Semiannual velocity variations around the 2008 M_w 7.9 Wenchuan Earthquake fault zone revealed by ambient noise and ACROSS active source data

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Abstract We continuously monitor the long-term seismic velocity variation of one of the major ruptured faults of the devastating 2008 M_w 7.9 Wenchuan earthquake in China from July 2009 to January 2012, jointly using accurately controlled routinely operated signal system active source and seismic noise-based monitoring technique. Our measurements show that the temporal velocity change is not homogeneous and highly localized in the damaged fault zone and the adjacent areas. Velocity variations from the active and passive methods are quite consistent, which both are characterized by $\pm 0.2\%$ seasonal variation, with peak and trough at winter and summer, respectively. The periodic velocity variation within fault zone exhibits remarkably positive correlation with barometric pressure with stress sensitivity in the order of 10^{-6} Pa^{-1} , suggesting that the plausible mechanism might be the crack density variation of the shallow subsurface medium of the damaged fault zone in response to the cyclic barometric pressure loading.

Keywords Velocity variation · 2008 Wenchuan Earthquake · ACROSS active source · Ambient noise

1 Introduction

Monitoring temporal variations of the subsurface velocity field, especially those at the seismogenic depths, can provide valuable insight into structural changes in a fault zone associated with the subsurface stress field. Such structural changes could be related to coseismic damaging and the subsequent postseismic healing (Schaff and Beroza 2004; Brenguier et al. 2008a), as well as earthquake nucleation and other preseismic phenomena (Niu et al. 2008). Knowledge on how a fault system responds to various loading is also critical for assessing potential earthquake risk (Hillers et al. 2014). Thus, the long-term monitoring of seismic velocity around active fault zones is of great importance.

There are basically three well-documented methods for reliably monitoring subtle velocity changes occurring in the shallow crust, based mainly on the type of the sources, including active source (Li et al. 1998, 2006), repeating earthquakes (Schaff and Beroza 2004; Schaff and Kim 2012), and ambient noise (Wegler and Sens-Schönfelder 2007; Brenguier et al. 2008a). The applicability of the repeating-earthquake technique is restricted to specific regions ascribed to its sporadic nature, and it also suffers from the poor time resolution due to the episodic pattern of occurrence. Chemical explosive sources were widely used before, but became more and more restrained due to its unfriendly nature to the environment. More environment friendly active sources, such as the Vibroseis (Korneev et al. 2000), Accurately Controlled Routinely Operated Signal System (ACROSS) (Ikuta and Yamaoka 2004), and piezoelectric transducer (Yamamura et al. 2003; Silver et al. 2007; Niu et al. 2008), have been invented recently and used for the monitoring purpose. Moreover, some continuous natural sources, for instance, the persistent and

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localized microseismic sources near Kyushu Island of Japan (Zeng and Ni 2010; Wang et al. 2012) and in the Gulf of Guinea (Xia et al. 2013), may also act as repeatable sources for detecting temporal variations of crustal structure.

Recently, passive seismic monitoring based on ambient noise data (also known as Passive Interferometry Imaging, PII hereafter) has gained increasing attention and interest from the seismological community. It has been demonstrated experimentally and theoretically that the noise cross-correlation function (hereafter refer to as NCF) from ambient noise data recorded by a pair of seismic stations computed over a long period of time tends to converge toward the Green's function for wave propagating inside the medium between the two stations (see Wapenaar et al. 2010a, b; Snieder and Larose 2013 and references therein). Temporal variations of seismic velocity can be computed by comparing NCFs formed at the same station pair at different time periods. One tempting advantage of the PII method is that the continuous recording of ambient seismic noise allows for actively monitoring a specially interest region without using an active source. The PII method has been used to monitor changes occurring in the Earth's crust at various length scales, ranging from a few kilometers to hundreds of kilometers, including earthquake source regions (Cheng et al. 2010; Takagi et al. 2012; Obermann et al. 2014), fault zones (Wegler and Sens-Schönfelder 2007; Brenguier et al. 2008a), sedimentary basins (Meier et al. 2010), volcanoes (Sens-Schönfelder and Wegler 2006; Brenguier et al. 2008b; Martini et al. 2009; Obermann et al. 2013a), and oil/gas reservoirs (de Ridder and Biondi 2013). One known problem with PII is the trade-off between the time resolution and measurement precision caused by the frequency-dependent convergent rate of the NCFs (Hadziioannou et al. 2011). Moreover, the calculated NCFs can be distorted by the non-uniform distribution of noise sources (Froment et al. 2010), which could lead to spurious velocity perturbations (Zhan et al. 2013). Nevertheless, highly repeatable controlled sources, such as the ACROSS, are not affected by such uncertainties. Thus, there is a great need to compare velocity variations based on the PII technique with those derived from a controlled-source data.

The great Wenchuan earthquake (12th May 2008, M_w 7.9) occurred in the Longmenshan fold-thrust belt at the eastern margin of the Tibetan plateau in Sichuan, China. Field survey (Pan et al. 2014) and seismic reflection studies (e.g., Wu et al. 2014) have revealed that the devastating earthquake produced a series of complicated and high-angle dextral thrust fault structures, including the Wenchuan–Maoxian fault (F1), Beichuan–Yingxiu fault (F2) and Guanxian–Anxian fault (F3) (Fig. 1). Moreover, the southwestern and northeastern segments of

the Longmenshan fault belt are characterized by significant different rupture processes and aftershock distributions, as well as different Pn velocities (Li et al. 2012). The Wenchuan Earthquake Fault Scientific Drilling (hereafter referred to as WSFD) project was implemented to better constrain the damage zones of these thrust faults, as well as to better understand other geological and physical phenomenon associated with the rupture (Li et al. 2013; Wang et al. 2014).

Several studies have reported the coseismic velocity reduction in the source region and the adjacent area based on ambient noise data (Chen et al. 2010; Cheng et al. 2010; Froment et al. 2013; Liu et al. 2014; Obermann et al. 2014). In this study, we focus on one of the newly ruptured faults (Guanxian–Anxian fault) and monitor the long-term velocity changes occurring inside the fault zone and its surroundings. Shortly after the occurrence of the Wenchuan earthquake, we set up a monitor system near the WSFD drilling site. The system is composed of the ACROSS as the active source and a temporary linear seismic array that runs across the fault. By analyzing the recordings of the ACROSS source signal, Yang (2013) successfully detected a 0.3 % coseismic velocity drop associated with a nearby M_L 5.6 aftershock, indicating the high detectability of the monitoring system. In this work, we analyzed both the ACROSS active source data collected in the period of 12/2009–04/2011 and passive ambient noise data recorded between 07/2009 and 01/2012. Our goal is to reveal temporal velocity changes around the Guanxian–Anxian fault, as well as to compare the velocity variation measurements from both the PII method and man-controlled active source.

2 Active and Passive Data

The Guanxian–Anxian fault is roughly a pure thrust fault with a NE–SW orientation and southwestward dipping. We installed the ACROSS monitoring system on the SE side of the fault, i.e., the footwall side, ~5.5 km away from the surface exposure of the fault (Fig. 1). We use a similar type ACROSS source that was used by Ikuta and Yamaoka (2004) in Japan. It is composed of two 10-ton-weighted eccentric rotators, which are synchronized and rotate with a peak frequency up to 10 Hz in opposite directions, yielding a net vertical force that vibrates the earth.

We installed a seismic station (S0) near the ACROSS (<3 m), recording the source signal of the source. We installed another 8 seismic stations (S1–S8) along a line roughly perpendicular to the fault, with a station spacing of ~1 km, which intersects with the fault near No. 3 well of WSFD project. Each station is equipped with a short-period seismometer (CMG-40T) that has a flat velocity response

Fig. 1 Map showing the experiment site. **a** Locations of the ACROSS active source and stations are indicated by *solid star* and *solid black triangles*, respectively. The *black line* indicates the surface exposure of the Guanxian-Anxian fault. *Inset* shows the epicenter of the Wenchuan Earthquake (beach ball) and the major faults around the experiment site, as well as the national meteorological station (*solid square*) whose records of barometric pressure and precipitation are used in this study. The *open square box* on the Guanxian-Anxian fault indicates the experiment site. **b** Depth cross-section view of the Guanxian-Anxian fault zone along the linear seismic array. The *thick dark gray line* represents the fault, and the *gray shadow* marks the possible damaged zone along the fault associated with the Wenchuan earthquake. The *thick vertical black line* denotes the well trajectory of No. 3 Well of the Wenchuan Earthquake Fault Scientific Drilling (WFSD) project

in the frequency range from 0.5 Hz (2 s) to 100 Hz, together with a Reftek-130B 24-bit data acquisition system, which digitizes and records data continuously with a sample rate of 200 Hz.

In addition to the seismic recording, we also set up other sensors near the source site to record barometric pressure, temperature, and precipitation. These sensors, however, stopped operation after 04/2011, so we used the meteorological data from a nearby national station (solid square in Fig. 1), which is about 70 km away from the experimental site. The readers are referred to Yang (2013) for more detail about the ACROSS active source experiment. The ACROSS active source had been routinely operated from 12/2009 to 04/2011. After 04/2011, the linear seismic array continued recording passive data till 01/2012, when they were closed. The ~18 months operation of the ACROSS monitoring system provides an excellent dataset for measuring temporal changes of seismic velocity in the shallow crust around the fault. In addition, the continuous recording also allows using PII technique to make similar measurements, providing a unique opportunity to compare the two techniques and valid the results from the two datasets.

3 Methods

In this section, we describe the procedures to process the passive seismic noise data and active source data, as well as the techniques to use them to extract the NCFs from passive continuous data and the transfer functions (TFs) from the active source data.

3.1 NCFs from ambient noise

We used the vertical component of the recording to compute daily NCFs for all the 36 station pairs (2-combination of a set of 9). Since all the stations are equipped with the same sensor and digitizer, we did not remove the instrumental response to avoid numerical errors of deconvolution. In order to compute the daily NCFs, we first cut the continuous records into daily segments, and down-sampled the data from 200 to 5 Hz. We computed the cross-correlation functions between different station pairs using a running time method. We employed a 2048-point (409.6 s) long time window and a 256-point step in the calculation, resulting in 1679 normalized cross-correlation functions for

Fig. 2 **a** Reference NCFs between station S0 and the other stations (S1–S8). The arrow indicates the fault surface trace. Coherent direct phase is marked with the solid black line. **b** All the reference NCFs in this study, with the cross-fault station pairs, the NW side and SE-side station pairs showing at the top (G1), middle (G2), and bottom (G3), respectively

one pair of daily records. If a gap was present in one of the seismograms in the running time window, the cross-correction function of that time window was not computed. For each time window, any linear trends and means were removed from the seismograms. We also applied spectral whitening in order to enhance the ambient noise signals. To do so, we set amplitudes in the frequency band between 0.3 and 1 Hz to 1, and 0 elsewhere while keeping the phase spectra unchanged. We computed the cross-correlation functions in the frequency domain for each running time window, and then linearly stacked them to form the daily NCFs. Note that we focused on the relatively high frequency band to avoid the strong signals related to microseisms. Also since the ACROSS source signal lies in the frequency range of 2–10 Hz, thus is expected to have little effect on the NCFs here.

We manually examined all the NCFs and removed the ones with poor SNR. We then stacked all the NCFs from each station pair and use it as the reference NCF for that station pair. Figure 2 shows the reference NCFs between station S0 and all the other stations. For each station pair, we further arranged the daily NCFs sequentially and stacked them using a running bin with a size of 30 days and an incremental step of 6 days in order to enhance signal-to-noise ratio (SNR) and to detect robust changes in the NCFs.

Assuming that a homogenous change in velocity perturbation ($\delta v/v$) across the study is (~ 10 km length scale), then we expect that the change in travel time, $\delta\tau$, of the

coda wave with an lapse time τ accumulates linearly as it propagates through the scattered medium, i.e. $\delta\tau/\tau = -\delta v/v$. We adopted the stretch technique (Wegler and Sens-Schönfelder 2007) to estimate the velocity variation within the time window from 15 to 55 s (Fig. 3a). Figure 4 shows an example of the stretch technique. Both the causal and acausal parts of the NCFs are used simultaneously. The correlation coefficient between the reference NCF and the current stretched NCF within the chosen time window is also registered as coherence index.

3.2 TFs of the ACROSS sweeps

The sweep of the ACROSS rotators is accurately controlled by a computer. In our case, we started to rotate the two eccentric rotators at a frequency of 2 Hz, gradually accelerated them to the peak frequency of 10 Hz, and then slowed them down to a complete stop. This cycle took about half an hour. We ran the ACROSS 14 sweeps per day during the relative quiet nighttime (local time: 22 p.m. to 05 a.m.).

We took the half-hour long records of the vertical component of all stations, while the ACROSS was running and used them to compute the TFs at stations S1–S8. We followed the similar procedures of computing NCFs to calculate the TFs, except that we used the frequency band of the ACROSS signal (2–8 Hz). Note that the highest frequency (8 Hz) is slightly lower than the peak frequency

Fig. 3 **a** Reference NCFs between station S0 and the other stations, with the large-amplitude direct phase (–10 to 10 s) being masked in order for better showing the later arrivals (coda). The gray shadow areas indicate the time window (15–55 s) used for measurement of velocity variation in this study. **b** Reference TFs from ACROSS active source. The gray and black lines mark the coherent first and the second high-amplitude P-wave, respectively, both with an apparent velocity of ~ 5.0 km/s. Note that there is a 0.12 s time delay for the first P-wave of the cross-fault stations, which may be due to the low velocity within the fault zone. The gray shadow areas mark the time windows that were employed for the delay time measurement. We used a cross-correlation method to measure the delay time between the TFs recorded at different times and the reference TF at each station

Fig. 4 An example showing the reference NCF (black line) and a subsequent NCF (red line) computed at the station pair S1–S8. While the direct phase shows no noticeable delay between the two NCFs, the coda waves exhibit clear time shifts between the two traces in both the casual and acausal part of the NCFs. Note here the coda waveforms have been amplified for the purpose of illustration

of ACROSS active source (10 Hz). We considered the S0 records as the ACROSS source signals and TFs at stations S1–S8 were obtained by cross correlating their records with the S0 records. For each sweep, we obtained one set of TFs, resulting a total of 14 TFs at each station of the linear array from the 14 daily sweeps. We further stacked the 14 individual sweeps to obtain a daily TF for each station.

We followed the similar steps as that of NCFs to process the daily TFs to measure temporal subsurface velocity variations. As mentioned above, we only have 8 TFs,

which are the causal Green's function between the source and the linear array stations. For each station, we first computed the reference TFs by stacking all the TFs obtained for that particular station, which are shown in Fig. 3b. The TF of S1 is not shown here, because we found significant culture noise around station S1 within the studied frequency band, thus were not able to obtain robust TFs. We also binned the daily TFs every 30 days with a time resolution of 6 days.

We measure delay times of the main arrival with respect to the reference TFs using a cross-correlation based method, which includes a cosine fitting in order to obtain sub-sample precision (Wang et al. 2008). The gray shadow areas in Fig. 3b marked the time main arrival centered at on the 7 TFs. Based on waveform modeling, Yang (2013) identified the main high-amplitude arrival here as the direct P wave. We then computed the velocity perturbation based on the delay time. It should be noted that we used the direct P waves for velocity change estimation, instead of the coda waves as that were used for the ambient noise data analysis, since the low signal-to-noise ratio of the coda wave, as well as the short duration, prevents us from obtaining reliable velocity change measurements.

4 Results

The NCFs show a coherent phase with an apparent velocity of ~ 4.5 km/s (the black solid line in Fig. 2a), indicating it

Fig. 5 **a** Temporal velocity variation (*open circles*) from NCFs of station pair S1–S8 during the period from July 2009 to January 2012. The *gray shadow* represents the winter season (from October to April of the next year). The coherence indexes are also shown (*open blue triangles*), which are quite stable and well above 0.8. **b–d** Similar to **a**. Averaged temporal velocity variations from PII method over cross-fault (G1), NW-side (G2) and SE-side station pairs, respectively, with vertical bars denote the statistically standard error. **e** The same as **b**. Temporal velocity variation from ACROSS active source averaged over cross-fault station pairs during the period from December 2009 to April 2011, with *vertical bars* denote the statistically standard error. **f** The monthly averaged barometric pressure (*red line*) and the daily average precipitation (*blue line*). Also shown is the barometric pressure recorded at the ACROSS site (*cyan line*). Note the barometric pressure is also shown in **b** (*red lines*) for comparison

is likely a body wave. There are also coherent later arrivals up to at least 55 s, which are most likely scattered waves. The rich contents in body waves of the NCFs are probably due to the proximity of station pairs (less than 10 km, Fig. 1). Also, the NCFs of the cross-fault station pairs appear to have waveforms slightly more complicated than the others, which are likely distorted by the fault zone.

We divided NCFs into 3 groups, G1, G2, and G3. The first group (G1) includes all the cross-fault station pairs, while the other two groups contain the station pairs located at the NW (G2) and SE (G3) sides of the fault, respectively. As expected, most of the G1 NCFs show more complicated waveforms than those of the G2 and G3 groups (Fig. 2b). This feature is also observed from the TFs of the ACROSS source. The apparent velocities of both the first and the second P-wave are approximately 5.0 km/s (Fig. 3b), consistent with that of NCFs (Fig. 2a). Note that there is a 0.12 s time delay for the first P-wave of the cross-fault stations, which may be due to the low velocity within the fault zone. The TFs of stations in the vicinity of the fault (S4 and S5) have multiple later arrivals, which show similar characteristics to the fault zone trapped waves (Li et al. 2006).

As an example, we show the temporal velocity variation measured from NCFs of the station pair S1–S8 in Fig. 5a. There are some gaps in the measurements due to either lack of continuous recording at one of the stations or poor SNR of the NCFs during those periods. The observed velocity exhibits a strong semiannual variation with a high in winter (shaded areas in Fig. 5) and a low in summer (Fig. 5). We also show the coherence indexes between the 30-day bins and the reference NCFs at the top of the figure, which are generally above 0.9 (Fig. 5a), indicating that the velocity changes shown here are robust. In general, the measured indexes are higher and more stable in the dry winters than in the rainy summers.

The temporal velocity changes shown in Fig. 5a are measured from the coda waves in the causal window (15–55 s) and the acausal window (–55 to –15 s). We also made the measurements with a single time window (either in the causal or acausal sides) and other time windows (e.g., 25–65 s, 35–75 s), and obtained roughly the similar variations. We thus believe that our measurements are robust and reflect the true changes in the medium beneath the study area.

We applied the PII techniques to all the station pairs and obtained the temporal velocity changes for all the station pairs. We found that pairs within a group roughly show similar variations, although there are distinct differences between different groups. We further computed the average velocity changes of each group, which are shown, respectively, in Fig. 5b–d, respectively. We used method of Weaver et al. (2011) in estimating errors in the measurements. The semiannual variation shown in Fig. 5a can be

seen from all groups with different variation amplitude. The G1 group shows the largest variation, with amplitude of $\sim 0.2\%$, followed by the NW G2 group ($\sim 0.1\%$) and the SE G4 group ($\sim 0.06\%$).

The delay times estimated from the ACROSS TFs exhibit a variation very similar to the PII measurements. In Fig. 5e, we show the averaged velocity change measured from the cross-fault TFs at station S5, S6, S7, and S8, which is remarkably consistent with the NCFs results.

The barometric pressure measurement near the experimental site is also averaged over a 30-day period a 6-day increment (red line in Fig. 5f), along with the daily precipitation (blue line). The barometric pressure from our sensor at the ACROSS site is also shown (cyan line) here for comparison, which appears to be almost identical to that of the national meteorological station. We thus employed the national station records, which cover the entire 2.5-year period of our measurements, to represent barometric pressure changes in our experiment site. We also plot the barometric pressure records in Fig. 5b, and it appears that the estimated velocity changes follow closely with the barometric pressure. The excellent correlation between the two suggests that the observed velocity variation is highly likely caused by the barometric pressure change.

In order to obtain a quantitative relationship between two, we plot the observed velocity changes as a function of barometric pressure (Fig. 6). The stress sensitivities of the velocity perturbation estimated from the passive and active source data are both positive, and agree well with each other. Both are in the order of 10^{-6} Pa^{-1} (Fig. 6a), consistent with the previous field measurements with various types of sources (e.g., Yamamura et al. 2003; Silver et al. 2007; Wang et al. 2008). The velocity–stress sensitivity at the hanging-wall side (i.e., measured from the NW station pairs alone) is approximately one order of magnitude lower than the cross-fault measurements, in the order of 10^{-7} Pa^{-1} (black symbols in Fig. 6b). The sensitivity is more subtle in the footwall side measured from the SE station pairs, which is another order of magnitude smaller, $\sim 10^{-8} \text{ Pa}^{-1}$.

5 Discussion

It has been known that the Earth's ambient noise field fluctuates seasonally, and such fluctuations in noise source can result in small changes in the NCFs computed in different times at a fixed station pair. Applying the stretching technique to these NCFs can lead to artificial temporal variations that are irrelevant to the subsurface velocity field (Zhan et al. 2013). Thus, it requires special caution in interpreting the temporal variations measured from NCFs, and it is highly preferred to validate the measurements from such signals with another type of data. Thus, the

Fig. 6 Velocity changes are shown as a function of barometric pressure. **a** Velocity changes from passive (*open black circles*, Fig. 5b) and active (*open blue diamonds*, Fig. 5e) methods for the cross-fault station pairs, with stress sensitivities 1.58×10^{-6} and $1.21 \times 10^{-6} \text{ Pa}^{-1}$, respectively. **b** Velocity changes from passive method for the station pairs on the NW-side (*open black circles*, Fig. 5c) and SE-side (*open blue diamonds*, Fig. 5d) of the target fault, with a stress sensitivity of 3.09×10^{-7} and $0.37 \times 10^{-7} \text{ Pa}^{-1}$, respectively

active source measurements here are of great importance. The ACROSS active source used in this study possesses high repeatability, and has been proved to be capable of detecting subtle seismic velocity changes (Yang 2013). As stated above, the temporal changes in seismic velocity measured from the two datasets show a remarkable similarity, suggesting that the velocity variations observed here reflect the real structural changes within the rocks around the fault.

We could not determine the lateral distribution of the velocity variations which was done by Obermann et al. (2013a, 2014), largely due to the inadequate raypaths. Nevertheless, the observed changes in seismic velocity from the cross-fault station pairs (G1) are significantly larger than those estimated from the single-side pairs (G2 and G3), suggesting that the observed velocity changes are not distributed homogeneously within the subsurface medium and the apparent velocity variations in the vicinity of the fault are larger than the other areas. Large structural damages caused by major earthquakes are usually confined within the highly localized zones, known as the fault zones. Seismic velocity within the fault zone could be $\sim 20\%$ – 40% lower than that of the surrounding wall rocks (Li et al. 2006). The effective width, a fault damage zone, usually varies from a few hundreds of meters to a few kilometers at the surface and tapers to a few hundreds of meters at depth, depending on the rupture history of the fault (Li et al. 2006). While details on the depth varying fault zone are very difficult to be constrained, we speculate that the width of the damaged zone is at the order of kilometers due to the large coseismic slips observed in the area (up to 4 m, Pan et al. 2014). Additionally, the well log

from the WSFD project indicated that the rock crack density down to the depth of ~ 1.5 km is significantly higher than that of the surrounding rocks, which were interpreted as the fault zone. The high crack density implies that the effective seismic velocity inside the fault zone is more sensitive to external stress loading than that within the wall rocks (Yang et al. 2014). This explains why the cross-fault group shows a velocity–stress sensitivity approximately one order of magnitude higher than the other two groups.

As mentioned above, the primary phase shown on the TFs of the ACROSS source is the S wave, which penetrates to a depth of ~ 2 km based on the epicentral distance (~ 5 – 10 km). Wegler and Sens-Schönfelder (2007) argued that the coda wavefield of the NCFs consists mainly of body waves instead of surface waves. Obermann et al. (2013b) investigated this through numerical analysis of the depth sensitivity of coda waves, and found that the coda waves comprise both body and surface waves, and the proportion of body waves dominates the depth sensitivity and increases with lapse time. Obermann et al. (2013b) also pointed out that frequency plays a crucial role in the depth sensitivity analysis, with longer period coda waves more sensitive to velocity perturbation at greater depth (Obermann et al. 2014). Assuming diffusion approximation, the coda waves on the NCFs in the studied frequency range (0.3–1 Hz) are mostly sensitive to the velocity structure down to 1–2 km deep, which is comparable to the sampling depth of the direct S wave on the TFs. Given the fact that the NCF coda waves and TF S waves sample the very similar part of the fault zone, it is not surprising that the observed temporal changes in seismic velocity based on the

two types of data are nearly identical. Unfortunately, because of the frequency band limit of our short-period data, the temporal changes in the periods longer than 5 s cannot be determined reliably. Consequently, we are unable to detect temporal changes in deeper depth to resolve possible depth variation of velocity perturbation.

Seismic velocity of subsurface medium is controlled by the intrinsic physical properties of crustal rocks, such as stiffness, crack density, orientation, and inter-connectivity, as well the extrinsic physical status, such as the in situ stress level and fluid saturation. Any comprehensive changes among these elements can lead to dynamical velocity perturbations.

Generally speaking, a cracked medium can be interpreted as a two-phase medium, consisting of the solid rock matrix and cracks saturated with fluids. Therefore, crack properties (crack density, orientation, inter-connectivity) and the amount of fluid are the two critical parameters controlling the seismic velocity of cracked medium. All the other factors (in situ stress, barometric pressure, precipitation, etc) affect the subsurface velocity through changing these two parameters. For example, coseismic velocity drops associated with large crustal earthquakes are often attributed either to the creation of cracks induced by strong shaking near the surface or to the opening of cracks due to stress drops at depth (Brenguier et al. 2008a). Other observations on the temporal changes of the subsurface medium, such as changes induced by perturbations of barometric pressure (Silver et al. 2007; Wang et al. 2008; Niu et al. 2008), thermo-elastic effects (Meier et al. 2010), and ground water-level fluctuations (Wegler and Sens-Schönfelder 2007; Hillers et al. 2014), can be all explained either by a change in crack density or by an alteration in fluid saturation.

Numerous studies (e.g., Martini et al. 2009; Meier et al. 2010; Froment et al. 2013; Obermann et al. 2014) have reported seasonal variations of the subsurface velocity field. The observed variations were usually interpreted as a result of the dynamic interaction between the atmosphere and the solid earth. In this study, we found that the measured velocity changes correlate almost perfectly with the observed barometric pressure, suggesting that barometric pressure loading is likely responsible for the velocity changes. The estimated velocity–stress sensitivity is in the order of 10^{-6} Pa^{-1} , which agrees reasonably well with most of the previous subsurface measurements (e.g., Yamamura et al. 2003; Silver et al. 2007; Wang et al. 2008). The stress dependence of seismic velocity is generally attributed to the opening/closure of microcracks embedded in the media, as the consequence of perturbations of the in situ stress level. The damaged fault zone has significantly higher crack density than the surrounding wall rocks (Yang et al. 2012, 2014), and consequently is more

Fig. 7 Schematic cartoon showing the possible fault structure and the preferred crack model of the target fault zone. Isolated cracks or compliant pore structures are dispersed in the fault zone and the adjacent areas, with significant higher crack density in the damaged fault zone than the host rocks. The direct body wave and the scattered coda wave sample the very similar part of the fault zone. An increase of barometric pressures leads to closure of cracks and therefore a decrease in crack density, resulting in higher seismic velocity

sensitive to the external stress loading (Fig. 7). This might also explain why the observed velocity perturbations are localized highly within the fault zone and the adjacent regions.

In principle, the effective seismic velocity is expected to be proportional to the effective stress P_e , which is defined as the difference between the confining stress P_c and the pore pressure P_p ($P_e = P_c - P_p$). Since barometric pressure can affect the confining and pore pressures in different ways depending on the connectivity of the cracks (Wang et al. 2008), its effect on seismic velocity could be complicated. Silver et al. (2007) observed both negative and positive velocity–stress sensitivities with respect to barometric pressure changes, which they categorized as the near-field and far-field effect, respectively. In the near-field case, an increase in barometric pressure has more influence on the pore pressure, resulting in an increase of crack density through crack opening. While in the far field, the increase of barometric pressure weights more on the confining pressure side, leading to closure of cracks and therefore a decrease in crack density. In our case, the observed positive velocity–stress sensitivity indicates that the far-field effect of the barometric pressure is likely in the dominant side. The cracks or compliant pore structures embedded in the poroelastic medium are randomly dispersed in the fault zone and adjacent areas, with most of them isolated from the atmosphere. Under such circumstances, an increase of the barometric pressure will raise

the confining pressure via increasing loads on the solid framework. The increased barometric loads serve to close cracks, consequently resulting in a net reduction of crack density and increasing seismic velocity along the wave propagation path, which provides a reasonable explanation for our observations.

Seismic velocities can also be strongly influenced by fluid saturation (Grêt et al. 2006). Based on the theory of O'Connell and Budiansky (1974), for a medium with penny shaped cracks, seismic velocity changes non-linearly with both crack density and saturation fraction. For a fixed crack density, wet rocks tend to have higher seismic velocities than dry rocks do. Thus, changes in the fluid saturation rate can also lead to changes in seismic velocity. Wang et al. (2008) attributed part of velocity changes they observed to be related to an increase of fluid saturation resulting from precipitation. In fact, the amount of precipitation in the study area, which is mainly related to the regular monsoon rainfall, in the study area, also shows a roughly semiannual variation, thus it can be an alternative mechanism to explain the observed temporal variations in seismic velocity.

Here, we argue that the observed semiannual variation at our experiment site is unlikely caused by precipitation. First, previous observations on how precipitation affects seismic velocity appear to vary considerably from case to case. Obermann et al. (2014) observed a seasonal velocity variation of 0.2 % in southwest part of the Sichuan basin, which they attributed to the monsoon rainfall. Obermann et al. (2014), however, did not observe such a variation in the part of the Sichuan basin that covers our experiment site. Second, rise of the ground water level related to precipitation could increase the pore pressure via surface loading, as well as the water saturation. As shown by O'Connell and Budiansky (1974), we expect to see a higher seismic velocity in the summer due to the increase of fluid saturation fraction resulting from the monsoon rainfall, which is opposite to our observation. Although high pore pressure tends to cause a velocity diminution (Froment et al. 2013; Obermann et al. 2014), such mechanism seems to affect directly only very superficial parts of the crust (Sens-Schönfelder and Wegler 2006; Hillers et al. 2014), and has very little effect below the surface environment layer (~ 200 m) (e.g., Niu et al. 2003, 2008). By contrary, the observed velocity changes in our case occurred in the region down to the depth approximately 1–2 km. We thus argue that while the velocity fluctuation in the shallow subsurface medium ascribed to precipitation may also contribute partly to the velocity variation, the observed semiannual variation in seismic velocity is primarily caused by barometric pressure. Nevertheless, heavy rainfalls in the monsoon season seem to have very local effects (change in river load, flooding, landslides, etc.), which

could change the shallow structure around the fault zone and contribute to waveform decoherence, resulting in reduced cross-correlation coefficients during the rainy seasons as shown in Fig. 5a.

Although we detected significant semiannual velocity variation within the fault zone and the surrounding region, we did not observed continuous velocity increase associated with fault healing process as the case of the 2004 M_w 6.0 Parkfield earthquake (Brennguier et al. 2008a). The logarithmic velocity recovery in the vicinity of the source region of Parkfield earthquake was governed by the postseismic stress relaxation within deeper parts of the fault zone. On the contrary, possible coseismic velocity reduction in our study region should be associated with shallow fault zone damage due to the strong ground shaking, which seems to undergo fast-recovering process. The seismic velocity tends to quickly recover to a stable level, typically within 100–200 days after the main shock, which is still significant smaller than the pre-event level, probably due to the permanent damage in the shallow fault zone. Such long-term recovery processes in the shallow crust were observed in the source region of the 2008 M_w 7.9 Wenchuan earthquake (Liu et al. 2014), as well as the 2008 M_w 6.9 Iwate-Miyagi Nairiku earthquake (Hobiger et al. 2012) and the 2011 M_w 9.0 Tohoku-Oki earthquake (Wu and Peng 2012; Takagi and Okada 2012). Unfortunately, our temporary seismic array was deployed about 1 year after the occurrence of the main shock, so we did not observe the coseismic and postseismic velocity changes associated with the 2008 M_w 7.9 Wenchuan earthquake.

6 Conclusions

We conducted a small-scale active source experiment near the Guanxian–Anxian fault, a newly ruptured fault zone during the devastating 2008 M_w 7.9 Wenchuan Earthquake, to continuously monitor the temporal changes in the subsurface velocity field. We computed NCFs and TFs from passive recordings and active data, respectively, and estimated the temporal velocity changes by applying the PII technique to NCFs and subsample delay time analyzing technique to the TFs data. Measurements from the two types of data are remarkably similar; both show a semiannual variation with an amplitude of ~ 0.2 % in the subsurface velocity field. We found that the observed temporal changes occurred mostly within the fault zone, which is featured by low seismic velocity and high crack density according to the well log of the WFSD project. The velocity–stress sensitivity is estimated to be in the order of 10^{-6} Pa $^{-1}$ within the fault zone, 10^{-7} Pa $^{-1}$ in the hanging-wall size and 10^{-8} Pa $^{-1}$ in the footwall side, respectively, which are consistent with previous studies. Our results

demonstrate that it is plausible to reliably monitor long-term velocity changes in a scale of a few kilometers with ACROSS active source, and PII is another promising way in mapping subsurface structural changes within the fault zones and the adjacent regions.

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