3D shallow structures in the Baogutu area, Karamay, determined by eikonal tomography of short-period ambient noise surface waves

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Abstract

Eikonal tomography based on ambient noise data is one of the most effective methods to reveal shallow earth structures. By tracking surface wave phase fronts, constructing travel time surfaces, computing the gradients of travel time surfaces to generate phase velocity maps, eikonal tomography avoids the ray tracing and matrix construction and inversion in the traditional surface wave tomography methods.

In this study, we collect continuous ambient noise data recorded by a dense seismic array in Karamay, Xinjiang to construct a 3D model of shallow structures using eikonal tomography. The seismic array consists of 35 stations with shortest interstation distance close to 1 km. 890 Empirical surface wave Green's functions (EGFs) between each station pair are retrieved by cross-correlating one or two months of continuous ambient noise data. From these EGFs, surface wave travel times in the frequency range of 1.8 to 4.0 Hz are measured by a Frequency-Time Analysis technique (FTAN). Then, eikonal tomography is adopted to construct Rayleigh wave phase velocity maps and estimate the phase velocity uncertainties. Finally, we invert the obtained phase velocity dispersion curves for 1D shear velocity profiles and then assemble these 1D profiles to construct a 3D shear velocity model. Major velocity features of our 3D model are correlated well with the known geological features. A shallow east-west velocity discontinuity is observed, which clearly reflects the lithological change between Baogutu formation (C_{1b}) and Xibeikulasi formation (C_{1x}) of lower Carboniferous system. Low shear velocities are observed beneath the location of porphyry copper deposit (V), possibly related to stockwork fracture and hydrothermal brecciation developed during the intrusion of deep magma in forming the deposit.

Keywords: ambient noise; eikonal tomography; phase velocity; Karamay.

1. Introduction

In the past decade, ambient noise tomography (ANT) has become a very popular method in revealing spatial variations of phase/group speeds of surface waves at different scales, from regional-global (e.g., [Shapiro et al., 2005](#); [Yao et al., 2006](#); [Yang et al., 2007](#); [Lin et al., 2007](#); [Saygin and Kennett, 2012](#); [Luo et al., 2012](#)) to local scales (e.g., [Picozzi et al., 2009](#); [de Ridder and Dellinger., 2011](#); [Gouédard et al., 2012](#); [Lin et al., 2013](#); [Mordret et al., 2013](#); [Hannemann et al., 2014](#)). ANT is based on the theory that cross-correlation of random wave fields recorded at two stations provides an estimation of Empirical Green's function (EGF) between these two locations (e.g., [Snieder, 2004](#); [Weaver, 2005](#); [Gouédard et al., 2008](#)). For receivers at the earth's surface, cross-correlations of ambient noise are dominated by fundamental-mode surface waves (e.g., [Shapiro and Campillo, 2004](#); [Kimman and Trampert, 2010](#)). In ANT, each station can be considered as a virtual source from which surface waves propagate to other stations. Surface wave travel time surface can be constructed in an area covered by a dense seismic array.

Based on surface wave travel time surface, an eikonal tomography method has been developed by [Lin et al. \(2009\)](#). The central concept of this method is to track surface wave phase fronts across a seismic array to construct the travel time surface, and then measure the local gradients of travel time surface to compute the local phase speeds. In a regional scale, [Lin et al. \(2009\)](#) obtain Rayleigh wave phase velocity maps using data from Earth Scope/USArray Transportable Array (TA) stations in western USA. On local scales, [de Ridder et al. \(2011\)](#), [Gouédard et al. \(2012\)](#), [Lin et al. \(2013\)](#) and [Mordret et al. \(2013\)](#) applied this technique to small-scale dense seismic arrays with thousands of stations to resolve the shallow structures in coastal areas. However, the huge numbers of seismometers and coastal research regions are harsh conditions for applying short-period ambient noise tomography. In our study, we succeed in applying ambient noise eikonal tomography to relatively sparse seismic array with only 35 seismometers deployed twice in Baogutu area, Karamay in inland

1 areas.

2
3 The sources of ambient noise at 1-50 s periods are considered to mainly come
4 from oceans(Stehly et al., 2006; Yang and Ritzwoller, 2008).However, noise at short
5 periods (<1 s) in a hinterland possibly originates from human activities (Draganov et
6 al., 2013). There are two main advantages of eikonal tomography when it is applied to
7 a seismic array deployed inland. First, eikonal tomography which accounts for bent
8 rays is more accurate than traditional straight-ray tomography in representing surface
9 wave propagation (Lin et al., 2009; Gouédard et al., 2012). Secondly, eikonal
10 tomography is less sensitive to the distribution of noise sources(Lin et al., 2013).

11
12 In this paper, we apply eikonal tomography to a small-scale array in Karamay,
13 Xinjiang to image the 3D shallow structure. The seismic array consists of 35 stations
14 with shortest interstation distance close to 1 km. 890 EGFs between two receivers are
15 retrieved by cross-correlating one or two months of continuous ambient noise data
16 recorded at all stations. From these EGFs, surface wave travel times in the frequency
17 range of 1.8 to 4.0 Hz are measured by a Frequency-Time Analysis technique (FTAN).
18 Then eikonal tomography is adopted to generate Rayleigh wave phase velocity maps
19 and estimate the phase velocity uncertainties. Finally, we invert the obtained local
20 phase velocity dispersion curves for 1D shear velocity profiles and then assemble all
21 these 1D profiles to construct a 3D shear velocity model.

22 2. Ambient noise cross-correlation

23 From 23 May to 25 July 2013, 35 3C-broadband seismometers with a sample
24 rate of 100 Hz were deployed in the Baogutu area, Karamay, as part of a regional
25 geological survey(Fig. 1). The target of the geological survey consists of two porphyry
26 copper deposits and a distinct north-south-trending fault zone which is geological
27 boundary between Baogutu formation (C_1b) and Xibeikulasi formation (C_{1x}) (e.g.,
28 Song et al., 2007; An et al., 2009; Shen et al., 2010; Zheng et al., 2014; Zong et al.,
29 2014). Continuous ambient noise data recorded by the dense seismic array provide

1 an opportunity to investigate 3D shallow structure of the Baogutu area.

2
3 The data processing procedures described by [Bensen et al. \(2007\)](#) is utilized to
4 obtain cross-correlations of ambient noise between all the possible pairs of receivers.
5 First, we cut continuous seismic data of vertical components to a series of one-hour
6 noise segment. Secondly, we perform spectrum whitening and temporal normalization
7 to normalize the data in both time and frequency domain, and then do
8 cross-correlation between all station pairs. Finally, cross-correlations on hourly
9 segments are all stacked together to get the final cross-correlations. 890
10 cross-correlations are obtained within the array. We define signal-to-noise ratio (SNR)
11 of the resulting cross-correlations as the ratio between the peak amplitude within the
12 predicted signal window of Rayleigh waves based on a velocity range from 1.5 to 4.0
13 km/s and the root mean squares (RMS) of the noise outside of the signal window.
14 [Fig.2](#) shows the record sections of ZZ component cross-correlations. Clear
15 fundamental-mode Rayleigh waves are observed in both negative and positive time
16 lags. However, it seems that one side of the cross-correlations has a higher frequency
17 content compared to the other side due to opposite directions of two main noise
18 sources. The azimuths of these two energies are consistent with the city of Karamay
19 in the northeast and the village of Daoban in the southwest ([Wang et al., 2014](#)). To
20 suppress the effect of an inhomogeneous source distribution, we construct the
21 symmetric components of cross-correlations by stacking the negative and positive
22 components.

23
24 To analyze the coherence of surface waves from cross-correlations, we construct
25 the normalized envelop for each EGF to approximate the wavefield emitted by a
26 virtual source of station ([Fig.3](#)). Each of the 1.5-3 Hz bandpass filtered
27 cross-correlations is normalized by the peak amplitude. We apply an envelope
28 function to fit the normalized cross-correlations in the time domain. The amplitudes of
29 normalized envelope functions are finally interpolated spatially. [Fig.3](#) reveals that clear
30 fundamental-mode Rayleigh waves emitted by the virtual source are observed in

almost all directions except in the NS direction. The discontinuous wavefront in the NS direction is related to inhomogeneous distribution of ambient noise sources. Nevertheless, reliable phase travel times can be obtained at nearly all azimuths, which is essential to map the accurate travel time surface. Moreover, [Lin et al. \(2013\)](#) has demonstrated that eikonal tomography can be applied even with an inhomogeneous distribution of noise sources as long as a well-defined wavefront can be identified for the wavefield emitted by virtual sources.

3. Eikonal tomography

Eikonal tomography is based on the eikonal equation ([Beydoun 1985](#)), which is derived from the Helmholtz equation ([Wielandt 1993](#))

$$\frac{1}{C_i(r)^2} = |\nabla\tau(r_i, r)|^2 - \frac{\nabla^2 A_i(r)}{A_i(r)\omega^2} \quad (1)$$

, where $\tau(r_i, r)$ denotes the travel time at position r , r_i is the location of the related effective source, and $\nabla^2 A_i(r)$ is the Laplacian of the elastic wave's amplitude at position r . Considering the high frequency or small spatial variation of the amplitude field compared with the gradient of the travel time surface, the second term on the right-hand of the Helmholtz equation can be neglected, then the Helmholtz equation, eq. (1), is greatly simplified as,

$$\frac{\hat{k}_i}{C_i(r)} \cong \nabla\tau(r_i, r) \quad (2)$$

, where $C_i(r)$ is the phase velocity at position r , and $\hat{k}_i$ is the unit wave number which contains the directionality of the phase velocity. The eikonal equation, eq. (2) is available when ignoring the second term on the right-hand in eq. (1).

We adopt eikonal tomography based on the eikonal equation to determine Rayleigh wave phase velocity maps at frequencies between the 1.8 and 4 Hz. For each frequency, we use frequency-time analysis (FTAN) ([Bensen et al., 2007](#); [Lin et al., 2008](#)) to measure travel time of positive, negative, symmetric-component of cross-correlation between each station pair with interstation distances longer than one wavelength, as [Luo et al. \(2015\)](#) has confirmed that accurate phase velocity can be extracted from station pairs with interstation distance as short as one wavelength

1 using ambient noise technique. We take the travel time measurements from the
2 component with the highest SNR among three components as our final travel time
3 measurements for tomography.

4 5 **3.1 Phase front tracking**

6 To construct a reliable phase travel time surface, a smooth surface fitting method
7 of Gridfit method proposed by [D'Errico et al. \(2006\)](#) is adopted to create a smooth
8 travel time surface on a 300×300 m grid ([Fig.4](#)). The theory of Gridfit method is based
9 on modified ridge estimator which is used to generate the surface where the bias
10 between fitting result and raw data is toward smoothness.

11 We follow the method described by [Mordret et al. \(2013\)](#) to divide the phase front
12 tracking into two steps. In the first step, we delete the areas where the resulting
13 difference between Gridfit method and spline interpolation method is larger than 5%.
14 To avoid the distortion of the travel time surface along the NS direction caused by the
15 inhomogeneous distribution of noise sources, we only retain regions where the
16 Laplacian of the travel time surface is less than $0.4 \text{ s}^2/\text{km}^2$ and measurements from
17 locations surrounded by at least 3 stations ([Fig.4a](#) and c). In the second step, we
18 re-interpolate the interstation travel times and remaining measurements ([Fig.4b](#) and
19 d).

20 21 **3.2 Phase velocity map**

22 A phase slowness map from each virtual source can be obtained through
23 computing the gradient of travel time surface for that source. Meanwhile, phase
24 velocity uncertainties can be estimated by the standard deviation of the mean
25 slowness ([Lin et al., 2009](#)). For each grid, we use an average of the phase slowness
26 computed from different virtual sources as the final slowness for that grid.

27
28 To examine the stability of eikonal tomography, we use different numbers of virtual
29 sources to compute the phase velocities ([Fig. 5a,b](#) and c) and associated
30 uncertainties ([Fig.5d, e](#) and f). As discussed by [Lin et al. \(2013\)](#), although the

large-scale patterns of phase velocity maps are generally similar, phase velocity variation is smoother and phase velocity uncertainty is reduced when more virtual sources are used. The average phase velocity uncertainty at 3 Hz is less than 40 m/s with 35 virtual sources. Some other studies with small-scale arrays also calculate the velocity uncertainty for eikonal tomography. For example, in the study of [Gouédard et al. \(2012\)](#), the average velocity uncertainty at 15 Hz is less than 2% with 1600 sources; in the study of [Lin et al. \(2013\)](#), the average velocity uncertainty at 1 Hz is less than 30 m/s when the number of virtual sources is 100.

In eikonal tomography, some parts of a study area may not be covered by stations for a particular virtual source, which leads to the different number of slowness measurements at different grids. [Fig.5\(g-i\)](#) shows the variation of number of slowness measurements in our study region at 3 Hz. For each grid, the uncertainties of slowness are mainly related to the number of measurements.

[Fig. 6\(a-c\)](#) shows phase velocity maps at 2 Hz, 3 Hz and 4 Hz, respectively. An obvious east-west velocity discontinuity appears in the 4 Hz phase velocity map ([Fig. 6c](#)), which is correlated well with the surface geological boundary between C_{1b} and C_{1x} . The phase velocity uncertainties at different frequencies are similar, having a mean uncertainty of 35.3 m/s at 2 Hz, 32 m/s at 3 Hz and 36.1 m/s at 4 Hz. We discover that the phase velocity uncertainties at lowest and highest ends of frequency are slightly higher due to fewer measurements.

3.3 Spatial resolution

We run checkerboard tests to estimate the resolution of phase velocity features constrained by eikonal tomography. Considering the phase velocities obtained from the data in Karamay, we construct checkerboard models with 5% velocity perturbation in a 2.5 km/s background velocity. The sizes of anomaly cells are 1, 1.5 and 2 km ([Fig.7a, d and g](#)). First, the Fast Marching method of [Sethian \(1999\)](#) and [Rawlinson et al. \(2006\)](#), a grid-based eikonal equation solver, is used to calculate the interstation

travel times for all virtual sources of our real data set. Next, these synthetic travel times are utilized to do the eikonal tomography as we do for the real dataset (Fig.7c, f and i).

For checkerboard tests with anomaly cells of 1.5×1.5 km and 2×2 km, the input pattern is well recovered at most of locations except at the western edge where station coverage is very sparse. For the checkerboard test with 1×1 km anomaly cells, the resolution at 3 Hz (Fig.7i) has degraded strongly in the peripheries and only the anomalies in the central region is recovered well.

4. Construction of 3D shear velocity model

From the Rayleigh wave phase velocity maps obtained by eikonal tomography, we extract local phase velocity dispersion curves in the range from 1.8 to 4 Hz at each grid. We then invert individual dispersion curves for 1D shear wave velocity profiles using an iterative linearized least-square inversion scheme of surf96 (Herrmann and Ammon, 2004). Finally, we assemble all the 1D shear velocity profiles to form a 3D model. In the inversion, we fix the V_p/V_s ratio and only perturb the shear velocity. As it is well known, a proper starting model is important in inversion. We utilize empirical formulas proposed by Xia et al. (1999) to determine the initial model. Xia et al. (1999) suggest that Rayleigh wave velocity with a wavelength (L) can be used in empirical formulas to estimate the initial values for the S-wave velocity of the layer at a depth of $0.63L$.

Fig.8 shows two examples of such inversion at two locations marked in Fig.1. The predicted Rayleigh wave phase velocity dispersion curves in the 1.8-4.0 Hz frequency range calculated for the inverted V_s profiles are plotted in Fig.8a, which fit the observed dispersions very well. The average 1D model for all locations is displayed in Fig.9a. Fig.9b shows the sensitivity kernels of Rayleigh waves calculated at various frequencies for the average 1D model. Rayleigh waves at different frequencies are sensitive to the Earth's structures at different depths. In general, high frequencies with

1 short wave lengths are sensitive to shallow structures and low frequencies with long
2 wave lengths are sensitive to relative deep structures. At the same frequency,
3 Rayleigh waves sampling C_{1x} are more sensitive to the shallower structure (< 400 m)
4 (Fig.9b) than those sampling C_{1b} (Fig.9c) due to low Rayleigh wave phase velocities
5 with shorter wave lengths at C_{1x} .

6
7 We use misfits of extracted local phase velocity dispersion curves from
8 tomography maps relative to the dispersion curves computed from 3D V_s model to
9 approximately evaluate the quality of local dispersion measurements. The
10 frequency-averaged misfits of dispersion measurements are plotted in Fig.10, which
11 vary from about 5 to 20 m/s at most locations with a mean misfit of 12.69 m/s and a
12 standard deviation of 3.8 m/s. The locations with misfits larger than 20 m/s are related
13 to the areas where phase velocity uncertainties are relatively high due to sparse path
14 coverage at the peripheries of our study region.

15
16 According to our tests, variations, like 10-15%, in the starting model do not affect
17 the inverted V_s model noticeably. Generally, the relatively narrow bandwidth (1.8-4.0
18 Hz) of the local phase velocity dispersions results in relatively strong non-unicity of the
19 velocity models. A testing V_s model (blue line in Fig. 11a) is taken from the inverted
20 one plotted in Fig. 8b, at location marked by triangle in Fig. 1. We repeatedly conduct
21 the same forward modeling for V_s models by generating uniformly distributed random
22 noise to the testing V_s model and then obtain 1000 velocity models (gray lines in Fig.
23 11a) with the synthetic dispersion curves (solid red lines in Fig. 11b) falling off the error
24 bars estimated by eikonal tomography. The average velocity and one standard
25 deviation perturbations of the 1000 performances are plotted as black and red lines in
26 Fig. 11a, respectively. In addition, the differences of each V_s and average velocity over
27 all the performances and all the depths are plotted as histogram in Fig. 11c. As
28 expected, the distributions of velocity differences follow a normal distribution with a
29 standard deviation of 42.9 m/s (Fig. 11c). Nevertheless, the standard deviation is
30 small for the main velocity anomalies.

We display the resulting V_s model of shallow structures in the Baogutu area at three different depths: 300, 400 and 600 m (Fig.11). At shallow depths (< 300 m; Fig.11a), we observe a distinct east-west velocity dichotomy with low velocities in the east and high velocities in the west. At greater depths (> 400 m; e.g., Fig.11b and c), the velocity anomalies are not correlated well with surface geological features. To better visualize the vertical variations of the V_s model, three cross-sections are displayed in Fig.12. A clear velocity discontinuity for all profiles in the top 400 m is also observed with low velocities in C_1b and high velocities in C_1x . A pronounced low velocity anomaly is imaged below the surface location of V porphyry copper deposit (e.g., Fig.12a, b and c) but a high velocity anomaly is observed below the surface location of VI porphyry copper deposit (e.g., Fig.12a).

5. Discussions

The shallow V_s structures constrained by eikonal tomography show clear correlation with surficial geological features. In the following paragraphs, we discuss in detail the predominant features of velocity model.

At shallow depths (< 300 m; Fig.12a), the obvious east-west shear velocity dichotomy with low velocities in C_1b and high velocities in C_1x is related to the geological trend of the Baogutu area and different rock characteristics between the two formations (C_1b and C_1x). C_1b is composed of volcanoclastic and volcanic rocks with fine grain and obvious stratification (Zong et al., 2014) and is the dominant host wall-rocks around the mineralized complex containing V porphyry copper deposit (Shen et al., 2010; Shen et al., 2012). On the contrary, C_1x is composed of volcanoclastic and volcanic rocks with coarse granule and weak stratification (Zong et al., 2014). The relative higher shear velocities in the shallow depth in C_1x can be interpreted in terms of the different lithological character and lack of mineralized complex. At greater depths (> 400 m; e.g., Fig.12b and c), the shear velocity patterns are not correlated well with the surface geological boundary, suggesting that C_1b and

1 C_1x are not vertically continuous strata.

2
3 Low shear velocities are observed beneath the V porphyry copper deposit
4 (Figs.12 and13), corresponding to mineralized complex caused by stockwork
5 fracturing and hydrothermal brecciation. Porphyry copper dikes are formed in the
6 stockwork fractures due to magma and fluid overpressures intrusion. (Rusk et al.,
7 2008; Hou and Cook, 2009). The depth of the low shear velocity at the area of V
8 porphyry copper deposit is shallow, less than ~500 m according to Fig.13a, b and c,
9 which is consistent with the well-logging data that V ore body extends to a depth less
10 than 500 to 600 m (Liu et al., 2009). The reason of the shallow extension depth of ore
11 body is that mineralized complex connected with porphyry copper has the
12 characteristic of shallow emplacement (Rui et al., 2006). However, high shear
13 velocities are observed beneath the VI porphyry copper deposit (Fig. 13a), which is
14 consistent with well-logging data that VI deposit is very low grade and lacks
15 mineralized complex.

16 Moreover, Love waves are mainly sensitive to horizontal polarized shear wave
17 velocity (V_{sh}) whereas Rayleigh waves are mainly sensitive to vertically polarized
18 shear wave velocity (V_{sv}). Analyzing the radial seismic anisotropy of V_{sv} and V_{sh} by
19 the combined inversion of Love and Rayleigh waves provides additional information
20 about deformation and flow within the research region such as geological formations
21 and mineralized complex (Luo et al., 2013), which is our current research topic.

22 6. Conclusions

23 In this study, we apply eikonal tomography in imaging high-resolution 3D shear
24 velocity structure of shallow depth in Baogutu area, Karamay, China. We process
25 continuous ambient noise records to obtain interstation Rayleigh wave travel times as
26 input for eikonal tomography between the 1.8 Hz and 4 Hz. The spatial resolution
27 assessed by checkerboard tests shows that the resolution of tomography at most
28 locations of our study is about 1.5 km and velocity features located in the central
29 survey area can be recovered at a 1 km scale, which is competent to image the

geological boundary and porphyry copper deposits in our study region.

Our 3D shallow structure is consistent with results from surficial geological surveys. A shallow east-west velocity dichotomy is observed with low velocities in C_1b and high velocities in C_1x across a surficial geological boundary. This observation is consistent with surficial geological features: C_1b is composed of volcanoclastic and volcanic rocks with fine grain and obvious stratification; but the compositions of C_1x are coarse granule and weak stratification. At the location of V porphyry copper deposits formed by intrusion of deep magma, shear velocities are low throughout the shallow depths, corresponding to stockwork fracturing and hydrothermal brecciation related to V porphyry copper deposit.

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1 Fig. 1. Location of our Karamay survey area and the distribution of broadband seismic
2 stations identified by black, red and blue squares. During the operation period, ten
3 stations denoted by red squares are removed to blue-square spots in the peripheries.
4 The observation time of each station is shown on the top as denoted by the different
5 colors. The geological boundary between C_1b and C_1x is identified by the black line
6 and the two porphyry copper deposits (V and VI) are marked by bold black lines. The
7 green circles mark the stations (C2 and P34) displayed in Fig.4. The blue triangle and
8 star denote the locations mentioned in Figs.8 and 9. In the top, the white star shows
9 the location of Karamay, with the topography shown as the background.]

10

11 Fig. 2. The record section of ZZ component cross-correlations filtered at 1-6 Hz
12 frequencies band with SNR larger than 5. The blue dashed line denotes the zero time
13 lag and the red dashed lines outline the prediction signal window bounded by the
14 move-out velocities of 1.5 km/s and 4 km/s.

15

16 Fig. 3. The normalized ambient noise symmetric-component cross-correlation
17 envelop plotted as snapshots at 0.6 (a), 0.9 (b) and 1.2(c) s lag times. These wave
18 fields are constructed for surface waves emitted by a virtual source (station C2)
19 denoted by the white star. The black concentric circles separated by 2 km intervals
20 represent the contours of source distances.

21

22 Fig. 4. 3 Hz Rayleigh wave phase travel time surfaces centered on two 'effective
23 sources' (station C2 and station P34) denoted by the white stars. (a) For station C2,
24 the first interpolated surface of travel times. Areas with excessively high curvature and
25 unconstrained areas are removed. (b) Second interpolation of the travel times. (c-d)
26 The first and final interpolated surface of travel times for station P34, respectively.

27

28 Fig. 5. (a-c) the 3 Hz Rayleigh wave phase velocity maps (the top panels), and (d-f)
29 their associated uncertainties (the middle panels) based on different numbers of virtual
30 sources. (g-i) The number of actual measurements at each grid. The number of virtual

sources in each plot is shown on the top and the locations are denoted by triangles of different colors (red, blue and black) in the inset. The imaging region is indicated by the green rectangle.

Fig. 6. Rayleigh wave phase velocity maps at 2 (a), 3 (b) and 4 (c) Hz, respectively.

Fig. 7. Checkerboard tests: (a, d, g): Input velocity maps with 2 km × 2 km, 1.5 km × 1.5 km and 1 km × 1 km checkerboard cells, the average velocity is 2.5 km/s. (b, e, h): Output velocity maps for 2 km × 2 km, 1.5 km × 1.5 km and 1 km × 1 km checkerboard input models, constructed by all synthetic travel times with interstation distance larger than 1 km. (c, f, i): Output velocity maps for 2 km × 2 km, 1.5 km × 1.5 km and 1 km × 1 km checkerboard models at 3 Hz.

Fig. 8. The comparisons of observed and calculated phase velocity dispersion curves obtained at two locations shown in Fig.1. The dispersion measurements and their uncertainties are derived by eikonal tomography. Blue and red dot with error bars represent observed Rayleigh wave phase velocity dispersion curves. Blue and red solid lines represent Rayleigh wave phase velocities calculated from V_{sv} shown in (b). (b) Inverted 1D model from dispersion measurements in (a). Blue and red dashed lines represent the related initial model obtained following by the formula of Xia et al. (1999).

Fig. 9. (a) Average 1D shear velocity model. (b-c) Depth Sensitivity kernels for 2, 3 and 4 Hz Rayleigh waves calculated based on the 1D models shown in Fig.8b (Fig. 9b is for red velocity model and Fig. 9c is for blue velocity model in Fig. 8b).

Fig. 10. The map of frequency-averaged misfits of observed local Rayleigh wave phase velocity dispersion curves relative to the predicted dispersion curves computed from the resulting 3D shear velocity model.

Fig. 11. A testing experiment for the reliability of inverted V_s model. (a) The blue line

represents the inverted V_s model plotted in Fig .8b (blue line). The gray lines represent 1000 V_s models, which are resulted from the addition of uniformly distributed random noise for each layer to the inverted V_s model. The black line and red lines identify the mean and one standard deviation perturbations of the 1000 V_s models distributions for each layer, respectively. (b) The red solid curves are the synthetic Rayleigh wave dispersion curves falling off the error bars, which are calculated from the 1000 V_s models (gray lines in Fig .11a). (c) The 1000 V_s models misfithistogram. The total counts are 6000 (1000 V_s models with 6 layers).

Fig. 12. (a-c) Maps of inverted 3D shear velocity model at 300, 400 and 600 m depths, respectively. The profiles of the three cross-sections shown in Fig.13 are marked in (b).

Fig. 13. Three cross sections of the 3D shear velocity model along profiles delineated in Fig.12b, plotted as perturbations relative to the average shear velocity model shown in Fig.9a. The contours outline 5% V_s velocity perturbation. The surface locations of the mines, V and VI, and geological boundary between C_{1b} and C_{1x} are denoted on the top of each panel.

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Figure 4

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