An Application of Ambient Noise and Earthquake Tomography in the Rigan Area, Southeast of Iran

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INTRODUCTION

The Rigan area is located in east central Iran (see Fig. 1), at the southern part of the Lut block and the northern edge of the Jaz Murian depression (Mirzaei, 1998). The study area is surrounded by different active strike-slip faults, that is, the Bam fault from the west-northwest, the Kahurak fault from the northeast, and the Nosrat–Abad fault zone from the east (Fig. 1). The region experienced large destructive earthquakes including the 26 December 2003 (M_w 6.6) Bam earthquake.

Available information on the lateral heterogeneity and the velocity structure of the crust for the Rigan area is quite limited. Tatar et al. (2003) used a local travel-time tomography method that uses the aftershock sequence of the 26 December 2003 (M_w 6.6) Bam earthquake and obtained a 1D P-velocity model for the upper 12 km of the crust. The ambient seismic noise, which is a surface wave with random amplitude propagating in different directions, can provide information about the lateral heterogeneity. The ambient seismic noise contains valuable information regarding the propagation paths between a pair of stations (Shapiro and Compillo, 2004; Shapiro et al., 2005). The empirical Green’s functions between each pair of stations can be extracted using cross-correlation of long time-recorded ambient seismic noise (e.g., Shapiro and Compillo, 2004). Recently, this method has been widely applied around the world to obtain high-resolution tomographic maps, that is, the average interstation spacing is less than 32 km. To improve the spatial resolution, we also took advantage of the short-time recorded ambient seismic noise at a local scale, that is, the average interstation spacing is less than 32 km. To improve the spatial resolution, we also took advantage of the short-time recorded ambient seismic noise at a local scale, that is, the average interstation spacing is less than 32 km.

In December 2010, a moderate earthquake with a magnitude of M_w 6.3 occurred in the Rigan area. The causative fault of the 2010 earthquake is not clearly identifiable on the near-surface geomorphology (Walker et al., 2013). The epicenter of the event was associated with a hidden part of the Kahurak fault (Maleki et al., 2012). The earthquake was followed by a large number of aftershocks, that is, more than 300 aftershocks with Nuttli magnitudes (Nuttli, 1973) \( M_N \) larger than 4.0 (Fig. 1). Maleki et al. (2012) relocated the earthquakes using a nonlinear probabilistic relocation algorithm, which was developed by Lomax et al. (2000) based on the data from permanent local and regional seismic networks. Maleki et al. (2012) also studied the spatial–temporal variations of the seismicity pattern using nonlinear earthquake location (Lomax et al., 2000) and Omori’s law (Omori, 1894) and recognized two mainshocks over an interval of 37 days. The first mainshock (M_w 6.3) and the second mainshock (M_w 6.0) occurred on a fault with a parallel trend to the Kahurak fault and the Bam fault at depths of approximately 13 and 10 km, respectively (Maleki et al., 2012; Walker et al., 2013). The first mainshock was followed by 137 aftershocks (\( M_N \geq 4.0 \), solid black circles in Fig. 1), whereas 159 aftershocks (\( M_N \geq 4.0 \), solid white circles in Fig. 1) occurred after the second mainshock (Maleki et al., 2012).

In this paper, we studied the near-subsurface structure of the Rigan area using the ambient seismic-noise tomography (ANT) method. By developing a new stacking method, we extracted the empirical interstation Green’s functions from the short-time recorded ambient seismic noise at a local scale, that is, the average interstation spacing is less than 32 km. To improve the spatial resolution, we also took advantage of the aftershock signals of the first mainshock and calculated the dispersion curves for both the ambient seismic noise and the associated aftershocks.

DATASET AND DATA PROCESSING

We processed the continuous data recorded at six short-period temporary Iranian Seismological Center (IRSC) network stations from 23 December 2010 to 6 January 2011. Each station was equipped with an SS-1 Kinematics sensor and the data were recorded with 50 sps (Fig. 1). We also used the waveforms of 30 aftershocks recorded at the six short-period stations. We used hypocenter information of the aftershocks that were relocated by Maleki et al. (2012) who used a nonlinear relocation algorithm developed by Lomax et al. (2000). The aftershocks are depicted in Figure 1. The horizontal and vertical uncertainties associated with the events are on the order of 4 and 7 km, respectively.

To process these data, we followed the common low-frequency method that was explained by Bensen et al. (2007). After the continuous data were divided to 10-minute windows (e.g., Cho et al., 2007) with 30% of the windows overlapping in
we calculated the root mean square (rms) values of the CCFs to-noise ratios (SNR) larger than 10. SNR was defined as the ratio of the peak amplitude of the surface-wave signal in the velocity interval of $3 \text{ s}$ and $5 \text{ s}$, respectively. More than 300 CCFs were stacked to extract GFs with signal-to-noise ratios (SNR) larger than 10. The stacking process was defined by the change in the gradient of the rms curve of some interstation CCFs over the period ranges of $3 \text{ s}$ and $5 \text{ s}$, respectively. The number of CCFs used in the stacking process was defined by the change in the gradient of the rms curve. We called this stacking method the rms stacking.

Figure 1. Central box, study area; black triangles, the temporary seismic stations deployed by the Iranian Seismological Center (IRSC) network; black and white circles, aftershocks of the first ($M_L 6.3$) and the second ($M_L 6.0$) mainshocks, respectively; solid black lines, main faults. The focal mechanisms of the first and the second mainshocks were obtained by Global Centroid Moment Tensor (Global CMT) catalog (available at http://www.globalcmt.org/CMTsearch.html, last accessed September 2013).

time length, and when the mean and the trends were removed, the data were decimated to 10 s (e.g., Bensen et al., 2007). Each window was prefiltered with a fifth-order Butterworth filter with corners between 0.1 and 1.0 Hz (e.g., Pedersen et al., 2007). Then, the data were normalized in both time and frequency domains. Unwanted events, such as earthquake and aftershock signals, instrumental irregularities (spikes), and human activities, were removed using a nonlinear one-bit normalization in the time domain. Spectral whitening equalized the energies of the signals in the frequency range of interest (e.g., Bensen et al., 2007; Lin et al., 2007).

After processing the cross-correlation functions (CCFs), we calculated the root mean square (rms) values of the CCFs in the velocity interval of $0.5$–$4.0$ km/s. Figure 2a,c shows the rms curve of some interstation CCFs over the period ranges of $3 \text{ s}$ and $5 \text{ s}$, respectively. The number of CCFs used in the stacking process was defined by the change in the gradient of the rms curve. We called this stacking method the rms stacking. More than 300 CCFs were stacked to extract GFs with signal-to-noise ratios (SNR) larger than 10. SNR was defined as the ratio of the peak amplitude of the surface-wave signal in the time window of interest to the rms of the noise window (Bensen et al., 2007; Pedersen et al., 2007). Our studies showed that the SNR enhancement that resulted from 1500 CCFs was 10% lower than the SNR that resulted from 300 CCFs at the expense of a high computation time. Then, the empirical Green’s functions were retrieved using nearly two days (180,000 s that was equivalent to 14.2% of the total time) of nonconsecutive noise time windows over the period range of $1 \text{ s}$–$10 \text{ s}$.

The interstation Green’s functions with respect to interstation distances over the period ranges of $3 \text{ s}$–$5 \text{ s}$ and $5 \text{ s}$–$9 \text{ s}$, are shown in Figure 2b and d, respectively. The interstation Green’s functions were approximately symmetric in time as depicted in Figure 2. Therefore, an average of positive and negative lags of the Green’s function was then calculated to simplify the data analysis and to enhance the SNR. Figure 3 shows an example comparison between the empirical interstation Green’s functions and an aftershock that occurred on 1 Jan. 2011 $M_L 3.1$ over the period of $1 \text{ s}$ to $10 \text{ s}$. The station-pair separation and the epicentral distance are on the order of 28 km.

We applied a multifilter technique to obtain the group velocities of the Green’s function and the aftershock signals (Dziewonski et al., 1969; Herrmann, 1973). It should be noted that because of the Heisenberg uncertainty principle (Mallat, 2009), an inherent uncertainty exists in determining the time and frequency resolutions using the multifilter technique (e.g., Mallat, 2009). The optimal resolution in the time and the frequency domains can be achieved using a suitable Gaussian’s operator coefficient, which considers the interstation and the epicentral distances (Shapiro and Singh, 1999). A value of 3.0 was chosen for the Gaussian’s operator coefficient for the interstation and epicentral distances less than 40 km, and a value of 6.25 was chosen for larger distances. The selected group-velocity dispersion curves, which were calculated from the ambient noise (solid black lines) and from the aftershock sequences (gray lines), are shown in Figure 4a. The group-velocity dispersion curves were calculated under the interstation distance–wavelength condition ($\Delta \geq 3\lambda$), and only the empirical GFs and the aftershocks with interstation and epicentral distances larger than 15 km were used in the processing, because at epicentral distances of less than 15 km, the Rayleigh wave is not well developed (Tibuleac et al., 2011).

**TOMOGRAPHY**

We applied a tomographic method to estimate the 2D group-velocity maps of the fundamental mode Rayleigh waves. We also used the iterative nonlinear inversion package of the fast-marching surface-wave tomography (FMST), which was developed by Rawlinson (2005), and Rawlinson and Sambridge (2005), to minimize the difference between the observed and the calculated travel times. This method is based on two main steps, including the forward calculation and the inversion steps. The Fast-Marching Method (FMM) solves the forward problem of the travel-time calculation. With the assumption of local linearity, the inversion step was used in the subspace inversion scheme. The FMM is based on the eikonal equation, which is formulated for the phase time of surface waves rather than the group time. However, it is valid to use an eikonal solver to describe the dissipation of the group energy if multipathing is not included, in which case
the interfering waves cause the group energy to follow notably different paths. In other words, comparable results can be obtained when the phase and group velocities have notably similar geographic pattern (Arroucau et al., 2010; Saygin and Kennett, 2010; Young et al., 2011; Saygin and Kennett, 2012). Young et al. (2011) obtained similar group- and phase-velocity maps using FMM in southeastern Australia. Thus, the nonlinear relationship between the travel time and the group velocity could be explained by repeated applications of FMM and subspace inversions (Rawlinson, 2005; Rawlinson and Sambridge, 2005). However, the nonlinearity was not significant for the group velocity; thus, the results from the first iteration were considered the optimal solutions.

We used checkerboard synthetics to evaluate the resolution of the model. The resolutions of the checkerboard tests depended mainly on the path coverage at specific periods and on the azimuthal distribution of paths (e.g., Yang et al., 2011). The input model for the checkerboard test is shown in Figure 5a. In this study, the checkerboard cell size was 11

#### Figure 2.
(a) and (c) show the rms values of the cross-correlation functions with respect to the number of the windows in period ranges of 3–5 and 5–9 s, respectively. (b) and (d) show all empirical interstation Green’s functions in period ranges of 3–5 and 5–9 s, respectively.

#### Figure 3.
An example of comparison of empirical interstation Green’s functions for station pair DSHT–HSRG with an aftershock that occurred on 01 January 2011 at $M_L$ 3.1, over the period ranges of 1 to 7 s. The station-pair separation and the epicentral distance are on the order of 28 km.
by 11 km ($0.1^\circ \times 0.1^\circ$) and the inversion cell size was fixed to 5.5 by 5.5 km ($0.05^\circ \times 0.05^\circ$). The optimal cell size used in the inversion was obtained after exhaustive synthetic tests. This optimal cell size served to recover robust and stable tomographic maps (e.g., Tramper and Sneider, 1996). The background group velocity was chosen to be 1.5 km/s, with a perturbing velocity of 0.2 km/s and a maximum error of 5% noise signal. The results of the checkerboard tests at 3 and 7 s periods are presented in Figure 5b,c. The resolution was fairly sufficient over most of the study area, except at the west-northwest and the northeast parts of the mode, in which strong smearing was noticeable.

Finally, we performed surface-wave travel-time tomography on a 5.5 by 5.5 km ($0.05^\circ \times 0.05^\circ$) grid size to produce

\[ \text{Figure 4.} \] (a) Selected dispersion curves of fundamental-mode Rayleigh waves calculated from the solid black lines, ambient noise and the gray lines, aftershock sequences. (b) Sensitivity kernel of fundamental-mode Rayleigh-wave group velocity at period lengths of 1, 3, and 7 s.

\[ \text{Figure 5.} \] (a) Input checkerboard test model with velocity perturbation of about $1.5 \pm 0.2$ km/s. (b) and (c) the recovery of checkerboard test model at periods of 3 and 7 s, respectively. The cell size of the checkerboard was selected to be $0.1^\circ \times 0.1^\circ$ and the inversion cell size was fixed to $0.05^\circ \times 0.05^\circ$. 
2D Rayleigh-wave group-velocity maps for the study area. The background group velocity was obtained by averaging the group velocities from the dispersion curves. The Rayleigh-wave group-velocity maps at period lengths of 1, 3, and 7 s are shown in Figure 6a, b, and c, respectively. Generally speaking, each group velocity was proportional to a given depth range for which the average value was proportional to one-third of the dominant wavelength (Yang et al., 2007). However, the sensitivity kernel of the group velocity was calculated to provide information about the depth resolution as a function of the period lengths (Fig. 4b). Thus, the period lengths of 1, 3, and 7 s were associated with the depths of 0.4, 1.5, and 4.0 km, respectively, based on the information available in the sensitivity kernel of the group velocity. However, the variation in elevation was less than 900 m. Therefore, we ignored the topographic variation effects on the inversion. Regularization parameters, such as the damping and the spatial smoothness, control the amplitude of the velocity perturbations. We selected and maintained the damping value at 5.0, 3.5, and 2.6, and the spatial smoothness factor at 1.5 km for the period lengths of 1, 3, and 7 s, respectively. We adjusted different damping values and compared the data misfits. It should be noted that we set the inversion iteration value to 1, because of the linearity between the group velocity and the travel time for the fundamental-mode Rayleigh waves.

**DISCUSSION**

The lateral resolutions of the tomographic maps greatly depend on the epicentral/interstation distances and the path coverage. However, in this study for the period ranges of 1 to 7 s, distances longer than 65 and shorter than 15 km were not used in the processing. Consequently, as shown in Figure 6a, the Kahrurak fault was not resolved in the upper 0.4 km of the subsurface.

A strong low-velocity anomaly can be observed in Figure 6a in the shallower part of the study area, which is caused by desert alluviums near the mountain front. In the deeper parts, shown in Figure 6b,c, the low- and high-velocity anomalies are separated by the north-northeast trend, which is apparently parallel to the trend of the Kahrurak fault zone. The hidden part of the Kahrurak fault was clearly resolved in our 2D group-velocity tomography maps for the Rigan area. As depicted in Figure 6b, both mainshocks and the associated strongest aftershocks occurred in a relatively high-velocity zone, which was bounded by relatively low-velocity structures at a depth of approximately 1.5 km. The north-northeast trends of the transitions between relatively low- and high-velocity anomalies, which were resolved in the deeper part of the study area, for example, between 1.5 and 4.0 km, were consistent with the focal mechanisms of the two mainshocks (see Fig. 1).

**CONCLUSIONS**

In this study, we showed an application of ambient seismic-noise tomography over the period ranges of 1 to 7 s that reveals the hidden part of the Kahrurak fault in the Rigan area, in southeast Iran. We showed that an interstation Green’s function could be extracted by the cross correlation of the short time recorded (~15 days) seismic-noise data. For this purpose, we reduced the cross-correlation time-window length to 10 minutes (e.g., Cho et al., 2007), and introduced a new stacking method to enhance the SNR of the interstation empirical Green’s functions.
To improve the spatial resolution of the tomographic model, we used the aftershocks of the first mainshock. We demonstrated that the fundamental Rayleigh-wave group velocities that resulted from the interstation empirical Green’s functions were similar to the Rayleigh-wave group velocities that resulted from the aftershocks of the first mainshock. We used the resultant group velocities to obtain 2D group-velocity tomographic maps at specific periods. Our results confirmed the existence of a hidden fault, which was previously reported (i.e., Maleki et al., 2012; Walker et al., 2013). An interesting feature that appeared in this study was a low-velocity structure along the Kahurak fault in the depth range of 0.4 to 4.0 km. According to the results presented in this study, the hidden part of the Kahurak fault is clearly resolved in 2D group-velocity tomographic maps and can be associated with a low-velocity region at depths deeper than 1.5 km.

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