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Key Points:

- The top ~20 m at the site has very low seismic velocities and very high V_p/V_s ratios
- Moderate PGA produces large coseismic S wave velocity reductions in the top ~20 m
- The coseismic changes are followed by rapid recoveries in less than ~240 s

Supporting Information:

- Supporting Information S1

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Imaging and Monitoring Temporal Changes of Shallow Seismic Velocities at the Garner Valley Near Anza, California, Following the M7.2 2010 El Mayor-Cucapah Earthquake

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Abstract Seismograms from ~700 local earthquakes recorded at various depths (0, 6, 15, 22, 50, and 150 m) by sensors of the Garner Valley Downhole Array in Southern California are used to analyze the shallow velocity structure and temporal changes of seismic velocities after the 2010 M7.2 El Mayor-Cucapah (EMC) earthquake. The direct P and S wave travel times between surface and borehole stations reveal very low shear wave velocities (178–259 m/s) and very high V_p/V_s ratios (6.2) in the top 22 m. Temporal changes of seismic velocities after the EMC earthquake are estimated using autocorrelations of data in moving time windows at two borehole stations (22 and 50 m) and seismic interferometry between multiple station pairs of the Garner Valley Downhole Array. The S wave velocity in the top 6 m drops abruptly by $14.3 \pm 3.3\%$, during the passage of surface waves from the EMC event with a peak ground acceleration of 39 Gal, and recovers in ~236 s. The average velocity reductions decrease with depth and are $10.9 \pm 3.1\%$, $8.5 \pm 2.1\%$, $6.3 \pm 2.1\%$, and $4.5 \pm 2.0\%$ in the top 15, 22, 50, and 150 m, respectively. Comparisons of seismic interferometry results between sensor pairs at 0–22 and 22–150 m indicate that statistically significant velocity changes are limited at the site to the top 22 m. Pore pressure data are in phase with the surface displacement and reach maxima when the highest velocity drop occurs, suggesting fluid effects contribute to the observed velocity reductions.

1. Introduction

The properties and dynamics of the heavily damaged material in the top few hundred meters of the crust are understood only in general terms, especially near fault zones, despite their great importance to observed seismic motion, crustal hydrology, subsurface reservoirs, reliability of underground facilities, and numerous other applications. The low normal stress at shallow depth renders the subsurface material highly susceptible to failure, masking details of deeper structures and processes. The shear wave velocities in the top few hundred meters can be 200–400 m/s or lower (e.g., Bonilla et al., 2002; Theodulidis et al., 1996; Zigone et al., 2019) and the attenuation coefficients can be as low as 1–20 (e.g., Aster & Shearer, 1991; Liu et al., 2015). Fault zone regions have enhanced rock damage in the top 2–4 km with 30–60% reduced seismic velocities compared to the surrounding rocks (e.g., Ben-Zion & Sammis, 2003, and references therein). Monitoring how subsurface materials respond to the occurrence of earthquakes and other loadings can provide important constraints for the rheology of shallow rocks, engineering seismology and various other topics.

Nonlinear site effects and temporal changes of seismic velocities of geomaterials have been documented in laboratory experiments (Ostrovsky & Johnson, 2001; TenCate et al., 2004), and in situ investigations based on the spectral ratio method (Karabulut & Bouchon, 2007; Nakata & Snieder, 2011; Sawazaki et al., 2006; Wu, Peng, & Assimaki, 2009; Wu, Peng, & Ben-Zion, 2009; Wu et al., 2010), cross correlations of the ambient noise (Brennguier et al., 2008; Hillers et al., 2015; Hobiger et al., 2012; Rivet et al., 2011), and repeating earthquakes (Peng & Ben-Zion, 2006; Poupinet et al., 1984; Rubinstein & Beroza, 2004; Schaff & Beroza, 2004). Laboratory experiments show reduction of seismic velocities under low confining pressure when dynamic strain is as small as 10^{-7} , while in situ observations imply minimum dynamic strains of 10^{-6} to 10^{-5} , corresponding to peak ground accelerations (PGAs) of ~30 Gal, are required to generate a velocity drop of less than 5% (Rubinstein, 2011; Wu et al., 2010). Sediment soils, with loose particles and low shear wave velocities, are more sensitive to strong motions than rocks and exhibit nonlinear and hysteretic behaviors.

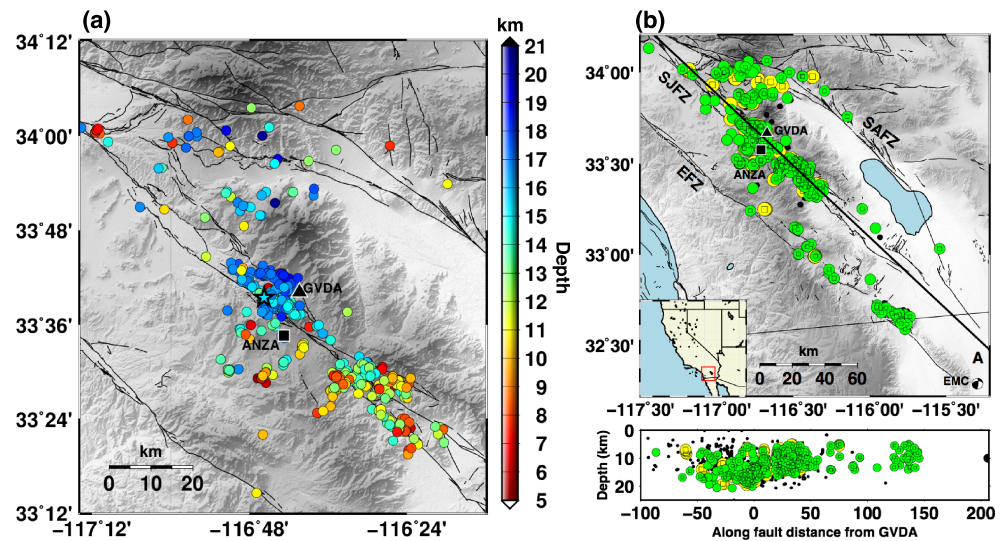


Figure 1. (a). All the events that have passed the selection criteria in the travel time analysis of direct arrivals. The star in the SW of GVDA shows the location of the example event (Figure 4). (b). All the events that are analyzed in the temporal change study. Black dots represent events that are eliminated by the selection criteria. Yellow and green symbols represent events that occurred before and after the EMC earthquake respectively. Squares and circles indicate events that are used for autocorrelation and seismic interferometry calculations, respectively.

Beresnev and Wen (1996) presented evidence for nonlinear soil behaviors (e.g., resonance frequency shift and deamplification) during large earthquakes with PGAs above 100 Gal. Deamplification by a factor of 2 was also observed at the Los Angeles basin related to the 1994 Northridge earthquake compared with its aftershocks (Field et al., 1997). Bonilla et al. (2005) found dilatant behavior of cohesionless soils that amplify surface motions from occasional high-frequency spikes in a low-frequency carrier in accelerograms. Lawrence et al. (2008) and Johnson et al. (2009) observed increasing reductions of shear wave velocity with applied force load, especially in the top few meters. TenCate et al. (2016) showed that crack orientation could have significant effects on nonlinear elasticity. Based on these observations, the strain/PGA threshold for nonlinear behavior is highly site dependent, and nonlinear site effects may be more pervasive and prominent than usually thought.

The shallow crust has often relatively soft sediment in the top tens of meters consisting of weathered and partially wet materials over relatively stiff bedrocks. Such shallow structures can be best analyzed with borehole data for velocities (Gibbs, 1989; Gibbs et al., 1990, 2001; Nakata & Snieder, 2012), attenuation (Aster & Shearer, 1991; Fletcher et al., 1990; Hauksson et al., 1987), dispersion (Liu et al., 2005), and anisotropy (Coutant, 1996; Liu et al., 2004). The Garner Valley Downhole Array (GVDA) was established in 1989 close to Lake Hemet, northeast of Anza, California (Figure 1, black triangle). The site has borehole seismic and pore pressure sensors at various depths (Figure 2) and provides valuable data sets to study shallow velocity structures and local site responses to various loadings.

In the present study, we analyze the velocity structure in the top 150 m of the GVDA site, and temporal changes of seismic velocities following the 2010 M7.2 El Mayor-Cucapah (EMC) earthquake with an epicenter 206 km away in Baja California. The data used in the analyses are described in section 2. In section 3, we first investigate the shallow velocity structure using travel times of direct *P* and *S* waves recorded by GVDA stations and then evaluate temporal changes of seismic velocities by autocorrelation and seismic interferometry. The imaging results indicate very low shear wave velocity (178–259 m/s) and very high V_p/V_s ratio (6.2) in the top 22 m. The analysis of velocity change indicates 14.3% *S* wave velocity reduction in the top 6 m after the EMC event and that the velocity reduction concentrates in the top 22 m. Assuming logarithmic recovery, the local velocity recovered approximately to the original state in ~236 s. Correlations between pore pressure data and velocity changes suggest that fluid interactions with shallow soft materials contributed to the observed temporal changes. The results are discussed and summarized in the final section 4.

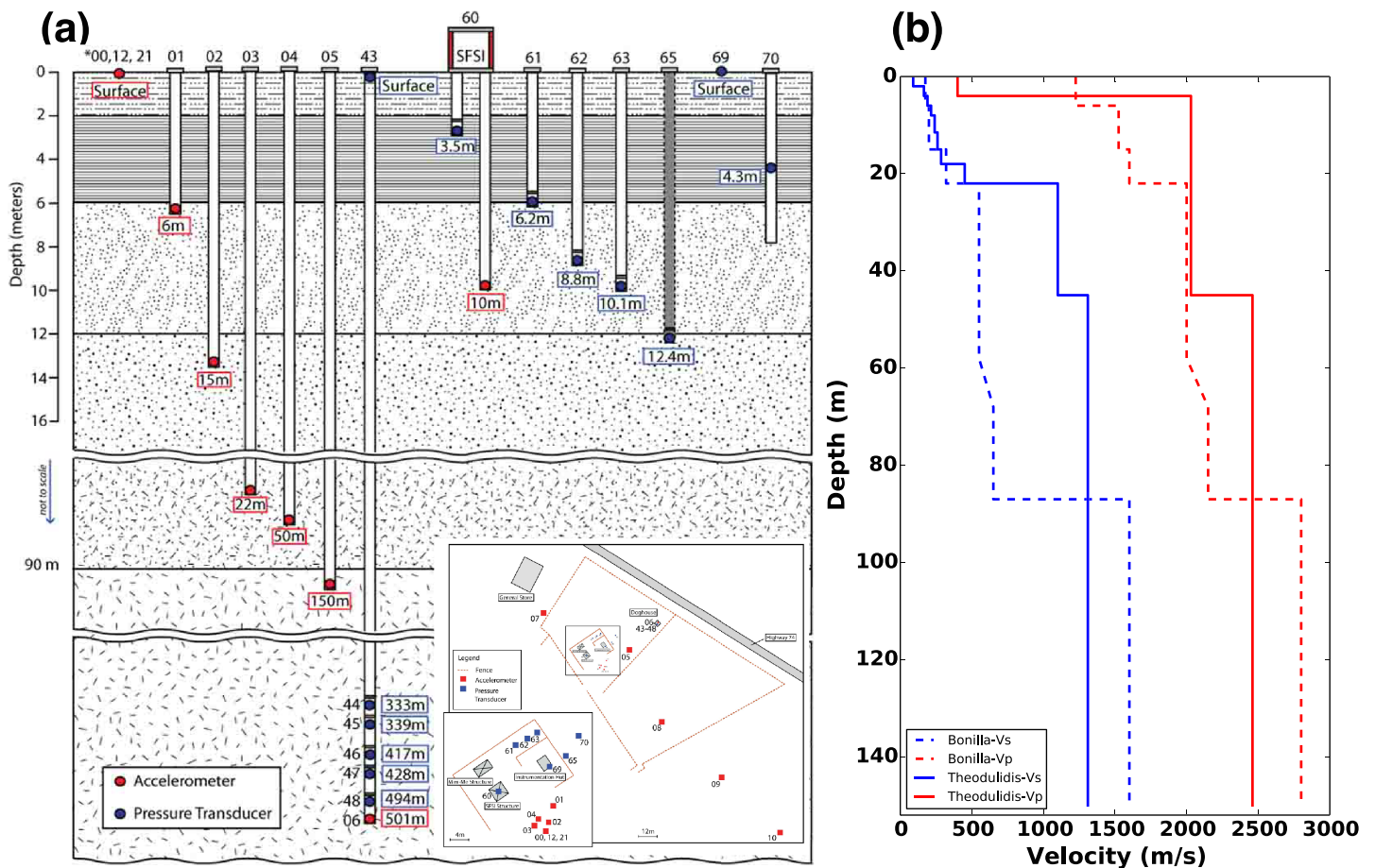


Figure 2. (a). The conceptual cross-section placement and surface layout (lower right corner inset) of the GVDA (available from <http://nees.ucsb.edu/facilities/GVDA>). Please note that the plot is not to scale vertically. Numbers on the top represent channel location codes and depths are labeled beside the sensors. Different symbols in each layer represent different materials from top to bottom: silty sand to sandy silt, sandy silt to clay, sand to sandy silt, sand to silty sand, weathered granite, and granite bedrock. Seismic Stations (red dots) 00–05 and pore Pressure Sensors (blue dots) 60–63 are used. (b). Velocity models of the GVDA site in the top 150 m from Bonilla et al. (2002) (dashed lines) and Theodulidis et al. (1996) (solid lines).

2. Data

The GVDA (Figure 1) was deployed in a narrow valley within the Peninsular Ranges Batholith in Southern California, about 7 km from the San Jacinto fault and 35 km from the San Andreas fault, and started recording in the summer of 1989. The station layout of the GVDA is shown in Figure 2a. The site is characterized by alluvium in the top 18 m with a gradual transition to decomposed granite from 18 to 25 m, decomposed granite between 25 and 87 m followed by a transition zone from 87 to 95 m, and granodiorite as bedrock deeper than 95 m (Bonilla et al., 2002; Steller, 1996). The attenuation coefficients of P and S waves in the top 150 m are in the ranges 15–50 and 10–30, respectively (Bonilla et al., 2002). Figure 2b shows velocity models at the site from earlier studies of Theodulidis et al. (1996) and Bonilla et al. (2002). The former constructed a velocity profile with results at different depths from previous studies, while the latter obtained velocity values by modeling accelerograms and using information from the velocity logs, core samples, and borehole suspension logging. Both models have a low shear wave velocity (<200 m/s) in the top few meters, and the model of Bonilla et al. (2002) also includes a strong reflection interface at 87 m that traps most of the reflected phases.

We analyze seismic data from local earthquakes in 2010 (Figure 1) recorded at the surface (00) and five borehole (01–05) sensors at 6-, 15-, 22-, 50-, and 150-m depths (Figure 2a). Pore pressure recordings in the liquefiable layers at depths of 3.5, 6.2, 8.8, and 10.1 m are also examined to investigate the influence of subsurface fluid on the local site response. The seismic sensors and pressure transducers have a sampling rate of 200 Hz.

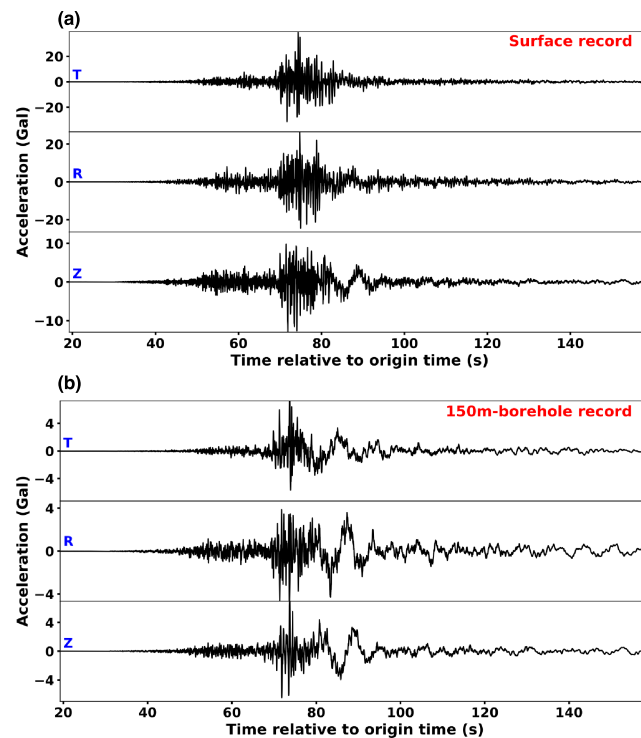


Figure 3. The transverse (*T*), radial (*R*), and vertical (*Z*) accelerograms (in Gal) from the EMC earthquake at the surface (a) and 150-m borehole (b) stations of the GVDA. Zero time corresponds to the occurrence time of the EMC event. Please note the different y scales.

We first remove the seismic instrument responses and rotate the east-west (HNE) and north-south (HNN) horizontal seismic recordings to the radial (HNR) and transverse (HNT) directions. Then we inspect the direct *P* and *S* wave travel times for additional results on the shallow velocity structure at the site. Multiple stations are analyzed further to evaluate the temporal changes of seismic velocities at various depth ranges following the EMC earthquake. The EMC earthquake recordings at the surface and 150-m borehole stations are presented in Figure 3, showing a PGA of about 39 Gal when the surface waves arrive.

3. Analysis and Results

3.1. Analysis of Direct *P* and *S* Waves

To estimate the seismic velocities in the top 150 m, we calculate the *P* and *S* wave travel times from stations at depth (6, 15, 22, 50, and 150 m) to the surface by short time window cross correlations. We first band pass the waveforms at 1–30 Hz and pick the direct *P* and *S* arrivals at the surface and 150-m borehole stations by automatic picking algorithms (Ross et al., 2016; Ross & Ben-Zion, 2014). The arrival times at other stations are estimated assuming a constant velocity in the top 150 m. This preprocessing provides accurate reference times for cross correlation, eliminating the possible bias caused by later arrivals. The waveforms are then interpolated from 200 to 1,000 samples per second to obtain subsample accuracy. The travel times are obtained by cross correlating the target waveforms (i.e., borehole traces) with reference records at the surface in a short time window (0.15 s for *P* and 0.3 s for *S*), with one third of the window before the estimated arrivals. During the process, we discard the record if (1) the event is shallower than 5 km; (2) the average signal-to-noise ratio (SNR) is smaller than 2 in the 1- to 30-Hz band. The SNR is defined as the spectral ratio between the signal window (*P/S* waves for the cross correlation) and the earlier noise window; (3) the cross correlation coefficient is smaller than 0.5; and (4) the travel time value has an absolute deviation from the median value outside the 80th percentile. Locations of about 450 events that pass the selection criteria are shown in Figure 1a.

Figure 4 illustrates results from an example event marked with a star in Figure 1a. The shifted surface record based on the delay time from cross correlation fits well with the waveforms at depths, implying the reliability

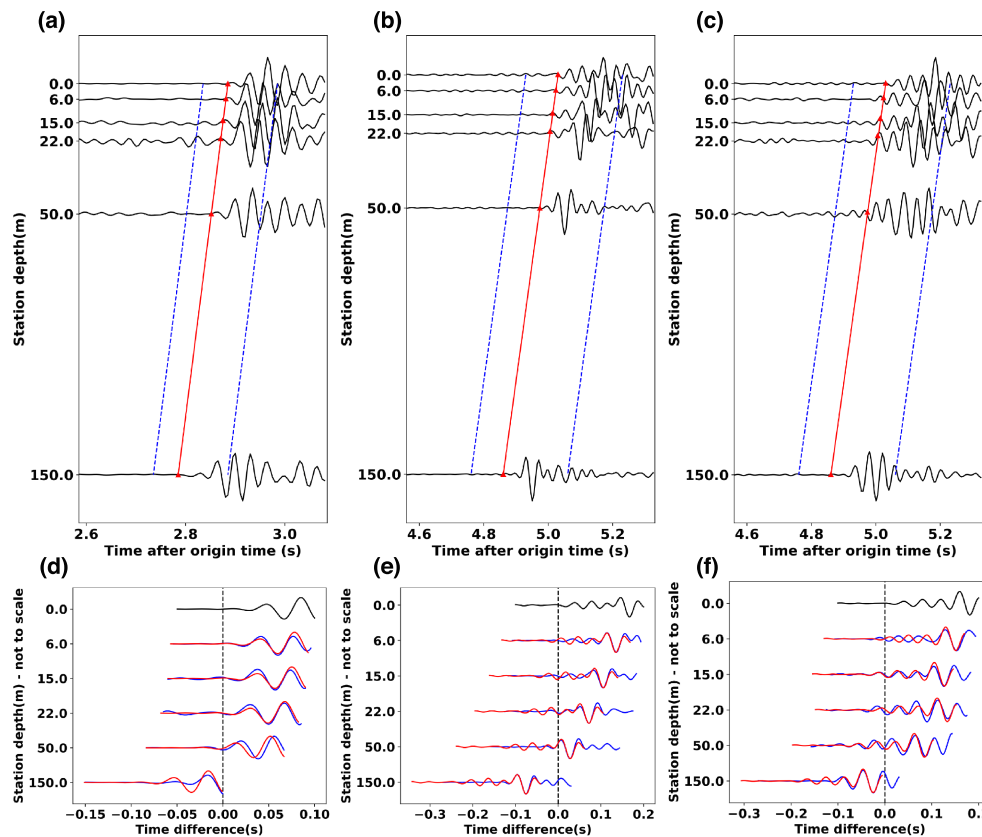


Figure 4. Example results from one event (shown as a star in Figure 1a). (a–c) The 1- to 30-Hz band-pass-filtered vertical (P wave), transverse (SH wave), and radial (SV wave) component waveforms from the six GVDA stations. Red triangles and solid lines represent the estimated direct arrivals, and blue dashed lines show the time windows used for cross correlation. (d–f) The interpolated waveforms in the cross correlation time windows. Surface records are in black, and borehole records are in blue. The red waveforms represent surface records shifted based on the time delay obtained from cross correlation.

of the obtained travel time. The travel time histograms at the five borehole stations are plotted in Figures 5a–5c. The distributions show small absolute deviations, except that the P wave travel time at the 6-m borehole (i.e., Station 01, blue symbols in Figure 5a) is less than 0.005 s, below the resolution of the data with a sampling rate of 200 Hz. Also the obtained average P wave velocity in the top 15 m is over 2 km/s, implying a wavelength much larger than the station distance. We therefore discard the corresponding P wave velocity. The median travel times, velocities, and V_P/V_S ratios are summarized in Table 1 and Figures 5d and 5e, assuming constant velocities between two successive stations.

The results from the radial and transverse components are similar, and the obtained P and S wave velocities are consistent with previous studies (Figure 2b). The top 22 m is characterized by a low shear wave velocity (178–259 m/s) and large V_P/V_S ratio (6.2), corresponding to a Poisson's ratio of ~ 0.49 , close to properties of saturated clay. This is compatible with the geologic site conditions of soft soils and close-to-surface water table (Bonilla et al., 2002). The S wave velocities increase quickly with depth to 961 m/s between 50 and 150 m, and the V_P/V_S ratios decrease to 2.1, corresponding to properties of the bedrock. The heterogeneous seismic velocities within the top 150 m may significantly modify surface seismic recordings, making it difficult to directly infer deep structural properties without deconvolving the response of the shallow materials. The coexistence of water and soil in the top layer could increase the site sensitivity to strong shakings during large earthquakes, providing a good opportunity to study nonlinear site behavior with fluid interactions.

3.2. Temporal Changes of Seismic Velocities After the EMC Earthquake

The autocorrelation function (ACF) can retrieve the zero-offset reflection seismogram at a station (Claerbout, 1968). The secondary peak in the ACF indicates the zero-offset reflection from an interface, the location of

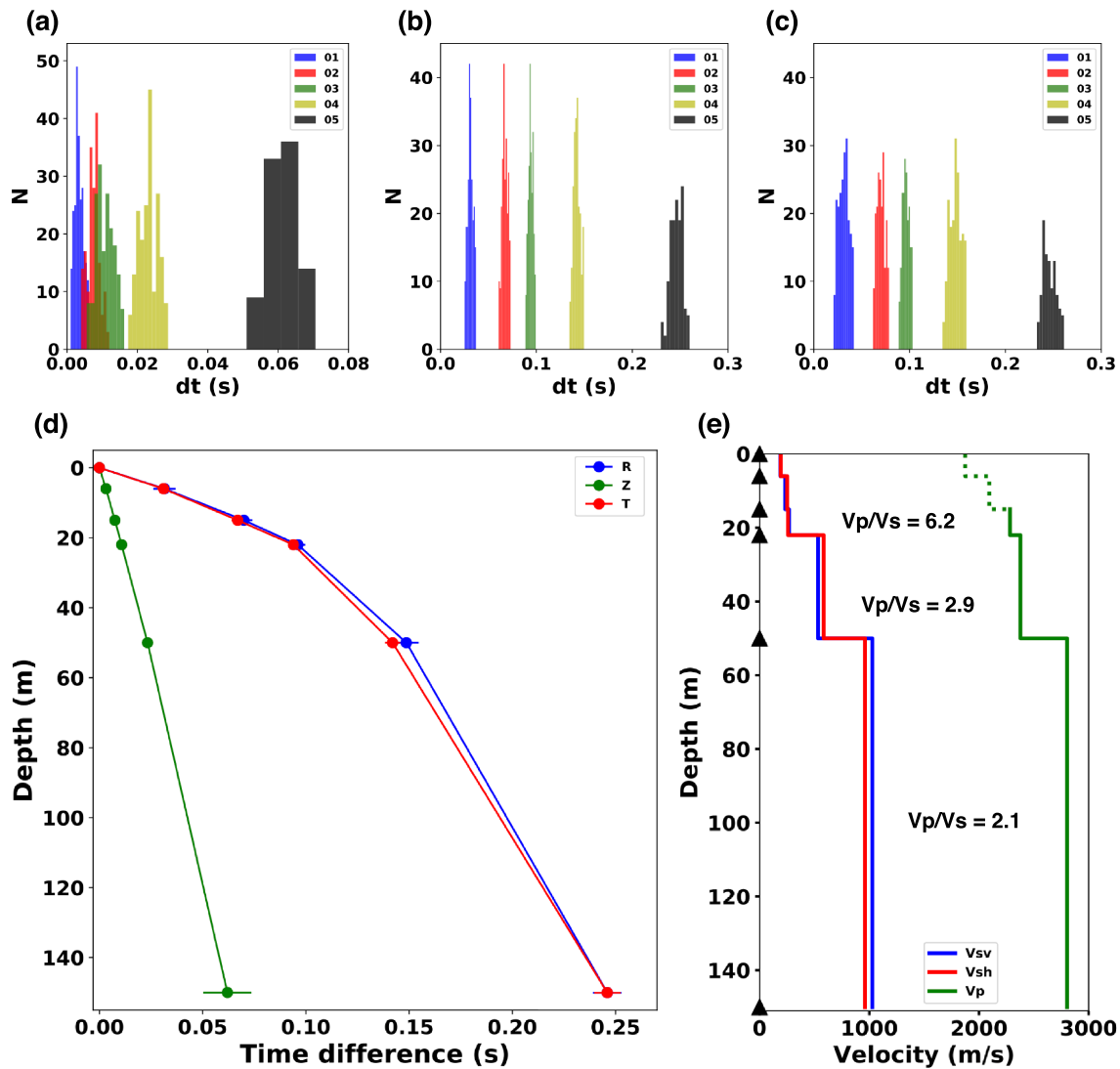


Figure 5. Results from delay time analysis of direct phases. (a–c) The travel time histograms between the surface station (00) and five borehole stations (01–05) from vertical (a), transverse (b), and radial (c) components, respectively. (d) Median (dots) and one standard deviation (error bar) of P (green), SH (red), and SV (blue) wave travel times from boreholes to the surface station. (e) The obtained velocities and V_p/V_s ratios assuming constant seismic velocities between two adjacent stations. The P wave velocity in the top 15 m is plotted with dotted line because of the limited data resolution.

which can be derived based on the travel time and local velocity structure. Travel time variations of the zero-offset reflections reveal material changes (e.g., change of seismic velocities or interface location) between the station and reflection interface and can be used to track the evolution of the subsurface materials in response to various sources (e.g., strong ground motions from large earthquakes and seasonal loadings). The impulse response function (IRF) between two sites can be obtained by deconvolving the waveform of a target site with

Table 1
Average Velocities and Velocity Reductions Between Stations

Depth (m)	0–6	6–15	15–22	22–50	50–150
V_P (m/s)			2283	2379	2804
V_{SH} (m/s)	178	250	259	583	961
V_P/V_S ratio			6.2	2.8	2.1
dV_{SH}/V_{SH} (%)	14.3	7.7	3.4	2.0	2.0

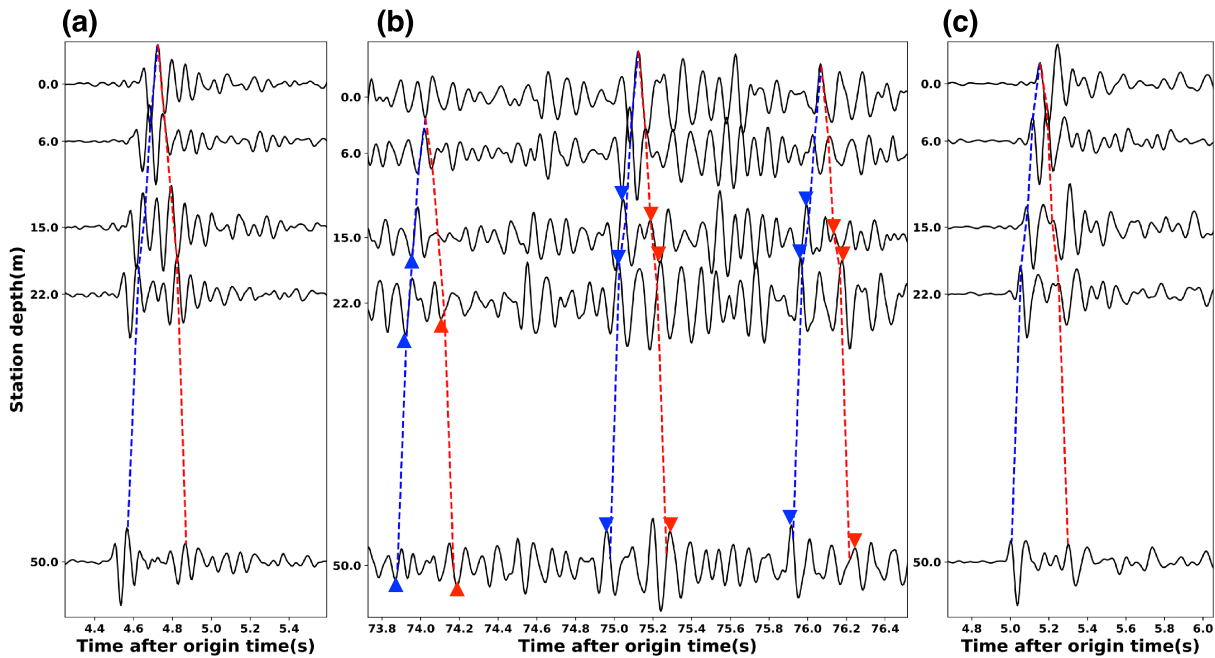


Figure 6. Transverse component waveforms showing free surface reflections before (a), during (b), and after (c) the EMC earthquake. Blue and red dashed lines represent the upgoing and downgoing waves centered at the surface station predicted by the reference travel time obtained from section 3.1. Blue and red triangles in (b) label the observed direct and reflected waves from the EMC earthquake, which are delayed with respect to the reference travel time (blue and red dashed lines).

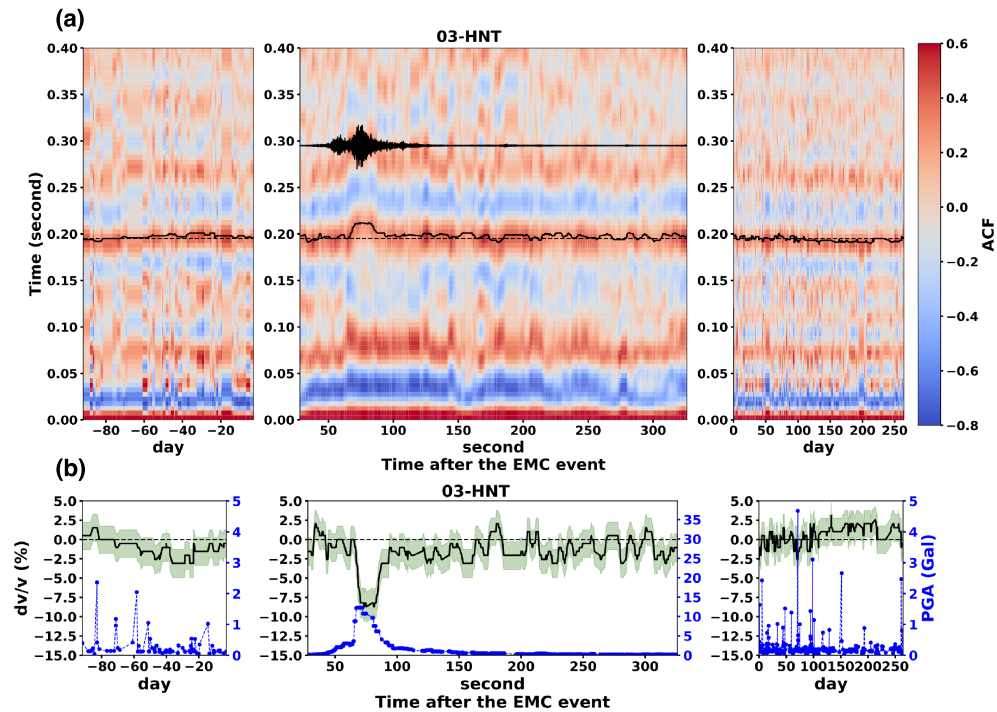


Figure 7. (a). The 10- to 30-Hz autocorrelation functions (ACFs, color representing amplitudes) normalized with zero-lag amplitudes, reference travel time (black dashed lines), and ACF peaks (black solid lines), corresponding to free surface reflections from the transverse component at the 22-m borehole station (03). The left, middle, and right panels are the results before, during, and after the EMC earthquake, respectively. Horizontal axes represent the time relative to the occurrence time of the EMC earthquake, and vertical axes indicate the lag time for autocorrelation. The 10- to 30-Hz waveforms from the EMC event are plotted in the middle panel. (b) The corresponding velocity changes (black solid lines), velocity reduction oscillations (green shaded areas), and PGAs (blue dashed lines with dots) inside the autocorrelation time windows. Please note the irregular time axes, different time and PGA scales between the middle and left/right panels.

Table 2
Average Velocity Changes From Surface to Borehole Depths

Depth (m)	0-6		0-15		0-22		0-50		0-150
Method	IRF 01-03	IRF 01-04	IRF 00-02	IRF 00-03	ACF 03	IRF 00-04	ACF 04	IRF 00-05	
Reference travel time (second)	0.0315	0.032	0.064	0.094	0.0985	0.142	0.146	0.245	
Velocity reduction $\pm$ error (%)	14.3 ± 3.3	14.1 ± 1.8	10.9 ± 3.1	8.5 ± 2.1	8.7 ± 3.1	6.3 ± 2.1	6.9 ± 3.1	4.5 ± 2.0	

that of a reference site (often at downhole or bedrock site). Tracking the IRFs of surface-borehole station pairs is often used to monitor the surface layer, where velocity reductions are the most pronounced in response to strong motions. Bonilla et al. (2019) resolved a 30% *S* wave velocity reduction from IRF analysis between a surface and 300-m-deep borehole station in Japan, and a 50% reduction from ACF analysis at the surface station, after the M_w 9 2011 Tohoku earthquake with a PGA of 585 Gal. Using these two methods, we monitor temporal changes of seismic velocities at the GVDA site with a PGA of about 39 Gal in response to the 2010 $M_{7.2}$ EMC earthquake.

3.2.1. Autocorrelation

Figure 6 shows SH waveforms before, during and after the EMC earthquake with clear free surface reflections. Data of the 150-m sensor does not contain clear surface reflections because a strong 87-m-deep reflection interface traps most of the energy at shallower depth and is therefore not shown. The travel times are consistent with those predicted by the velocity model in section 3.1 (blue and red dashed lines in

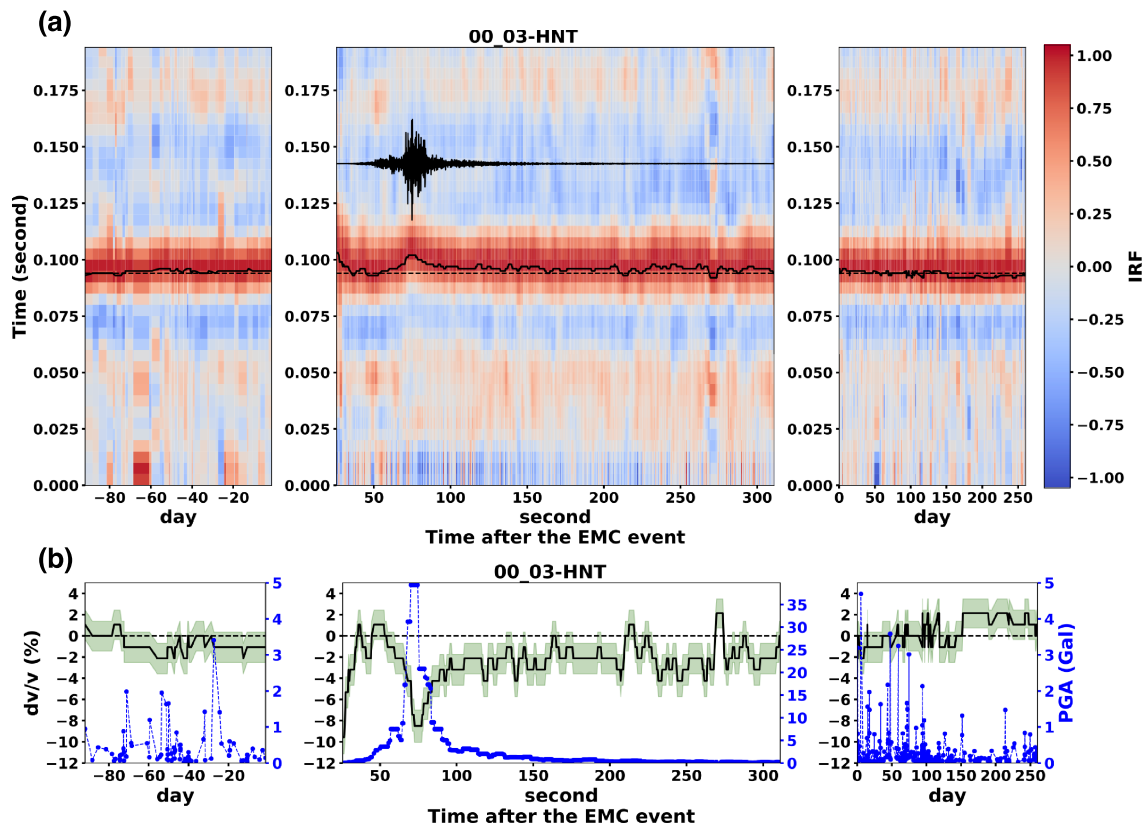


Figure 8. (a). The 1- to 30-Hz impulse response functions (IRFs, color representing amplitudes) normalized with the maximum amplitudes, reference travel time (black dashed lines), and IRF peaks (solid black lines) from the transverse component between the surface station (00) and 22-m borehole station (03). The layout is similar to Figure 7. The 1- to 30-Hz waveforms at the shallower station (in this case, Station 00) from the EMC earthquake is plotted inside the middle panel. (b) The corresponding velocity changes (black solid lines), velocity reduction oscillations (green shaded areas), and PGAs (blue lines with dots) inside the deconvolution time windows. Please note the irregular time axes, different time and PGA scales between the middle and left/right panels.

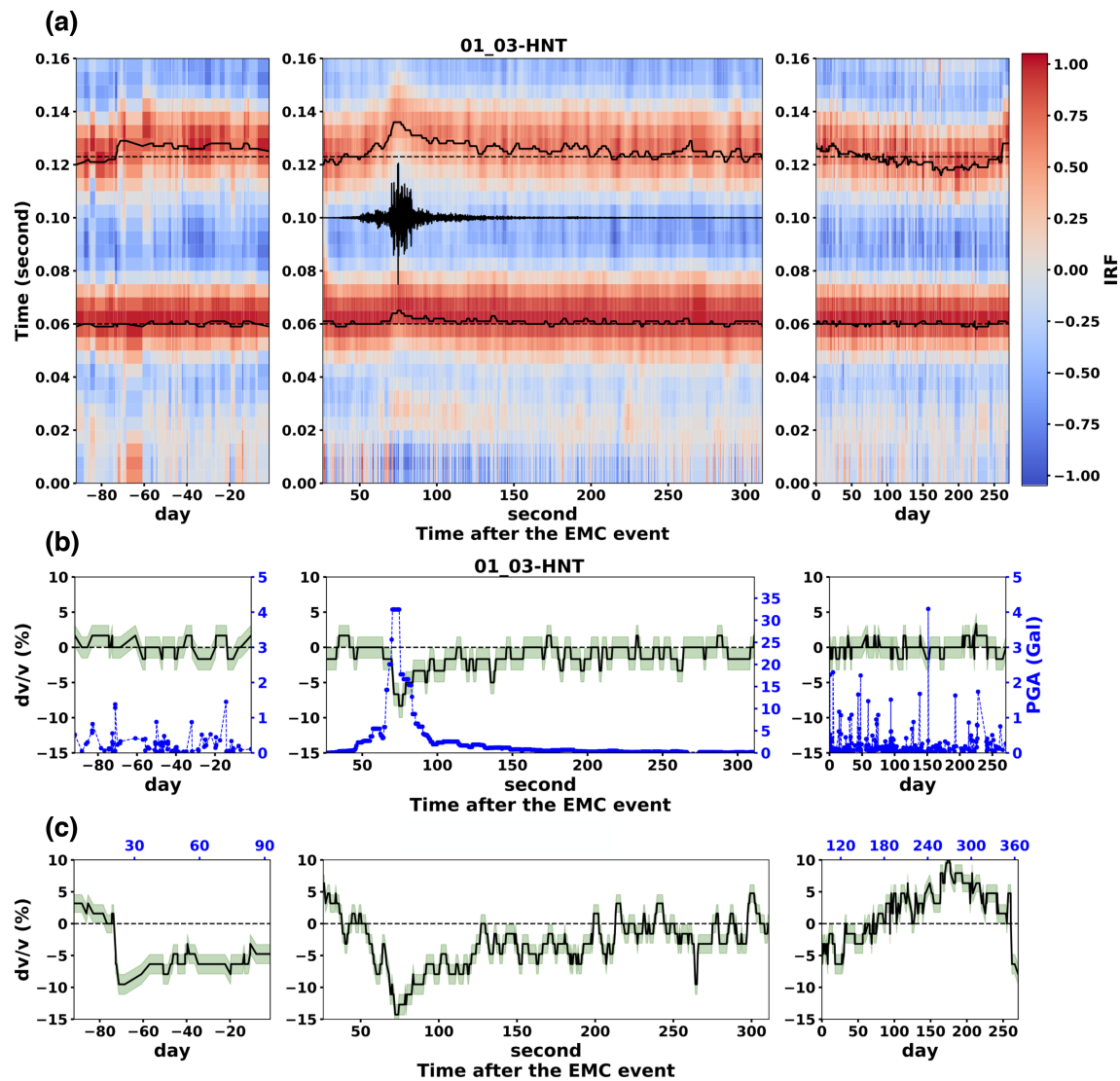


Figure 9. The 1- to 30-Hz IRFs (a) and corresponding velocity changes (b, c) from the transverse component recorded by the 6- (01) and 22-m (03) borehole stations. (b) The average velocity changes between Stations 01 and 03, obtained from the IRF primary peak at ~ 0.06 s. (c) The average velocity changes in the top 6 m, obtained from the IRF secondary peak at ~ 0.12 s, corresponding to waves traveling from Station 03 to surface then reflected to Station 01. The blue horizontal axis ticks in (c) indicate Julian days in 2010.

Figure 6), but during the EMC earthquake the SH waves (blue and red triangles in Figure 6b) are delayed with respect to the reference travel time. Since the direct and free-surface-reflected waves are not well separated at the 6- and 15-m borehole stations, the ACFs are analyzed at the 22-m (03) and 50-m (04) borehole stations.

We first automatically pick the *S* wave arrivals (Ross et al., 2016) and band-pass filter the waveforms at 10–30 Hz. For all the earthquakes before and after the EMC event, the ACFs are calculated in 5-s time windows starting 1 s before the *S* wave arrivals. For the EMC earthquake recordings, ACFs are obtained in 5-s moving time windows with 90% overlap. We then apply a 10- to 30-Hz band-pass filter to all the ACFs, normalize with the zero-lag amplitudes, interpolate them from 200 to 1,000 samples per second, and pick and track the secondary peaks' locations, that is, the two-way travel times between the station and reflection interface (i.e., free surface). The obtained two-way travel times (t) are smoothed with a five-point median filter to eliminate outliers. The reference two-way travel time t_0 is defined as the median travel time from all the small events with PGAs less than 5 Gal. The corresponding velocity change is

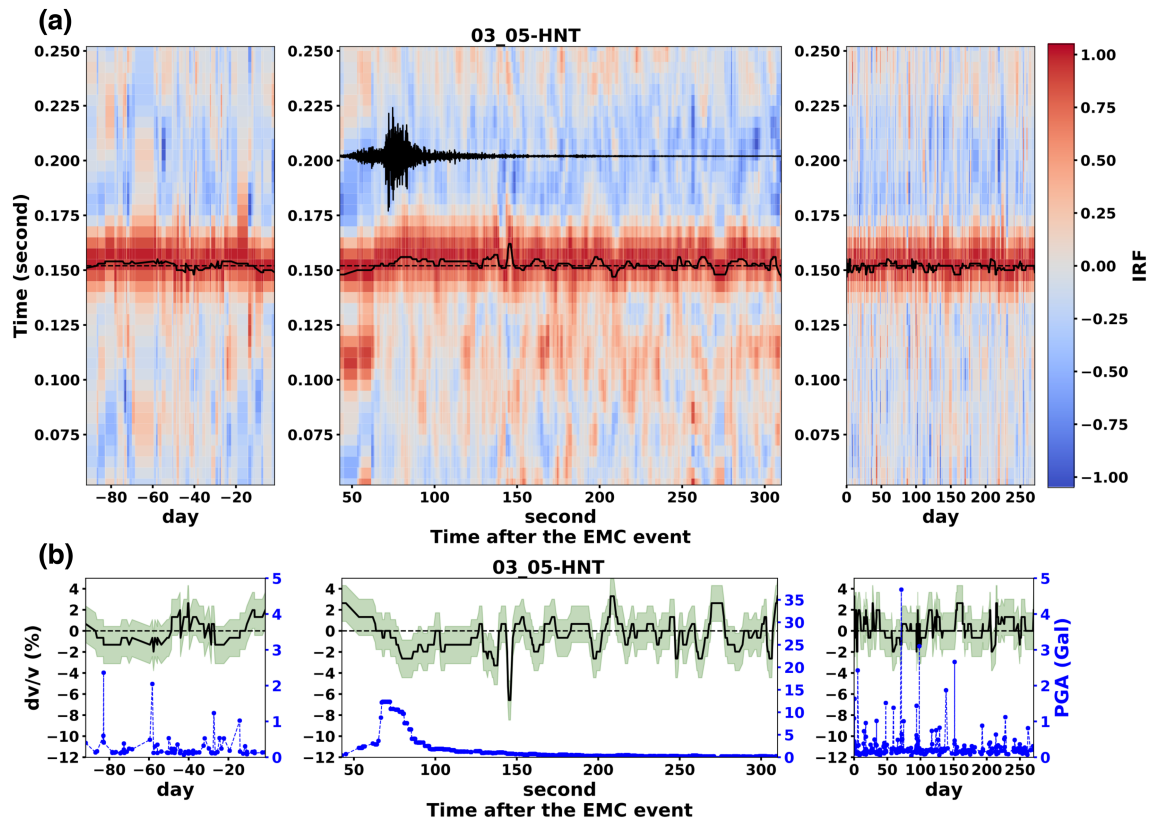


Figure 10. The 1- to 30-Hz IRFs (a) and corresponding velocity changes (b) from the transverse component between the 22- (03) and 150-m (05) borehole stations. The layout is the same as Figure 8.

$$\frac{dv}{v_0} = \frac{v-v_0}{v_0} = -\frac{t-t_0}{t_0}, \quad (1)$$

with v_0 and v being the velocities before and after the strong motion respectively. Velocity reduction oscillations caused by uncertainties of reference travel time are obtained as the velocity reductions with respect to the travel times that are one standard deviation away from t_0 for all small events. The velocity reduction error is defined as the maximum background velocity variation with respect to t_0 from all the small events.

We filter out the signals below 10 Hz since the surface wave energy from the EMC earthquake dominates the waveforms in this frequency band. The 5-s time window is chosen because it generates a good balance between stabilizing the results and emphasizing the signal strength of reflected phases. We also discard the event or time window from the EMC earthquake if the average SNR is smaller than 2 at 10–30 Hz, or the ACF's secondary peak value is smaller than 0.2. Except for the EMC earthquake, events are discarded if they are shallower than 5 km, or generate PGAs larger than 5 Gal at the GVDA site, or the absolute deviation of the corresponding two-way travel time from t_0 is outside the 80th percentile.

Figure 7 shows the ACFs and velocity reductions from the transverse component (HNT) at the 22-m station (03). ACFs from the radial component (HNR) show similar results and are given in Figure S1 (supporting information). The reference two-way travel time t_0 (Figure 7, black dashed lines) is 0.195 s, corresponding to the *SH* wave two-way travel time between the station and free surface. The average *S* wave velocity in the top 22 m drops abruptly by $8.7 \pm 3.1\%$ (with 3.1% indicating the maximum background velocity variation) when the PGA reaches the maximum (~ 39 Gal) at the surface wave arrival ~ 75 s after the EMC earthquake. This is consistent with the observation from waveforms (Figure 6) that the direct and free-surface-reflected *S* waves are delayed between the surface and 22-m borehole stations relative to the reference. The recovery process only takes ~ 200 s. The recovery time scales are estimated in section 3.2.3.

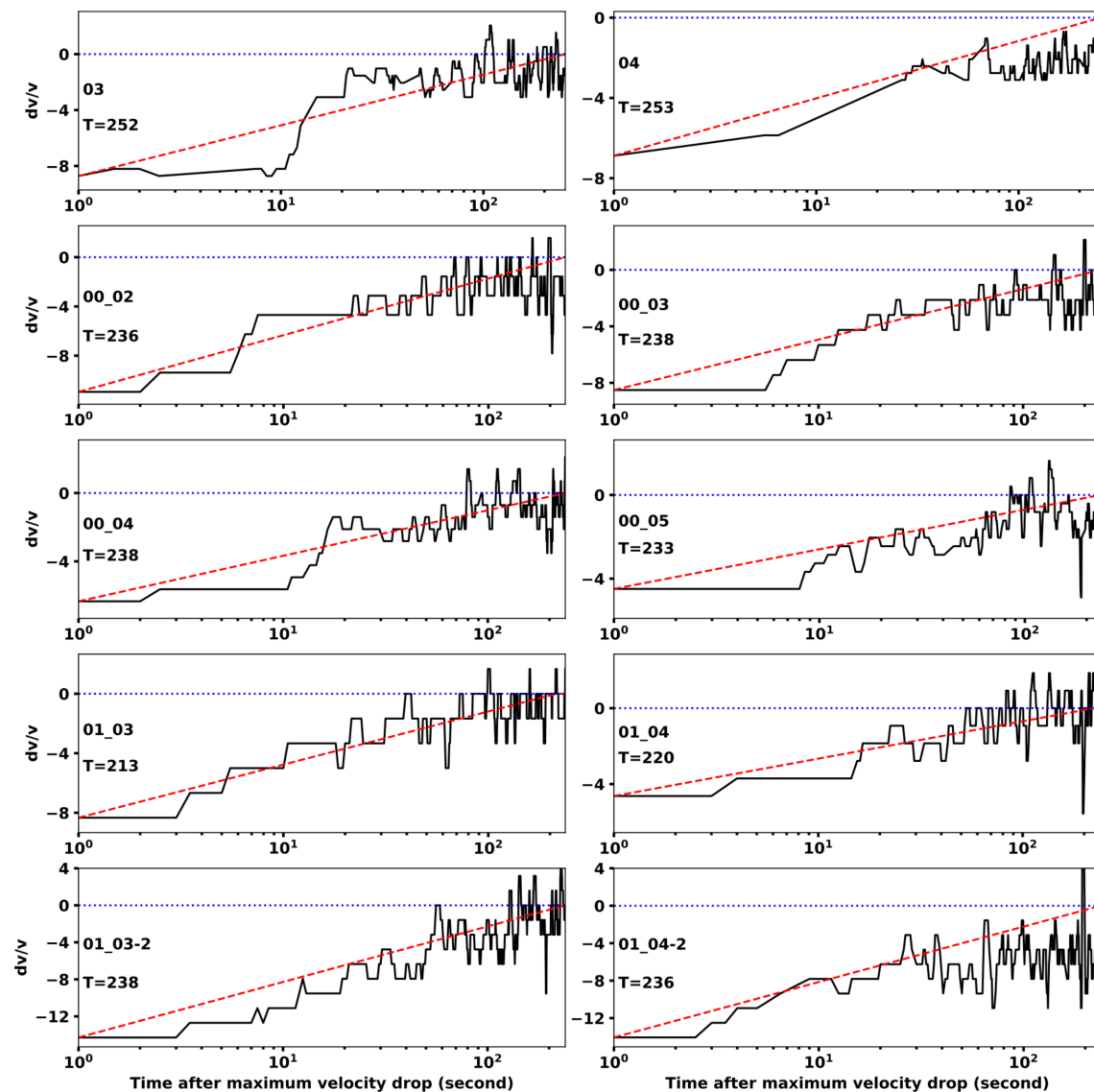


Figure 11. Recovery time estimation of velocity changes. Black and red lines represent observed dv/v curves and the best fitting linear functions. Each plot is labeled with the station/station pair numbers and the corresponding recovery time (T) in seconds. The top two panels show results based on ACF analysis at Stations 03 and 04, middle six subplots are IRF analysis results from six station pairs, and bottom two panels show results based on the secondary peaks of IRFs from 01-03 and 01-04, respectively.

The transverse and radial component ACFs from the 50-m borehole station (04) are presented in Figures S2 and S3 (supporting information). Compared with the results from Station 03, the velocity variations show a similar pattern with an abrupt drop followed by a gradual but quick recovery process after the EMC event, with the reference two-way travel time (0.292 s) consistent with the local velocity structure. The average S wave velocity in the top 50 m decreases by $6.9 \pm 3.1\%$ when the surface waves arrive and recovers in ~ 200 s. Table 2 summarizes the reference travel times and velocity reductions from the analysis.

3.2.2. Seismic Interferometry

The evolution of the S wave IRFs from different sensor pairs are also analyzed for temporal changes of seismic velocities at various depth ranges. In the same time windows defined in the ACF study (section 3.2.1), we apply the multitaper spectral analysis method (Prieto et al., 2009; Thomson, 1982) with five tapers and 1% water level control to calculate the spectral ratios and then perform inverse FFT to obtain the IRFs. We band-pass filter the waveforms and IRFs at 1–30 Hz and interpolate the IRFs from 200 to 1,000 samples per second. The IRF peak locations correspond to the S wave travel time t between the two stations,

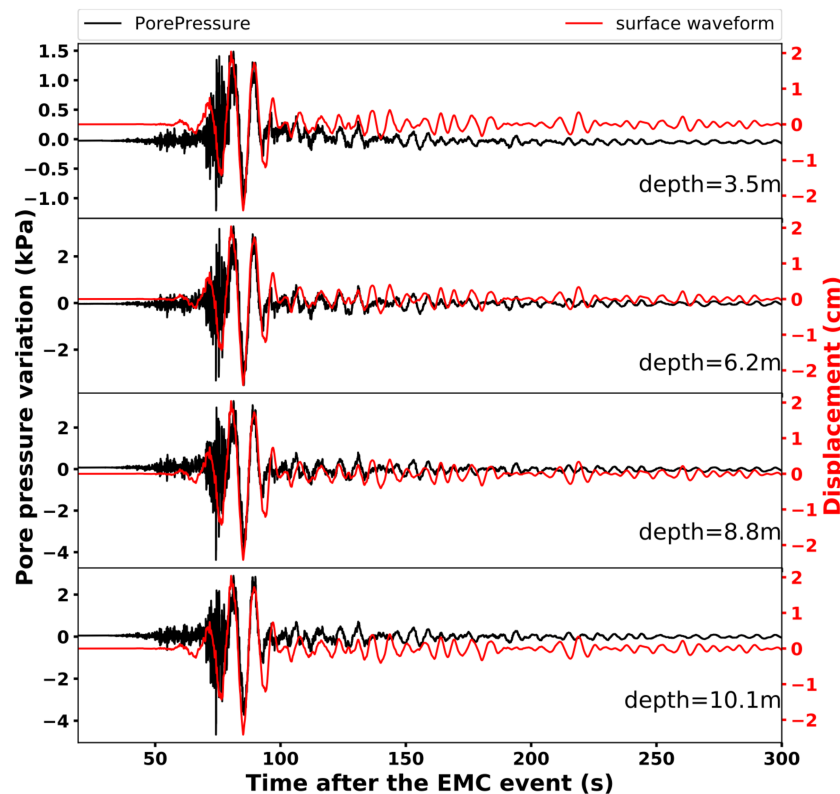


Figure 12. The pore pressure recordings (black lines) from the GVDA at different depths and transverse displacement (blue lines) at the surface station from the EMC earthquake high-pass-filtered at 0.1 Hz. Mean values of each pore pressure trace are removed.

which is smoothed with a five-point median filter to eliminate outliers. The reference travel time t_0 is defined as the median value from all events with PGAs smaller than 5 Gal. Selection criteria and result uncertainties are similar as in the ACF analysis, except that the SNR is defined in the 1- to 30-Hz band and an event is discarded if the IRF value at the picked peak is smaller than 0.8 after normalized by the maximum IRF amplitude. With the highest frequency of 30 Hz, IRFs are stable and reliable between stations separated by at least one wavelength at 30 Hz, e.g., 00-05, 00-04, 00-03, 00-02, 01-03, and 03-05.

Figures 8–10 and S4–S7 (supporting information) illustrate the IRFs and velocity reductions from the transverse components at seven station pairs; Table 2 summarizes the results. Radial components show similar results and Figure S8 (supporting information) provides an example for Station Pair 01-03. The travel times between stations are consistent with the local velocity structure obtained in section 3.1. The velocity variation patterns are similar to the observations from the ACF analysis, with an abrupt reduction at the largest PGA followed by a gradual recovery in ~ 200 s. In Figures 8 and S4–S6 (supporting information), the maximum velocity reductions are $10.9 \pm 3.1\%$ (00-02), $8.5 \pm 2.1\%$ (00-03), $6.3 \pm 2.1\%$ (00-04), and $4.5 \pm 2.0\%$ (00-05), averaged from the surface to 15, 22, 50, and 150 m, respectively. The IRFs from Station Pairs 01-03 (6–22 m) and 01-04 (6–50 m) are shown in Figures 9 and S7 (supporting information). The primary peak ($t^{(1)}$) indicates direct waves traveling from 03/04 to 01, and the secondary peak ($t^{(2)}$) corresponds to free-surface-reflected waves recorded by Station 01 at 6-m depth. The change of $t^{(2)} - t^{(1)}$ represents the temporal change of the material in the top 6 m, shown in Figures 9c and S7c (supporting information). The average velocity changes between 6–22 and 6–50 m calculated from the primary peaks are $8.3 \pm 3.3\%$ (Figure 9b) and $4.6 \pm 1.9\%$ (Figure S7b in the supporting information), respectively. The velocity reductions in the top 6 m obtained from the two station pairs are consistent with each other with values of $14.3 \pm 3.3\%$ and $14.1 \pm 1.9\%$. In addition, travel times in the top 6 m show potential seasonal variations, delayed in spring and summer. More data are required to confirm and analyze the observations, which is not the main target of the current study and left for a future work.

Given the results, we consider the velocity drops to be 14.3%, 10.9%, 9.6%, 6.3%, and 4.5% from the surface to 6, 15, 22, 50, and 150 m. Combined with the corresponding travel times (Table 2), the velocity reductions are calculated to be 7.7% between 6 and 15 m, 3.4% between 15 and 22 m, 2.0% between 22 and 50 m, and 2.0% between 50 and 150 m (Table 2). The calculated velocity drops at 22–50 m and 50–150 m, 2.0%, are comparable to the background velocity variations ($\geq 2.0\%$) from small events, implying no statistically significant velocity reduction below 22 m. This is consistent with IRF results from Stations 03–05 (Figure 10) showing no clear velocity reduction between 22 and 150 m.

3.2.3. Recovery Time Estimation

We estimate the recovery times of the velocity changes by assuming a logarithmic recovery function (e.g., Dieterich & Kilgore, 1996; Johnson & Sutin, 2005). Starting at time (t_0) when the maximum velocity reduction occurs, the velocity change is written as

$$v(t) = \begin{cases} a \cdot \log_{10}(t - t_0 + 1) + b, & t_0 \leq t \leq t_0 + T \\ 0, & t \geq t_0 + T \end{cases}, \quad (2)$$

where a and b are constants and T is the recovery time. In the logarithmic scale, we find the best fitting function $v(t)$ by tuning the parameter T that minimizes the L_2 norm between $v(t)$ and the observed dv/v . Figure 11 presents the observed and best fitting velocity change curves based on all the analyses. The recovery times are similar at all stations with a median value of 236 s. The somewhat larger recovery time from ACF analysis at Stations 03 and 04 may be caused by the large uncertainties in the estimation resulted from noise and modification of waves in the surface layer.

4. Discussion and Conclusions

We analyze interstation velocities in the top 150 m at the GVDA site, using travel times of direct phases recorded by the surface and borehole stations at different shallow depths, and track the ACFs and IRFs for in situ temporal changes of seismic velocities related to the 2010 M7.2 EMC earthquake with an epicenter 206 km from the site. Tables 1 and 2 summarize the results. The derived interstation seismic velocities and V_p/V_s ratios (Figure 5e) are consistent with those from previous studies, and the S wave velocity reductions between various station pairs using the ACF and IRF techniques are compatible with each other. The Garner Valley site exhibits in top 22-m properties of saturated soils with a very low shear wave velocity (178–259 m/s) and large V_p/V_s ratio (6.2), followed by transitions to stiff soils and bedrocks between 22–150 m. During the arrival of the surface waves from the EMC event with a PGA of ~ 39 Gal, the local S wave velocity does not have statistically significant changes below 22 m but abruptly decreases by 3.4%, 7.7% and 14.3% at depths 15–22, 6–15, and 0–6 m, and then recovers to the original levels quickly in ~ 236 s.

The travel time estimations of direct phases based on cross correlating the surface recordings with multiple borehole waveforms, especially the vertically propagating P waves, may be affected by the free surface effect at this site. However, P waves at the surface station vanish only when the incident angle is strictly 0 at the free surface. This is usually not the case even for vertically propagating P waves because of possible topography variations. This reduces the free surface effect and enables clear observation of the vertically traveling P wave at station 00 (Figure 4). Also, the selection criteria require the cross correlation coefficient between surface and borehole recordings to be larger than 0.5, which helps to exclude possible cases when the surface recordings are significantly affected by the free surface effect.

Zhan et al. (2013) showed that velocity change measurements using the noise-based stretching method could be biased by frequency content changes. However, the travel time estimation from the ACFs and IRFs should not be significantly affected by frequency content variations, considering the very short propagation time between stations and relatively narrow frequency band in our analyses although the shallow Q values are small. We therefore use 1- to 30-Hz signals for estimating the local velocity structure and IRF analyses and analyze the ACFs using 10- to 30-Hz data since ACFs from the EMC recordings are significantly distorted by surface waves below 10 Hz. Figure S9 (supporting information) presents an example ACF analysis (transverse component at Station 03) in two narrower frequency bands (10–20 and 15–30 Hz), and Figure S10 (supporting information) shows IRF results in 10–30 Hz at Station Pair 01–03 (transverse component). The ACFs at 10–20 Hz and IRFs at 10–30 Hz are almost identical to those between 10–30 and 1–30 Hz in Figures 7 and 9, while the 15–30 Hz ACFs are less reliable because of the relatively low SNRs of EMC

waveforms in this frequency band. These comparisons show that the obtained results are not affected significantly by attenuation and are independent of the used frequency bands given enough SNRs. Another possible effect that may distort the ACF is local scattering that may cause variations in the ACF side lobes (Figure 7, 0.05–0.1 s). However, the travel time variations do not correlate with the variations in the ACF side lobes. Since the free surface reflection results are more coherent and stronger than local scattering, the measurement of travel time (and velocity change) is not affected by scattering.

It is difficult to constrain the depth range of observed velocity changes without borehole data, considering the very high sensitivity of shallow materials to ground shaking compared to the deeper materials. Chandra et al. (2015) showed that material in the top 15 m at the Garner Valley site behaves nonlinearly already for strain $\geq 10^{-6}$, while below 50 m the response remains linear for strain up to 10^{-4} . This is consistent with our observations on velocity reductions in the top 22 m associated with weak ground motion. Rubinstein and Beroza (2005) showed that temporal changes of *S* wave velocities caused by the 2004 Parkfield earthquake concentrate in the top 100 m. Modeling results from Yang et al. (2019) suggest that temporal changes of seismic velocities in the top 1–3 km can produce observable changes in analyses using Rayleigh waves with periods of 5–20 s. Therefore, resolving the depth range of temporal changes based on sensitivity kernels of Rayleigh waves can be subjected to large uncertainties without information (e.g., from borehole data) on changes in the shallow crust.

Based on the results of this work and previous studies (e.g., Karabulut & Bouchon, 2007; Sawazaki et al., 2006; Wu, Peng, & Ben-Zion, 2009), most of the velocity reductions recover in a short time after the changes occur. This is consistent with the general $\log(t)$ healing of material properties seen in laboratory experiments (e.g., Dieterich & Kilgore, 1996; Johnson & Sutin, 2005). The resolved changes depend strongly, in addition to material properties and loading amplitudes, on the time resolution of the analyses. Given the rapid early recovery and limited temporal resolution, all measurements of temporal changes of seismic velocities should be considered as lower limits. With cross correlations of ambient noise, observed velocity reductions are typically a fraction of 1% because of averaging over days to weeks across wide areas of several kilometers (e.g., Brenguier et al., 2008; Hobiger et al., 2012; Rivet et al., 2011). Analyses of repeating earthquakes and cross correlations of earthquake waveforms with time steps of minutes resolve velocity changes of 1–3% (e.g., Peng & Ben-Zion, 2006; Roux & Ben-Zion, 2014). Using the spectral ratio method and 6-s time windows, Wu et al. (2010) resolved a velocity drop of less than 5% in the top ~100 m generated by weak loadings of ~60 Gal. Bonilla et al. (2019) resolved with ACFs and time steps of 1–2 s about 50% velocity drop in the upper ~100 m during the passage of strong surface waves (~585 Gal). Our calculations of ACFs and IRFs in 5-s time windows with 90% overlap and applying a five-point median filter have a time resolution of ~2.5 s and resolve about 14.3% changes in the top 6 m from loadings of ~40 Gal. Using borehole data at depths (6, 15, 22, 50, and 150 m), we achieve a high spatiotemporal resolution in monitoring the behavior of shallow materials.

Detailed analysis of heterogeneous shallow structures remains highly challenging, especially in the presence of fluids. Nonlinear mechanisms could be easily triggered in the presence of water in porous media (Shapiro, 2003; Zinszner et al., 1997), particularly in the low saturation range (Van Den Abeele et al., 2002). Earthquake-induced permeability changes were inferred in source regions (Nur, 1972) and observed in the shallow crust (Rojstaczer et al., 1995; Wang et al., 2004). Large excess pore pressures are also observed simultaneously with the significant temporal change of seismic velocities at the Garner Valley site in response to ground motions larger than 100 Gal (Steidl et al., 2014). Considering the high water table and large velocity reduction (14.3%) in response to a relatively weak ground motion (~39 PGA), fluid interactions with the subsurface materials likely contribute to the observed results. Figure 12 shows that pore pressure variations at depth down to 10 m are in phase with the displacement data on the surface. The largest velocity drop occurs ~75 s after the EMC event, when the absolute pore pressure values increase with the surface wave arrivals. Deeper than 22 m with decomposed granite, the pore pressure oscillates because the fluids flow in and out from essentially fixed cracks. However, the soft sediments in the top 22 m likely fail partially and move with the fluids. It is possible that the initial pulse of pore pressure weakens (loosens the bonds between soil grains) the poorly consolidated near-surface material, and the material remains in this weakened state (with lower shear modulus and shear wave velocity) while the pore pressure oscillates. This may contribute to the large coseismic velocity drops (14.3% in top 6 m) in response to a small PGA, and the relatively rapid recovery (~236

s) when the material settles down and bonds between soil grains reform strengthening the soil matrix. The precise details of the failure and recovery mechanism should be modeled using additional information (e.g., soil grain size and permeability) about the site.

Analyses of borehole data highlight the significance of shallow structures on observed ground motion and temporal changes of properties, especially in the top 100 m. Despite intensive theoretical research (Bentahar et al., 2006; Johnson & Rasolofosaon, 1996; Lyakhovsky et al., 1997, 2009; Sens-Schönfelder et al., 2019; Sleep & Hagin, 2008; TenCate & Shankland, 1996), the behavior of the shallow and heavily damaged crustal rocks is only understood in general terms. Detailed in situ spatiotemporal monitoring of shallow structures is critical to clarify the properties and processes of subsurface materials. Such information is needed to estimate the stability of the significant infrastructure that exists at the subsurface during ground motion generated by future moderate and large earthquakes.

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