iranian Plateau is used to extract interstation EGFs with Rayleigh waves as dominant arrivals, from stacks of ambient noise cross-correlations to investigate group and phase velocity variations. Surface wave velocity maps at different periods usually provide the first information in determining shear wave velocity models. Thus, this study is the first attempt to obtain fundamental mode Rayleigh wave group and phase velocity maps in the study area by using ambient noise data.

The Iranian Plateau, located in Alpine–Himalayan orogenic belt, features a complex tectonic setting, which results from the collision and convergence of the Arabian Plate towards Eurasia. This convergence is approximately 22 mm yr–1, and accommodated by deformation along the Zagros fold, Alborz and Kopeh-Dagh Mountains, Makran subduction and lateral displacement of the blocks in Central Iran (e.g. Tatar et al. 2002; Vernant et al. 2004; Masson et al. 2005). The major tectonic structures associated with various features in the study area, are shown in Fig. 1.

Both subduction and rifting models have been suggested for the tectonic setting of the collision zone, while the subduction model has been widely accepted. The Main Zagros Fault (MZF)
2 DATA AND METHOD

We process continuous vertical component seismograms recorded by 34 broad-band and five mid-band stations deployed in the Iranian Plateau (Fig. 1). We used mid-band station waveforms (Nanometrics TRIL-40s) recorded for 6 months between 2010 July 1 and 2010 December 30. The raw data for broad-band stations (Guralp CMG-3T-360s and CMG-3ESP-120s sensors) varies between 8 and 24 months, starting from 2009 and ending in 2010.

The data processing is carried out in two steps. First, we employ the method of cross-correlation procedure described by Bensen et al. (2007) to extract Rayleigh wave EGFs between stations. The second step is to measure group and phase velocity dispersions. Here we briefly describe two steps.

In the first step, all data are processed on a daily basis. The mean, trend, and instrument response are removed where the resulting transformed waveforms are velocity records. Next, the data is decimated to one sample per second, to retain only the effects of microseisms which include spectral contents of frequencies lower than 0.5 Hz. Finally, the one-bit normalization procedure is applied to suppress the influence of earthquake signals and instrument irregularities followed by spectral normalization, and then bandpass filtering between 0.01 and 0.5 Hz.

Stacks of daily vertical component cross-correlations are calculated for all station-pairs. The causal (arrivals at positive times) and acausal (arrivals at negative times) parts of cross-correlations are folded and further stacked followed by a negative time derivative. Our data selection for extracting only reliable measurements is based on signal-to-noise ratio (SNR) and inter-station distances. We determine SNR by dividing the power of maximum amplitude in the EGF signal by the power of the last 50 s of the record and retain all traces with SNR higher than 10 for a tomographic inversion. Also, dispersion measurements with inter-station spacing of less than three complete wavelengths will be removed through the second step.

In the second step, the fundamental mode Rayleigh wave group velocities are manually extracted through the multiple filter technique of Herrmann & Ammon (2002). Additionally, the method of Yao et al. (2006) was applied to measure phase velocity dispersions. In this method, a narrow bandpass filtering is applied on the EGF records to construct a time-period diagram. Then, the time-period records are aligned and transformed to velocity-period image through the use of far-field approximation. On the velocity-period image, the dispersion of phase velocities can be easily identified. It should be noted that the filtering process was carried out at central periods from 8 to 40 s with 1 s intervals. A detailed description of this method has been presented in Yao et al. (2006).

In order to estimate the ambient noise fundamental mode Rayleigh wave group and phase velocity across the study area, an iterative, non-linear inversion method of fast marching surface tomography (FMST; Rawlinson 2005) is applied, in which the differences between observed and calculated traveltimes are minimized through updating of propagation paths. This method is carried out in two steps. First, we apply the fast marching method (FMM), a grid-based finite difference solution process of eikonal equation to construct first-arriving wave fronts through the area by using traveltime data (Rawlinson & Sambridge 2005). This step is known as so-called forward problem. The most important advantage of this method over other ray tracing methods is that traveltimes are computed for all gridpoints of the medium. This is an effective tool, particularly when we have poor coverage of stations (or sources). In the second step, using the subspace method, the local linearization of the problem is performed relative to the current model to match the perturbation of the model parameters with the observed traveltimes. This is done by minimizing the objective function given by:

$$
\varphi(m) = \left(g(m) - d_{\text{obs}} \right)^T W^{-1}_g \left(g(m) - d_{\text{obs}} \right) + \varepsilon (m - m_0)^T W^{-1}_m (m - m_0) + \eta^T D^T D m
$$

where $g(m)$, $d_{\text{obs}}$, $W_i$, $m_0$, $W_m$, $D$, $\varepsilon$, and $\eta$ are predicted group (or phase) traveltimes, observed traveltime values, data weighting matrix, reference model, model covariance matrix, smoothness matrix, damping parameter, and smoothing parameter, respectively. The non-linearity is addressed by iterative application of the two steps, in which once the model perturbations is estimated, propagation paths are updated and retraced by the FMM scheme. The combination of the two steps provides stable and robust results, even in strongly heterogeneous media (Rawlinson et al. 2010).
3 RESULTS

The 8–50 s bandpass filtered cross-correlation record section with SNR higher than 10 are shown in Fig. 2. In the record section, the surface waves are dominant while a considerable noisy energy exists in the records. We think that it is probably due to the high level of seismic activity in the study area, and the corresponding effect could not be completely removed through the normalization process. Furthermore, we clearly observe the presence of reciprocity of surface waves in cross-correlations, although in different size and intensity. This reciprocity behavior reduces the level of bias in velocity measurements between stations (Yao & Van der Hilst 2009), since the ambient noise sources are two sided (not necessarily the same size) along a profile. But the spatial distribution of ambient noise sources and path effects may still influence tomographic results in terms of directionality (Pedersen et al. 2007), azimuthal and radial anisotropy (Yao & Van der Hilst 2009). On the other hand, a level of bias could be introduced even for an isotropic distribution of ambient noise sources, due to the unknown medium heterogeneity (Tsai 2009). However, although we do not provide a quantitative measure of such biases, we measure the normalized azimuthal dependence of energy flow across the region to estimate azimuthal energy contributing to CCFs construction. In this way, we compute rose diagram of SNR (normalized by the square root of the corresponding inter-station distances) at period bands centred on 8, 20, and 32 s, to estimate the energy flow which traverses the region. Fig. 3 clearly shows that at period of 8 s the most energy comes from south, while at longer periods (20 and 32 s) the ambient noise energy is relatively homogeneous, allowing possibly less biased velocity estimations resulting from the effect of directionality. It should be noted that the average normalized SNRs are 0.11, 0.15, and 0.17 for periods 8, 20, and 32 s, respectively.

An example of phase velocity measurements is presented in Fig. 4. An example of group velocity measurements, together with examples of phase and group velocity error calculations, is shown in Fig. 5. Our velocity (or traveltime) error estimates are calculated from cross-correlations from 21 stations for which we have time records of approximately 2 yr. We compare dispersion measurements (group and phase) corresponding to 3 months cross-correlations with 1-month overlap. The results indicate that at periods between 8 and 40 s, the average group and phase velocity uncertainties are approximately of the order of ~0.04 to ~0.1 and ~0.02 to ~0.07 km/s, respectively. We think that most part of velocity uncertainties results from attenuation at short periods, while at longer periods the lack of suitable inter-station distances and longer time duration stacking predominately affect our velocity measurements. As shown in Fig. 6(a), we have approximately 34 per cent selected measurements at period of 8 s, while for periods longer than 15 s and up to 40 s, there are approximately 50 per cent selected measurements with 5 per cent reduction at longer periods. According to the average SNRs from directionality analysis (Fig. 3) and to the average inter-station distances (Fig. 6b), it is likely that high attenuation in the Iranian uppermost crust (Cong & Mitchell 1998) affects higher frequencies. However, as shown in Fig. 6(c), the phase velocity errors are estimated to be less than group velocity errors through all consideration periods.

After selection of data as mentioned above, the Rayleigh wave dispersion measurements were used to invert for 2D velocity maps...
Figure 4. An example of phase velocity measurements from narrow bandpass filtering of EGF record for path indicated with solid line in Fig. 1. White solid line represents the normalized EGF filtered at the central period of 30 s. White dashed line indicates the smoothed dispersion curve.

Figure 5. (a) and (b) The resulting two-sided cross-correlation functions for paths indicated in Fig. 1. The shaded boxes display the velocity window between 2 and 4 km s$^{-1}$. Information about direction of most energetic signals and signal-to-noise ratio are also shown on the panels. (c) Group velocity dispersion measurements for paths indicated in Fig. 1. Grey lines show dispersion measurements for 3-month stacks. The final estimations over full time interval, resulting from stacking similar 3-month stacks, are shown by black lines. Solid line represents the dispersion curve corresponding to path which intersects different tectonic structures (path 1), while dashed line corresponds to path intersects a uniform tectonic feature (path 2). (d) Same as (c) but for phase velocity measurements.
at each period as a starting model, the inversion is performed on
of resolvable structures by real data (Trampert & Sneider 1996).
75 km grid for the inversions since this prevents possible leakage
tivation distribution. But certainly smaller class of anomalies may be
method is better suited than the ray-tracing methods for our sta-
poor coverage of rays. This implies that the wave front construction
have been recovered reasonably well in the whole region, allowing
scale length resolved at all periods and at the most places of the
8 s, for which we have minimum ray paths, while the resolution
teriods from 8 to 40 s, although some individual dispersion curves
range up to ~100 s. Since the resolution of surface wave tomogra-
depends primarily on path coverage, the azimuthal distribution of
paths, and the inversion method, a checkerboard resolution test
was conducted at each period to recognize how well various veloc-
velocity anomalies could be recovered in the study area. The tests were
performed using theoretical velocity models that are divided into
grid cells of the same sizes as the real model. The velocity field in
each cell is defined by continuously smooth B-splines oscillating
between 20 per cent above and below a background velocity of
2.8 km/s. Synthetic traveltimes are calculated by adding a Gaussian
noise with standard deviation of average traveltime uncertainties in
real data at each period. Different sets of damping and smoothing
factors were investigated for different cell sizes to reconstruct the
true velocity models. Fig. 7 shows the results of tests at period of
8 s, for which we have minimum ray paths, while the resolution
at the other periods is relatively better than what is shown here.
The results show that the 150 by 150 km grid cell is the minimum
scale length resolved at all periods and at the most places of the
study area, although large smearing occurred at the edge of the
network as expected. On the other hand, 200 by 200 km anomalies
have been recovered reasonably well in the whole region, allowing
valid interpretation of the results at this scale even for places with
poor coverage of rays. This implies that the wave front construction
method is better suited than the ray-tracing methods for our sta-
tion distribution. But certainly smaller class of anomalies may be
resolved where density of paths allows. Therefore, we use a 75 by
75 km grid for the inversions since this prevents possible leakage
of resolvable structures by real data (Trampert & Sneider 1996).

Finally, based on the average observed group (or phase) velocity
at each period as a starting model, the inversion is performed on
the real data. A tomographic map was produced for many different
combinations of regularization parameters at each period until the
misfit of the computed data falls below the acceptable threshold. Ideally, the elbow of the L-shaped curve of data misfit versus model
perturbation and roughness indicates the optimum trade-off region
to select an accurate damping and smoothing factors. The smoothing
is based on the second derivative of the velocity field. We found that
the smoothing factors are similar to the values in the synthetic tests
but the damping values slightly differ where the inversion satisfies
the data. Although a wide range of parameters was investigated, the
large scale pattern did not show major differences. For example, Fig. 8
displays model variance versus variance reductions through
different damping factors for group velocity map at period of 20 s.
Fig. 9 represents the results of group and phase velocity tomography
for three selected periods that are most sensitive to structures at
different depth ranges.

To further assess the implications of tomographic results, we in-
verted phase velocity maps to 1-D S-velocity depth profiles. In this
way, in each grid node with 75 km spacing, we extracted phase ve-
locities at each period from the corresponding phase velocity maps
to obtain dispersion curve in each node. The dispersion curve is
then inverted for 1-D model of shear wave velocity using the lin-
erized least-square method of Herrmann & Ammon (2002). The
inversion in all nodes is conducted with the same starting model.
We considered the estimated average shear wave velocity model
by Maggi & Priestley (2005), the standard AK135 model (Kennett
et al. 1995), and the average of both models as the initial models
(Fig. 10), but the Moho depth in different tectonically sub-regions
was considered to be the average crustal thickness calculated by the
receiver function studies (e.g. Hatzfeld et al. 2003; Paul et al. 2006;
Rham 2009; Sodoudi et al. 2009). To accommodate velocity gradi-
ents, the initial velocity models are sampled by 2 km thick layers,
and then the inversion was repeated 25 times. After completion of
the inversion for all nodes, the average 1-D profiles from all initial
models were collected into a 3-D volume, and the estimated shear
wave velocities are transformed to an image by a cubic interpo-
lation. Although a full 3-D inversion is required to well-constrained
absolute depth estimates, this approach is useful for determining
depth sensitivity for better interpreting the tomographic results. In
this paper, we present shear wave velocity structure along two depth
profiles crossing some tectonically important areas (Fig. 11).

4 DISCUSSION
The method of tomographic scheme that we employed here was
based on solving the eikonal equation to produce 2D group and
phase velocity maps. Theoretically, a problematic question is raised
when one is applied an eikonal solver for group traveltimes. Ac-
tually, in the case of surface waves, the eikonal equation can be
formulated for distributions of phase traveltimes not for group trav-
eltimes. But such a tomographic scheme works well if there is not
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Fig. 6. (a) Number of interstation paths as a function of period. (b)
Average interstation distance versus period for all qualified dispersion mea-
surements. (c) Average surface wave group (solid line) and phase (dashed
line) velocity uncertainties.

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Average interstation distance versus period for all qualified dispersion mea-
surements. (c) Average surface wave group (solid line) and phase (dashed
line) velocity uncertainties.
distance through all consideration periods is about 450–650 km (Fig. 6b), and in spite of low density of long paths, the individual distances do not exceed 1200 km. These considerations imply that multi-pathing and off-great circle effect may induce subtle changes to produce meaningful group velocity maps in our study. Thus, these effects are not as severe of a problem at these short distances, and various tomographic methods may not change the major patterns of group velocity maps from those presented here.

On the basis of group, phase, and shear wave velocity patterns, we discuss the results in terms of the major tectonic structures in the shallow part of the crust, mid-crust, and lower crust or uppermost mantle.
4.1 Group and phase velocity maps

The group and phase velocity distribution correlates well with the characteristic features of the structural map of the region. In general, at short periods (<15 s), the surface wave velocities are affected by the upper crustal structures and well indicate structures which are characterized by low velocities associated with major sedimentary basins. Mutually, velocities at longer periods (>15 s) tend to sample deep structures.

At period of 8 s (Fig. 9), we observed the most conspicuous structural features at Zagros Fold and Thrust Belt (ZFTB) and Sanandaj-Sirjan metamorphic Zone (SSZ), which were not previously revealed in other studies. A relatively continuous high velocity zone beneath the SSZ is associated with sedimentary and metamorphic Paleozoic-Cretaceous rocks, while the ZFTB is characterized by low velocities in the thick Mesozoic-Cenozoic sediment cover up to 12 km (Moutthereau 2011). This gradient boundary is approximately separated by MZF. The noticeable low velocities extending from South Caspian Basin toward Central Iran, likely reflect thick alluvial deposits and inter-mountain sediments in Alborz and salty deposits in Kavir Desert (KD) (Rieben 1966). The low velocity KD is well distinguishable from high velocity dispersed mountain ranges in Central Iran and northern Lut Desert (LD) to the east. Despite of low density of crossing paths, the southern LD is also characterized by average velocities, while the northern LD is dominated by high velocities likely due to intermediate to acidic igneous rocks (Rham 2009). Another prominent feature characterized by the lowest velocity anomalies are found approximately along the UDMA with volcanic and sedimentary rocks. These low velocity anomalies can be explained by the presence of magmatic rocks, and probably reflect lava flows through to pyroclastic deposits, in the upper crust (Alavi 1994; Rham 2009).

At period of 20 s, the density of ray paths allows more robust tomographic imaging, and the results depict high velocity zones that extend over most of the mountain ranges including ZFTB, SSZ and Alborz. Compared with 8 s group and phase velocity maps, the size of amplitude of high velocities noticeably increases in the 20 s maps oriented along the strike of ZFTB and SSZ. In contrast, these high velocities decline at periods higher than 20 s. At period of 20 s, the velocity maps exhibit high velocities associated with KD, Central Iran, and Alborz Mountain range. In Alborz, high velocities extend to the west and east parts, but the central part of the mountains is characterized by low velocities. This behavior is more prominent in 20 s phase velocity map.

At period of 32 s, surface wave velocities sample the lower crust and uppermost mantle, and primarily are useful indication of relative general crustal thickness through the sub-regions with different structural features. In our study, of particular significance, is an important high velocity contrast on UDMA, parallel to SSZ and ZFTB structures. The same pattern is observed for the KD where both group and phase velocities form a sharp high velocity region. At this period and in group velocity map, there is a roughly triangular high velocity feature covering the Alborz and South Caspian Basin. In both group and phase velocity maps, the eastern part of the Alborz seems to be characterized by higher velocities. This implies that the western part of this region is underlain by a thicker crust than the eastern part. According to evidences from other studies (e.g. Tatar et al. 2007; Sodoudi et al. 2009), the Moho depth is not flat across the Alborz mountain range. We also note that the low velocity feature beneath the Koppeh-Dagh mountain range at the north and LD at the south of Iranian Plateau is persistent at most periods higher than 20 s (not shown here), from which we infer the thick crustal structures.

4.2 Shear wave velocity model

Fig. 11 shows the shear wave velocity models along two profiles crossing some key regions of interest down to 50 km depth. As shown in Fig. 11 (profile AA’), the very low shear wave velocities are persistent from surface to mid crust beneath the central part of the Alborz mountain range. According to evidences from local body-wave tomography (Mostafanejad et al. 2011) and the presence of volcanic materials (Jackson et al. 2002), these low velocity anomalies are in part associated with the presence of a magma chamber in the upper crust. In this region the Moho also appears to be deeper than in the northeastern part. The joint inversion of receiver function and surface wave analysis by Rham (2009) found that the Moho depth varies from ~42 to ~50 km beneath the Alborz range. Also, it has been denoted from the gravity field study that the high topography Alborz range is not compensated by crustal root (Dehghani & Makris 1984), and the inferred crustal thickening in central part is due to magmatic addition at the base of the crust (Sodoudi et al. 2009). Although we could not provide constraints on the entire lithosphere structures, the low shear wave velocity in the whole crust is more likely due to elevated crustal temperatures concerning thin lithosphere inferred from receiver function study by Sodoudi et al. (2009) and young magmatism.

Assuming ~4.2 km/s as a proxy for crustal thickness and the lateral variations of vertical profiles (Fig. 11), the results imply a deep Moho discontinuity (~50 km) beneath ZFTB and SSZ, and a shallow Moho depth beneath UDMA and Central Iran. Earlier studies, for example, the receiver function and Bouger gravity anomaly modeling by Paul et al. (2006) and receiver function studies by Hatzfeld et al. (2003) found that the crust has an average thickness of ~56 and ~42 km in northern ZFTB-SSZ and UDMA, respectively. Furthermore, teleseismic P wave and surface wave measurements analysis by Kaviani et al. (2007, 2009) and Shad & Shomali (2010) revealed similar results in this region. The concept of crustal thickening beneath ZFTB and SSZ, and crustal thinning beneath UDMA is compatible with previous surface wave studies (e.g. Maggi & Priestley (2005); Shad & Shomali 2010), but our velocity estimations are quite different from their large-scale surface
wave tomographic results. This is mainly because very few of the waveforms included in previous studies are fit at periods shorter than 40 s. Their results have been probably affected by deeper structures with high and low velocity upper mantle of Arabian platform and Central Iran, respectively. It is worth noting that our results are more consistent with the surface wave study documented by Rham (2009), for which the period range is the same of this study.

The profile BB’ (Fig. 11) indicate that the suture zone, where the Arabian platform is subducted under the Central Iran, occurred approximately along the boundary of UDMA and SSZ. Also, some researchers have proposed that the subducted slab has broken-off beneath the suture zone (Molinaro et al. 2005; Shad & Shomali 2010), but this is not easily reconciled with our results due to the lack of sufficient observations at longer periods (>40 s).

Figure 9. Interstation ray path distribution, phase and group velocity maps at periods 8, 20 and 32 s provided by a 2-D non-linear inversion procedure. Information about the number of ray paths, period and reference velocity are also shown at the right upper part of panels. Some tectonic features are shown by dashed lines. The solid lines indicate the location of vertical profiles shown in Fig. 11.
5 CONCLUSIONS

This study is an investigation to gain insight into the whole crustal structure of the Iranian Plateau by using Rayleigh wave group and phase velocity measurements from cross-correlation of ambient noise data in the period range of 8–40 s. The phase velocities were inverted for shear wave velocity models. Our velocity patterns are in agreement with previously found features and, in addition, the results indicate new features corresponding to major sedimentary structures and crustal thickness through the study area. The vertical profile through the Alborz mountain range shows a thick crustal structure and a prominent low velocity zone.
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References


Rawlinson, N., 2005. FMST: Fast Marching Surface Tomography package, Research School of Earth Sciences, Australian National University, Canberra ACT 0200.


