Seismic velocity change patterns along the San Jacinto fault zone following the 2010 M7.2 El Mayor-Cucapah and M5.4 Collins Valley earthquakes

G. Hillers\textsuperscript{1}, M. Campillo\textsuperscript{2}, F. Brenguier\textsuperscript{2}, L. Moreau\textsuperscript{2}, D. C. Agnew\textsuperscript{3}, Y. Ben-Zion\textsuperscript{4}

\textsuperscript{1}Institute of Seismology, University of Helsinki, Helsinki, Finland
\textsuperscript{2}Institut des Sciences de la Terre, Université Grenoble-Alpes, Grenoble, France
\textsuperscript{3}Scripps Institution of Oceanography, University of California San Diego, La Jolla, U.S.
\textsuperscript{4}Department of Earth Sciences, University of Southern California, Los Angeles, U.S.

Key Points:

• We monitor the response of the crust to two triggered deep creep events using noise correlations and strainmeter data
• The obtained seismic velocity change patterns vary along and across the San Jacinto fault zone
• The time, frequency, and lapse time dependence of the velocity changes suggest an upward migrating deformation pattern

Corresponding author: Gregor Hillers, gregor.hillers@helsinki.fi
Abstract

We study temporal changes of seismic velocity ($dv/v$) in the crust around the central section of the San Jacinto fault zone (SJFZ), southern California. Focusing on a 200 day long period in 2010, our analysis resolves two tens-of-days-long successive episodes of reduced velocities that are compatible with signals from the long base strainmeter at the Piñon Flat Observatory. The timing and spatial coincidence of the $dv/v$ patterns with properties of two proposed slow slip episodes suggest that the velocity changes reflect the response to deep creep events that follow the M7.2 El Mayor-Cucapah earthquake and the M5.4 Collins Valley earthquake that occurred 94 days later on the San Jacinto fault. The main slip during the creep events was proposed to occur below 10 km depth. Wavefield properties suggest sensitivity to medium changes above this source zone, in the top 10 km. The distribution of the $dv/v$ reductions shows a strong difference between large values to the west of the SJFZ and significantly smaller amplitudes to the east. The similarity to seasonal velocity changes implies that the results may be controlled by the contrast of mechanical properties across the fault, such as fault-perpendicular shear modulus variations. The imaged $dv/v$ sequences are proxies for evolving deformation and hence stress in the surrounding crust. Our analysis extends the spectrum of methods that can be used to study earthquake and fault interaction, fault zone rheology and dynamics, triggering, and the interplay between creep episodes and earthquakes.

Plain language summary

Geologic fault zones not only slip during small and large, potentially devastating earthquakes. Some faults creep aseismically, which means the slip motion is too slow to generate earthquake waves. To develop a more complete picture of fault behavior it is important to detect and locate such slip transients. A particularly interesting problem is how earthquakes and episodic creep events, which tend to occur in a zone below the earthquakes, interact with each other. Creep events are usually detected using satellite-based methods. In this study we used the ambient seismic waves or noise to detect and locate the changes in the rock properties around the San Jacinto fault zone, southern California, that are caused by two successive creep events along that fault. We applied the most modern seismic analysis techniques to make this first observation of rock changes driven by creep events along a continental fault, but we also show that a co-located deformation-meter picked up the same signals. The observed rock deformation patterns—they are not uniform along and across the fault—and their temporal evolution can help us better understand the processes that happen on and off faults in earthquake-prone regions.
1 Introduction

The 4 April 2010 M7.2 El Mayor-Cucapah (EMC) earthquake occurred on the Laguna Salada fault system in Baja California, Mexico [Hauksson et al., 2011; Wei et al., 2011b] (Figure 1a). It triggered widespread phenomena on many faults in the well instrumented southern California plate boundary region. Analyses of evolving aftershocks and postseismic deformation in the Yuha Desert to the northwest of the EMC rupture indicate that deep localized afterslip propagated in this direction and was followed by a volumetric diffusion-like process [Ross et al., 2017]. The triggered aftershocks in the Yuha Desert include the M5.7 Ocotillo event 72 days later on 15 June 2010. Observed UAVSAR satellite and GPS data in the Yuha Desert show asymmetric migrating deformation patterns after the EMC event that are more pronounced to the northeast of the EMC rupture, along with triggered shallow slip by the Ocotillo aftershock that extends to the Elsinore fault [Donnellan et al., 2018]. To the north of the rupture the EMC event triggered shallow slip on multiple larger and minor faults in the greater Salton Trough area and in the Imperial Valley [Wei et al., 2011a; Donnellan et al., 2014], similar to previous regional earthquakes [Hudnut et al., 1989; Rymer et al., 2002]. The coseismic timing of the slip and the lack of significant afterslip suggest that the triggering involves waves excited by the EMC event. Dislocation modeling confined the slip to the uppermost few kilometers within the velocity strengthening zone that is characterized by unconsolidated sediments with high pore pressure. Only at the southern end of Imperial Fault the initial shallow slip was followed by fault slip at depth [Donnellan et al., 2014]. Seismicity rates in the Salton Sea Geothermal Field increased after the EMC event [Meng and Peng, 2014], and noise-based monitoring resolved seismic velocity reductions in this area [Taira et al., 2018]. Similar to behavior seen after many medium and large earthquakes, the relative velocity variations $dv/v$ are characterized by a coseismic drop followed by a gradual recovery. The affected depth range is 0.5–1.5 km, comparable to the fault responses in the area.

Here we focus on the effects of the EMC event in the San Jacinto fault zone (SJFZ) region. We apply noise analysis techniques to data from 22 seismic stations to study two tens-of-days-long transient seismic velocity change episodes around the Anza seismic gap (Figure 1). The lack of small and moderate to large earthquakes that defines the Anza gap [Sanders and Kanamori, 1984] is attributed to the localized, simple geometry of this relatively straight ~20 km long locked fault section. This is in marked contrast to the seismically active trifurcation area to the southeast (Figure 1b) where the fault splits into three branches [Hauksson et al., 2012; Ross et al., 2017; Trugman and Shearer, 2017; Cheng et al., 2018]. The observed
\(dv/v\) transients follow the EMC event and the July 7, 2010 M5.4 Collins Valley event that occurred 94 days later at 12 km depth on the SJFZ in the trifurcations area. The study extends the range of \(dv/v\) change observations in the region that have been associated with tidal deformation [Hillers et al., 2015a] and seasonal loading [Hillers et al., 2015b]. The results demonstrate the possibility to probe processes that are associated with crustal deformation, and hence stress redistributions, in an area that is characterized by increased seismic hazard.

The obtained velocity change patterns differ from previously observed drop-and-recovery responses to earthquakes of shallow material. The dependence of the observed \(dv/v\) variations on time, space, frequency, and lapse time indicate a complex evolution of the response. Peak amplitude reductions estimated above \(\sim 10\) km occur well after the EMC event, decrease from west to east across the Anza gap and are smallest in the area between the fault and the Piñon Flat Observatory (PFO) (Figure 1b). The similarity of this \(dv/v\) amplitude distribution to thermally-driven seasonal velocity changes [Hillers et al., 2015b] suggests that the response is related to the spatially variable lithologic and mechanical properties across the Anza region [Sharp, 1967; Allam and Ben-Zion, 2012; Lindsey et al., 2014; Fang et al., 2016]. A comparison of PFO long base laser strainmeter data with the seismic \(dv/v\) changes in the same area shows that seismic velocity reductions coincide with volumetric extension periods following the EMC and the Collins Valley events, and that the timing of peak seismic velocity variations tends to follow better the maximum strain rate instead of strain changes, which can indicate nonlinear material behavior [Rivet et al., 2011].

Key properties of the transient velocity change patterns are compatible with the proposed occurrence of two successive deep (\(>10\) km) aseismic slip events [Inbal et al., 2017]. These slip transients have been inferred from seismicity patterns and geodetic signals and are thought to be triggered by the EMC and the Collins Valley events. Following this scenario the \(dv/v\) changes and strain signals likely reflect the response to these creep episodes in the crust around the San Jacinto fault, albeit in a depth interval above the main slipping zones. Linking the observed effects to the proposed creep episodes puts these slip transients into context with implications for deep episodic creep [Wdowinski, 2009; Jiang and Fialko, 2016] and triggered aseismic motion [Meng and Peng, 2016] in the area. Our results show that the combined analysis of high quality strain data and seismic \(dv/v\) observations can provide additional constraints on aseismic deformation in a strike-slip faulting environment and can therefore contribute to our understanding of triggering processes and earthquake and fault interactions on regional and local scales. The employed techniques provide a complementary monitoring tool for defor-
mation processes along active continental fault segments by extending the resolution of geodesy, satellite, and seismicity and tremor based approaches.

In Section 2 we outline the analysis methods used. Section 3 contains a discussion of mechanisms that could affect the resolution of the transient velocity change patterns. After we demonstrate that the signals are not controlled by the wavefield changes associated with the EMC aftershock activity, we separate the slow slip responses from seasonal variations (Section 4.1), analyze the frequency and lapse time dependence (Section 4.2), estimate the spatial distribution of the time dependent response (Section 5), and compare the seismic velocity change time series to PFO strainmeter data (Section 6). The results and implications are discussed in the final Section 7.

2 Seismic data and analysis

We process data from 16 three-component broadband surface stations of the CISN and AZ network, and from six Earthscope/UNAVCO Plate Boundary Observatory (PBO) three-component 2-Hz sensors that are installed in boreholes between 120 and 230 m around the Anza seismic gap in the SJFZ area (Figure 1b). To study potential dependencies between properties of the ambient wavefield and the targeted $dv/v$ variations we consider cataloged information on the seismicity in the southern California plate boundary region [Hauksson et al., 2012] together with wavefield markers (Figure 2) and properties of the cross-correlations analyzed for the $dv/v$ estimates (Figure 3).

To compute single station wavefield markers from surface station data we first high-pass filter daily surface station seismograms above 0.05 Hz. We split these records from each station and component into 96 15 minute windows and estimate the spectral power. The horizontal power $H$ is the sum of the N and E component estimates, and the vertical power $V$ uses the Z-channel data. $H$ and $V$ are estimates of the arriving seismic energy, and the ratio $H/V$ is a proxy for the $S$-wave to $P$-wave energy ratio [Shapiro et al., 2000; Hennino et al., 2001]. From the 32 values per eight hours we can take any quantile of the three $H$, $V$, $H/V$ estimates for each frequency and average over three frequency ranges 0.2–0.4 Hz, 0.4–0.8 Hz, 0.8–1.6 Hz and finally over all stations in the network. Figure 2c shows the resulting 0.9 quantile network averages using an additional 3-point running median filter (thin lines), that is again smoothed with a moving average (thick lines) to highlight the long-term trend. The time series have been split at the time of the EMC earthquake for the temporal averaging to highlight the effect of the increased seismicity. $H$ and $V$ values are in decibel before scaled
by the median energy in the 50 days prior to the EMC event. For the \(H/V\) ratios in Figure 2d we show the 0.2, 0.5, and 0.8 network quantiles on a linear amplitude scale. The time series are scaled to the respective median energy in the 50 days prior to the EMC event. A below-average level for the low frequencies at longer times reflects the seasonal energy variation.

For the seismic velocity change analysis we construct two databases of noise cross-correlation functions between all sensors within 40 km distance using seismic records from 2008 to 2015. We remove the instrument response and whiten each daily record in the \(0.05-5\) Hz frequency range. The applied sign-bit clipping yields higher signal-to-noise ratios (SNR) of the correlation functions compared to clipping at some value that is based on the amplitude distribution in each processing window. Vertical component data from the day of year (doy) 1 to 300 are used to study seasonal variations between 2008 and 2015 (Figure 4). For this the correlations are stacked at \(\pm 2\) days and downsampled by a factor of two. A high-resolution analysis of the EMC and Collins Valley events responses in 2010 is performed using the daily nine-component (9C) correlations between \(t_0-50\) days to \(t_0+150\) days, where \(t_0\) is doy 94, 4 April 2010 (Figure 5). This window is centered on the EMC induced transient and its extent is limited by the computational cost of the employed analysis method. The velocity change signals in this window are compared to records from the long baseline laser strainmeter at the PFO site (Section 6, Figures 6a–c).

We apply three different approaches to the 2010 9C high-resolution database for estimates of daily \(dv/v\) variations between 0.2 and 2 Hz to assess the variability and the sensitivity to wavefield changes. The first two approaches estimate \(dv/v\) changes from variations between each daily correlation function and a reference stack compiled from the correlations in the analysis period. We use the time-domain stretching method [Grêt et al., 2005; Sens-Schönfelder and Wegler, 2006] and the frequency-domain Moving Window Cross Spectral (MWCS) or Doublet technique [Poupinet et al., 1984; Clarke et al., 2011] to measure the travel time changes. The final estimate in both cases is the average over all station and component pairs.

In the third method the \(dv/v\) time series is obtained by inverting estimated changes between all possible pairs of daily correlations [Brenguier et al., 2014; Gómez-García et al., 2018]. These individual estimates are obtained with the MWCS technique. This method does not rely on a reference function and also not on the assumption of stationary wavefield properties. The number of \(dv/v\) estimates used in this approach scales quadratically with the number of days per station and component pair, in contrast to the linear scaling using the reference stack. This can better constrain the \(dv/v\) model at times with no or low-quality estimates. The final \(dv/v\)
time series is obtained with a standard linear least-squares scheme [Tarantola and Valette, 1982; Brenguier et al., 2014]. Weights in the inversion scale with the similarity between the full waveforms. This similarity estimate is the scaled cross-correlation coefficient $cc$. For a network average, data from all stations and components with $cc \geq 0.85$ are inverted simultaneously for a given frequency band and lapse time window. In the inversion we use a correlation length parameter $\beta = 3$ days. The high similarity of results obtained with $\beta$ values between 1 and 7 days shows that the final $dv/v$ time series are not sensitive to variations within this range. Significant differences occur only if $\beta$ is varied over a wide range [Brenguier et al., 2014; Gómez-García et al., 2018]. The value employed does not degrade the daily resolution and hence does not affect the discussion of the transient onset. Our results are insensitive to the $\alpha$ parameter that scales the model covariance matrix in the inversion [Gómez-García et al., 2018]. This third method is also used for the low-resolution seasonal analysis.

The three measurements are repeated after applying an Singular Value Decomposition (SVD) based Wiener filter [Moreau et al., 2017] to the noise correlations. The filter removes incoherent data across the daily correlations. In contrast to previous filtering approaches [e.g. Stehly et al., 2011] the filter does not dependent on a model or reference waveform. The Wiener filter works on the singular values of data-patches of size $M \times N$. $M$ is the number of days along the calendar time, and $N$ is the number of samples along the correlation lag time. We apply a low-pass filter adapted to each $dv/v$ frequency analysis range before application of the Wiener filter and use then a fix $N = 5$. We use $M = 20$ days to mute fluctuations on the 10 days scale seen in Figure 3h and keep the largest $M/2$ singular values. This window size results in a temporal smoothing on the order of five days. As shown below, the significantly improved SNR in the noise correlations leads to better $dv/v$ estimates.

In addition to the earthquake catalog and wavefield markers, the indicators obtained from the correlation based $dv/v$ estimates include a marker $P$ of the peakedness of the daily $dv/v$ distributions for the two reference-stack based approaches. For the inversion based method, we consider how many $dv/v$ estimates associated with a daily correlation have entered the inversion (Figure 3g). This metric is controlled by the $cc = 0.85$ cutoff threshold. We define $P$ as the ratio between the distribution peak value and the distribution width. The width is estimated as the most narrow interval that contains some fraction of the total number of values in the distribution. The peak value and the width are inversely proportional to each other, so the ratio accentuates this trend. Large $P$ indicates peaked, narrow distributions and hence an overall well constrained component and network average, whereas small $P$ characterizes less
consistent or more variable estimates. The $P$ values in Figures 3a, 3d capture the evolution of the color-coded $dv/v$ distribution in Figures 3b–c, 3e–f. Small $P$ values are not problematic as such. They are expected if the spatial $dv/v$ pattern varies on scales that are smaller compared to the network size [Obermann et al., 2013; Hillers et al., 2015c]. However, together with other wavefield markers the temporal variation of $P$ can be indicative of periods with poorer $dv/v$ resolution.

### 3 Relation between wavefield variations and velocity changes

We find that the main variations in the $H$ and $V$ energy time series are dominated by the EMC aftershock sequence and not by properties of the local seismicity (Section 3.1). It is thus important to assess to what degree these wavefield changes affect the $dv/v$ estimates in the study area (Section 3.2).

#### 3.1 Seismicity and wavefield changes

The aftershock distribution in the EMC area (EMC box in Figure 1a; data in Figure 2a) shows the typical magnitude dependent decay pattern described by the Omori law, which means the larger aftershocks that excite waves with longer periods decay faster compared to the small magnitude events that have thus a longer lasting effect on the local and regional high-frequency wavefield. This frequency dependent effect is clearly seen in the $H$ and $V$ energy observed in the study area (Figure 2c). The $0.8 - 1.6$ Hz high-frequency wavefield shows excess energy for some 40 days after the main shock, whereas the $0.2 - 0.4$ Hz low-frequency estimates are affected only for a few days.

The seismicity increase at $t_0$ in the areas outside the EMC main shock area and outside the study area (red circles in Figure 1a; red data in Figure 2b) is mainly associated with EMC triggered activity to the north of the main shock zone along the San Andreas, San Jacinto, and Elsinore faults. Events occurring after $t_0$ along the San Jacinto fault (orange circles in Figure 1b; black data in Figure 2b) are thought to be triggered by deep creep that is in turn triggered by the EMC event [Inbal et al., 2017]. The persistently elevated level of small-magnitude seismicity after the EMC event in this area is more evident from a more complete detection list obtained with template matching techniques [Meng and Peng, 2014]. This seismicity does not exhibit the Omori-type behavior that characterizes the EMC aftershock sequence and the $H$ and $V$ time series. Thus the local seismicity does not govern the variations in these local wave-
field markers. Similarly, the M5.4 Collins Valley event in the San Jacinto fault area 94 days after the EMC event does not affect the wavefield properties on the here considered time scales.

Interestingly, the high-frequency energy peak 72 days after the EMC event (Figure 2c) is associated with the aftershock sequence of the M5.7 Ocotillo event in the EMC main shock area, and not with the two M4.9 and the M4.2 events that occurred 70 days later on 13 June 2010 within 23 seconds 8–9 km below the network in the study area (Figure 1a). This further highlights the effect of seismicity in the EMC area on the wavefield in the study area.

In contrast to the $H$ and $V$ energy estimates, the $H/V$ wavefield marker exhibits a remarkable stability (Figure 2d) on the eight hour time scale. The $H/V$ ratio fluctuations are on the order of 10% and therefore similar to the variability that Hennino et al. [2001] report for diffusive earthquake coda averaged over 15 s. The $H/V$ time series are unaffected by the strongly variable seismic energy release. We can discern, perhaps, a very small effect in the first 10–20 days after $t_0$ in the 0.4–0.8 Hz range, where the ratio remains slightly increased and the fluctuations are smaller. This behavior is compatible with an excess excitation of shear waves due to scattering. Hence despite the wavefield amplitude changes, and despite the changes in the flux direction associated with the intermittent source region around the EMC rupture the partitioning of the multiply scattered wavefield in the study area remains relatively constant. This stability supports our application of noise based analysis techniques.

3.2 Method comparison

The comparison of wavefield markers and $dv/v$ estimates associated with the three different methods suggests that the results obtained with the stretching method are biased by the effect of the remote EMC aftershock activity on the wavefield amplitudes in the study area [Zhan et al., 2013]. On the other hand, the amplitude variations and the changes in the flux direction do not seem to affect the phase based $dv/v$ estimates, which is compatible with theoretical and observational work [Colombi et al., 2014] and with the overall stable $H/V$ ratios. These conclusions are drawn from Figure 3, and are further supported by the lapse time dependent results discussed in Figure 5 that show that the velocity change estimates are related to medium changes, and not to source changes. The conclusions are also compatible with the findings of Hillers et al. [2015b] and Hillers et al. [2015c] who both favored MWCS based results over stretching method based results, because changes in the excitation pattern left a much weaker footprint in the MWCS-derived $dv/v$ signals.
We focus on the resolution of the transient signal after the EMC event. This signal is superimposed on the seasonal variation. We see that the abrupt velocity drop obtained with the stretching method coincides with the EMC earthquake (Figure 3b), and then gradually approaches the long-term trend. Similarly, the $P$ value in Figure 3a shows that the consistency over the network drops at $t_0$ and recovers over 70 days to the final level. As said, the widening of the distribution is not problematic as such, as it could be associated with lateral variations in the velocity changes. However, the low $P$ values also reflect the removal of many spurious measurements from the distributions that cluster at both ends of the wide grid search range. That is, the algorithm does not converge to stable, physically meaningful solutions. This behavior is similar to the evolution of the energy estimates in Figure 2c and suggests that the stretching method provides the least stable estimates because of the wavefield amplitude changes.

In contrast to the coseismic drop, both MWCS $dv/v$ results in Figures 3d to 3i show a more gradual velocity decrease after $t_0$ that is followed by the recovery to the long term trend. The associated black $P$ values in Figure 3d are remarkably stable and hence insensitive to the observed wavefield changes. The $cc$ based control parameter in Figure 3g associated with the inversion method shows a significant drop-and-recovery-behavior in the period following $t_0$ that strongly reflects the EMC aftershock induced wavefield changes. Importantly, however, in contrast to the stretching method based results the associated $dv/v$ estimates in Figure 3h show a different, i.e., unaffected, gradual response pattern that is similar to the MWCS reference stack approach.

We repeat this analysis using correlations that have been Wiener-filtered (Figures 3c, f, i). All results are cleaner and significantly improved, although the filter cannot completely remove the bias from the stretching results at the times of the strongest amplitude changes. The number of $dv/v$ measurements in the MWCS based inversion approach quantifies the positive effect of the filter. About $4.7 \times 10^6$ estimates with $cc < 0.85$ were removed from the total $12.5 \times 10^6$ compared to the removal of $9.7 \times 10^6$ estimates from the unfiltered data. The consistency between the results associated with each correlation tensor component and the results obtained with a joint inversion of all data (gray and black lines, Figures 3h, i) implies a high degree of mixing or randomization of the wavefield, again compatible with the conclusions drawn from the stable $H/V$ ratio. We continue with inverted $dv/v$ time series that we obtained from MWCS-analyzed, Wiener-filtered correlation functions.
4 Results

4.1 Separation of slow slip and seasonal responses

The annual network averages including borehole data are shown in Figure 4a for the 0.2–1 Hz range. Compared to the 2008, 2009, 2011–2015 results in gray, the 2010 curves in red show a different post-t0 \( \frac{dv}{v} \) behavior that we have been associating with the proposed aseismic transient triggered by the EMC earthquake. The \( \frac{dv}{v} \) inversion yields a neutral average over the analysis period, which explains the different \( \frac{dv}{v} \) levels at negative times of the 2010 velocity states compared to the data from the other years. This neutral average also controls the sensitivity of the pre-t0 results to the length of the analysis window, which is indicated by the difference between the long and short red curve.

We find that the difference between the 2010 signal and the multi-year average response depends on frequency and position. The clear difference shown in Figure 4a appears at low frequencies, and for the full network average. For a station subset to the west of the fault (Figure 4b), the seasonal signal is significantly above average and thus also dominates the 2010 response, so that the 2010 values are generally within the range of values obtained in other years. In all cases, however, the downward slope of the 2010 transient onset is opposite to the upward showing trend of the average seasonal effect at this time of the year, which also serves as a discriminator.

The amplitude of the seasonal variation decreases with frequency (Figures 4c, d). The implications for the depth resolution are discussed in Section 5.2. The spatial variability of the seasonal response is significant, with a weak or no signal in the region to the east of the fault, and large amplitudes in the area to the west of the Anza gap. These results are compatible with the view that thermoelastic strain governs the seasonal response in the area [Hillers et al., 2015b]. They show now in more detail that this effect likely results from the the lithologic and mechanical contrasts across the Anza region. This involves strong topographic gradients in the region (Figure 1b), material contrasts across the San Jacinto fault [Allam and Ben-Zion, 2012; Lindsey et al., 2014; Fang et al., 2016], changes in the lithology along the fault [Sharp, 1967], and perhaps also variable water saturation as suggested by hot springs in the study area (Figure 1b). The pivotal mechanical properties of the Anza section are indicated by the seismic quiescence of the locked fault segment in that area.

We separate the target signals from the seasonal trend by subtracting the ZZ low-resolution multi-year average from the 9C high-resolution 2010 data (Figure 4b) considering the high similarity between the ZZ and 9C results (Figure 3i). While the 2010 results are relatively sim-
ilar to the average seasonal signal at $t > 50$ days, the neutral-average effect makes it more difficult to study details of the response at $t < 0$ days. We varied the amplitude and level of the subtracted seasonal response to minimize this discrepancy. The overall evolution and shape of the resulting response pattern is not much affected using different scaling strategies, in particular at $t > 0$ days, but the pre-$t_0$ level remains somewhat arbitrary. We set the zero level as the average of the values from before the EMC event, between the EMC and Collins Valley events responses, and after the Collins Valley event related signals, i.e., from $t < 0$ days, $75 < t < 90$ days, and $t > 125$ days.

4.2 Frequency, lapse time, and regional dependence of the transient changes

The zonation reveals a significantly more detailed picture compared to the network averages discussed in the quality assessment. The division into a west, center, and east zone (Figures 4, 5) is based on the similarity or zero-lag correlation coefficient between the underlying 150 days long 2010 $dv/v$ time series obtained from station pairs or triplets in each area. We compute the $dv/v$ curves for five frequency bands between $0.2 – 1$ Hz and $0.6 – 3$ Hz using the same $20 – 50$ s coda window as in the broadband $0.2 – 2$ Hz quality assessment (Figure 5a). Results in the $0.1 – 5$ Hz range (not shown) are compatible with the $0.2 – 1$ Hz results but are characterized by significant fluctuations, even if we adapt the processing parameters to the longer wave periods and longer convergence times scales. We combine data from the surface and the borehole sensors for all but the $0.2 – 1$ Hz results, where the borehole data become too noisy. Above this range we do not find systematic differences in the $dv/v$ responses from the deconvolved $2$ Hz borehole sensor data and the surface stations. For the $0.3 – 1.5$ Hz (Figure 5b) and the $0.5 – 2.5$ Hz range we compute the responses in three lapse time windows $20 – 40$ s, $30 – 50$ s, $40 – 60$ s.

Similar to the seasonal response pattern, peak $dv/v$ amplitudes after the EMC earthquake are largest in the west zone and decrease across the fault towards the east zone (Figure 5a). The inverse frequency dependence of the transient $dv/v$ amplitude change, and the implications of the timing of the frequency dependent peak velocity changes in the west and center zones are discussed in Section 5.2. The transient duration is about $70 – 90$ days. The signals in the west zone show a clear change in the slope of the relaxation at the time of the two successive M4.9 and M4.2 events that occurred 70 days after the EMC event on the San Jacinto fault (Figure 1). We verified the ‘positive kink’ around $t = 50$ days in the west zone using multiple other zonation and averaging strategies. The weak east zone response appears rela-
tively independent on frequency above 0.3 Hz, and the minimum at $t_0+20$ days is reached earlier compared to the west and center zone results.

The main features in the lapse time $\tau$ patterns (Figure 5b) are the inverse $\tau$ dependent scaling of the maximum reductions—later windows, smaller peak amplitude—and the correlation with the peak timing—later windows, later peak timing. All these dependencies are also observed for the $0.5-2.5$ Hz results. In the multiple scattering regime associated with longer lapse times potential source effects are reduced and the change estimates better reflect the medium response. The late-coda results are thus an additional strong indicator that the $dv/v$ variations reflect the deformation induced by aseismic fault motion at depth, which is further discussed in Section 5.2.

The response associated with the Collins Valley earthquake at $t = 94$ days is clearly visible in all results. Compared to the post-EMC results the $dv/v$ reductions are smaller and the transient duration is shorter. Similar to the EMC related observations the velocity reductions in the west and center zones are larger at low frequencies, and the lapse time dependence is also consistent. Note that the time series in Figure 5 are smoothed with a 3-point moving average for better legibility, but we verified that the velocity drop is gradual and not abrupt, as can also be inferred from the timing of the peak about 15 days after the event.

As said, the different velocity states in 2010 and the multi-year average together with the neutral average of the inverted $dv/v$ data contributes to the pre-$t_0$ fluctuations seen in Figure 5. Overall, however, we can resolve the gradual onset of the velocity transient at or shortly after $t_0$. The pre-$t_0$ fluctuations on the 10 days scale that are not eliminated or muted by the filter decrease with frequency and lapse time. This means they decrease as a result of the better wavefield randomization through scattering that tends so reduce signatures of the noise excitation that is related to environmental triggers, changing atmospheric and ocean state patterns, and anthropogenic activity. The onset of the Collins Valley event response, however, is cleaner compared to the EMC event response. This is due to the much better alignment of the 2010 and the average multi-year response around $t = 100$ days (Figure 4b).

5 Constraining the spatial distribution

5.1 Regionalization of the time dependent response

Imaging the lateral distribution of velocity changes is based on the inversion of pair-wise $dv/v$ estimates from all station combinations using statistical end-member models for the scattered energy propagation [e.g., Obermann et al., 2014]. We tested this approach but found that
the inversion results were unstable due to fluctuations in the underlying \( dv/v \) time series from the pair-wise 9C data. The ambiguous definition of the reference and pre-\( l_0 \) level in the de-seasoning step, which is also essential for spatially consistent \( dv/v \) variation patterns, contributes to these problems. We found that data from more than two stations have to be averaged to arrive at sufficiently stable \( dv/v \) time series similar to those discussed in Figures 3–6. We expect that the filtering or decomposition [Hobiger et al., 2012; Richter et al., 2014; Wang et al., 2017] of daily sampled multi-year time series supports more stable spatial inversions.

We adopt a simpler imaging approach based on regionalization [Brenguier et al., 2008; Hobiger et al., 2014; Brenguier et al., 2014]. We found that a triangulation based zoning sufficiently trades-off stacking or averaging—using three times more data compared to the pair-wise approach—and spatial confinement. For this we build station triplets that correspond to all non-overlapping triangles. There are three sites with colocated surface and borehole stations. The associated triangle estimates can thus be build from two or more station triplets. For each triplet we subtract the low-resolution seasonal average from the high-resolution 2010 \( dv/v \) time series. For quality assessment, \( dv/v \) values of each triplet are associated with the number of daily correlation pairs with \( cc > 0.85 \) that entered the \( dv/v \) inversion scheme \( (O(10^5)–O(10^6)) \). These are scaled to triplet specific weights between zero and unity. These weights scale with the average inter-station distance. We then assign the triplet-average \( dv/v \) value to the center position of the triangle. The \( dv/v \) value can be the seasonal peak amplitude (Figure 4f), the estimate at any given day during the study period (Figure 7), or the maximum \( dv/v \) reduction during the EMC event response (Figure 8a). Next we compute weighted \( dv/v \) averages at the center locations of those triangles containing colocated surface and borehole stations, and average over locations that are closer together than 2 km. Finally, we fit a 2D distribution through the weighted values at the center locations [Wessel et al., 2013]. The imaging area is defined by the envelope around 10 km circles around the triangle center locations. A tension factor controls the scale of the variations, similar to the tuning parameters in the least-squares inversions [Obermann et al., 2014]. The chosen value yields distributions that contain the robust features obtained with a wide range of values.

The scattered wave propagation models—also referred to as sensitivity kernels—used in the inversions of spatially variable \( dv/v \) estimates imply that the largest sensitivity to medium changes is at the station location itself [Pacheco and Snieder, 2005]. This is why the more recent implementations of the simpler regionalization strategies [Hobiger et al., 2014; Brenguier et al., 2014] use station locations and not in-between locations as reference points. At least
the study of Brenguier et al. [2014], however, benefitted from a dense, regular network for this task. We also test this idea and compute \( \frac{dv}{v} \) time series using correlations between every ‘master’ station and its five nearest neighbors. We exclude connections between neighbor stations to emphasize the association of the obtained velocity changes with the master location. All other processing steps are the same as in the triangle case. The resulting peak seasonal amplitude distributions are very similar compared to the triangle approach. The key features of the \( \frac{dv}{v} \) response following the 2010 EMC event are also compatible considering the different positions of the triangle centers and the station locations. The two distributions differ most significantly to the north of the Anza gap and east of the fault, in the region between the KNW station and the PMD and PFO stations (Figure 1b). The master station approach puts significantly decreased amplitudes in this area, similar to those on the southwest side of the fault, therefore creating a more fault symmetric pattern compared to the one-sided triangle based results that has overall neutral values in this area. The master approach, however, does not include connections across this area between the KNW and PFO sites, and the five nearest-neighbor pairs cover only a limited azimuthal range. The large KNW amplitudes governed by the connections to the south across the fault are thus interpolated to the east. Because of the explicit delimitation of the triangles we favor the associated triplet results. It has also the appeal of yielding 43 instead of 21 different values before the weighted averaging is applied.

The space-time evolution obtained from the \( 0.3 - 1.5 \) Hz, \( 20 - 50 \) s results (Figure 7) show that velocities decrease first along the fault. The transient following the EMC event lasts for \( 65 - 70 \) days, and the maximum reduction occurs around 30 days after the event to the southwest of the fault, within the Anza gap. The transient reduction is modulated by a clearly resolved intermittent recovery period around 45 days—the above mentioned ‘kink’. Velocity variations are weakest to the northeast of the fault in the area between the fault and the PFO site. The region that exhibits the maximum seismic velocity changes 30 to 60 days after the EMC event shows within the location uncertainties at day 110 a remarkable gap in the peak velocity change associated with the Collins Valley event. This can potentially be associated with evolving mechanical properties of the successively deformed material. However, this triangulation based detail should be verified in future analyses using updated pair-wise based inversion results.
5.2 Estimating the depth resolution

Estimating the depth of the medium changes is crucial for constraining the underlying source properties. For scattered wave propagation, the velocity structure and thus the surface wave sensitivity (Figure 8b) together with the medium heterogeneity are the two main properties that control the resolvable depth regime of the medium changes, and the time, frequency, and lapse time dependent \( \frac{dv}{v} \) observations in Figure 5. At lapse times up to eight times the transport mean free time \( t^* \) the coda wave sensitivity is dominated by surface wave properties [Obermann et al., 2016]. Later, body waves dominate. The transport mean free time \( t^* \) is proportional to the inverse frequency \( f \) dependent transport mean free path \( l^* = l^*/c_E \).

Depth dependent body wave sensitivity kernels (Figure 8b) can be computed using analytical solutions of scattered energy propagation [Pacheco and Snieder, 2005] that depend on \( \tau \) and the diffusion constant \( D = l^*c_E/3 \), where \( c_E \) is an effective energy velocity that depends on the energy ratio of \( S \)-waves and \( P \)-waves at the surface. Here we use \( 1/c_E = 0.89/v_S + 0.11/v_P \) based on a ratio obtained with numerical simulations of wave propagation in an attenuation-free 3-D heterogeneous elastic medium [Obermann et al., 2016].

It follows that the \( \frac{dv}{v} \) value estimated at a given frequency \( f \) and lapse time \( \tau \) in the coda is governed by the \( \tau/t^*(f) \) controlled partitioning \( \alpha \), the \( f \) dependent depth resolution of the involved surface waves, and the \( \tau \) and \( l^*(f) \) dependent body wave sensitivity. These connections can be used to invert \( \tau \) and \( f \) dependent \( \frac{dv}{v} \) estimates for the lateral and vertical position of the actual velocity changes in the medium \( \frac{dv}{v} \) [Obermann et al., 2018], understanding that \( \frac{dv}{v} \propto \alpha(\frac{dv}{v})_{sw} + (1 - \alpha)(\frac{dv}{v})_{bw} \). The \( (\frac{dv}{v})_{\langle \rangle} \) estimates based on surface waves (sw) and body waves (bw) depend on the respective kernels \( K_{\langle \rangle} \) via \( (\frac{dv}{v})_{\langle \rangle} \propto K_{\langle \rangle}/\tau \cdot \frac{dv}{v} \cdot \Delta V \), where \( \Delta V \) is a local variation in a voxel \( \Delta V \). However, as even solutions to the simpler lateral 2D problem have to await cleaner \( \frac{dv}{v} \) estimates (Section 5.1), we focus here on a discussion of the dominating regimes to evaluate the implications of the \( f \) dependence of the seasonal and post-earthquake \( \frac{dv}{v} \) amplitude pattern, the \( f \) dependence of the timing of the peak velocity change, and the \( \tau \) dependence of the amplitude pattern and the peak timing.

Hillers et al. [2013] parametrized the lateral heterogeneity in the study area from the tomographic model of Allam and Ben-Zion [2012] and estimated scattering angle dependent 20 km < \( l < 150 \) km for 0.5 Hz Rayleigh waves. Anache-Ménier et al. [2009] estimated \( l \) in the 1–10 km range for 5–7 Hz body waves in the PFO area. The ratio of \( l \) and \( l^* \) depends on the scattering anisotropy. Clearly, the length scale \( l^* \) after which all information about the orig-
inal direction of propagation is lost is larger than the average distance \( l \) between two scattering events. Typical ratios for anisotropic scattering are around 1.5. For the body wave kernels in Figure 8b we thus use \( l^* = 50 \text{ km} \) for the lower and \( l^* = 10 \text{ km} \) for the upper frequency range, and an effective velocity \( c_E = 3.6 \text{ km/s} \). The associated low- and high-frequency mean free times are thus around 14 s and 3 s, and the typical wavelengths associated with \( c_E \) [Obermann et al., 2016] are in the 12 km and 2 km range, respectively. For a window average lapse time of 35 s this results in \( \tau/t^* \) ratios of 2.5 and 13, which suggests the coda at low- and high-frequencies is dominated by surface waves and body waves, respectively.

The frequency and depth dependent surface wave kernels \( K_{sw} = \partial c/\partial v_s \) in Figure 8b [Herrmann, 2013] are based on profiles from the tomographic velocity model of Fang et al. [2016]. They indicate Rayleigh wave sensitivity above the proposed slow slip zone, and for most of the used frequencies it is above 5–8 km. We assume here that the wide frequency ranges are dominated by the low-frequency content considering the frequency dependent decrease in coherency on the flank of the microseisms peak and the faster attenuation of waves with shorter period. Alternatively, the surface wave sensitivity vanishes at depths larger than 2/3 of the typical wavelength [Obermann et al., 2016], which yields very similar estimates. The low frequency kernels from around the PFO or East area in Figure 8b imply a 20% smaller peak sensitivity compared to the \( K_{sw}(z) \) values estimated in the area to the west of the fault. The otherwise similar depth resolution suggests that the laterally different \( dv/v \) amplitudes and time histories are not governed by processes or effects in different depth ranges.

We use a 3-D sensitivity kernel for coincident source and receiver assuming body wave energy propagation can be modeled as a diffusion process [Pacheco and Snieder, 2005]. At a given depth the 3-D kernel is integrated over the horizontal domain to yield \( K_{bw}(z, \tau) \) and normalized by \( \tau \) to ensure \( \int K_{bw}dz = 1 \) [Sens-Schönfelder and Wegler, 2011; Hillers et al., 2014]. The kernels for \( l^* = 50 \text{ km} \) and \( l^* = 10 \text{ km} \), proxies for the low- and high-frequency regime, suggest that the scattered body waves can probe the medium response in the depth regime of the proposed creep event, whereas the surface wave sensitivity is constrained to the region above 10 km depth. Note, however, that \( \int K_{sw}dz \) is on the order of 0.1, i.e., the body wave sensitivity in Figure 8b appears exaggerated relative to the surface wave’s.

For the assessment of the frequency and lapse time patterns we focus on the high-SNR \( dv/v \) results obtained in the west and center zones. We begin with the inverse frequency dependent amplitude variation. At the low frequencies for which \( \tau < 8l^* \) and hence surface waves dominate, the decreasing \( K_{sw} \) values with depth for longer periods (Figure 8b) suggest...
together with the inverse frequency dependence of the measured $\delta v/v$ values that the medium change $\delta v/v$ is larger at depth. The continuation of this trend into the regime that is dominated by body waves is compatible with the frequency dependent properties of the body wave kernels, because $K_{bw}$ values at depths at and above the suggested slip zone at 15 km (gray line in Figure 8b) decrease, too, with decreasing frequency due to the increasing scattering length scale. Another change-in-regime-related factor that contributes to the decreasing $dv/v$ estimates is the overall smaller body wave sensitivity [Obermann et al., 2013] considering the $\int K dz$ scaling. The $dv/v$ frequency patterns and the properties of surface and body waves imply thus that the medium change and hence also the deformation at depth is larger compared to the near-surface layers. This behavior holds for the seasonal loading (Figures 4c, d) and the proposed slow slip response (Figure 5a).

Next we assess the frequency dependent timing of the peak $dv/v$ reduction. Combining the inverse frequency dependent depth scaling of the $\delta v/v$ changes with the timing of the peak $dv/v$ reduction (Figures 5a, blue circles in 8b) we infer a delayed arrival of the maximum deformation state towards shallower depths. This suggests an upward migration of the medium deformation. An extrapolation of this pattern towards greater depth is compatible with the deep slip hypothesis. We emphasize the reverse trend at higher frequencies. The smaller circles associated with higher frequencies in Figure 8b tend to indicate smaller delay times. This is perfectly compatible with the notion that the two highest frequency bands are dominated by body waves that sample the deeper parts where the deformation arrives earlier. The shallow location of the circles that follows the surface wave kernels had thus to be modified considering the partitioning of surface and body wave sensitivity.

We iterate that the late lapse time results confirm the spatially variable seismic velocity reduction in the Anza area following the EMC and Collins Valley events. The lapse time dependent $dv/v$ reduction and delayed peak timing in Figure 5b appears to be governed by properties of the surface wave field considering that $\tau/t^*$ ratios for the three considered $\tau$ windows in the low-frequency regime are in the range 2–4. In this case, we attribute the $\tau$ dependence to lateral averaging effects [Obermann et al., 2016]. The decreasing $dv/v$ amplitude data can also be influenced by an increasing body wave partition due to the smaller kernel values. A body wave dominance $\tau/t^* > 8$ for this lapse time and period range requires $l^*$ to be not larger than 20 km. Although this scenario cannot be ruled out considering the uncertainties in the $l$ estimates and $l/l^*$ relation, a body wave dominance is not expected because the $dv/v$ peak timing does not reverse the overall trend.
6 Strain observations

We compare records from the long base laser strainmeter at the PFO site (Figures 6a–c) to the $dv/v$ time series from the east zone (Figure 6d). The records of the N-S, E-W, and NW-SE strain components (Figure 1b), $\varepsilon_{yy}$, $\varepsilon_{xx}$, $\varepsilon_{nw}$, are hourly sampled. Tidal components are removed. Twenty-four hour averages yield time series that are sampled at the same daily resolution as the $dv/v$ data. Volumetric strain $\theta = \varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz}$ is estimated from $\varepsilon_{zz} = \xi(\varepsilon_{xx} + \varepsilon_{yy})$ at the surface of a traction free half-space, where $\xi = -\nu(1-\nu) = -1/3$, and $\nu = 1/4$ is Poisson’s ratio.

Figure 6a shows in gray the hourly sampled strain data in relation to the 40 day average before the EMC event. Positive values correspond to extension, i.e., proposed right lateral slow slip on the $\sim130^\circ$ trending fault yields large extensional strains in the E-W direction, large compressional strains in the N-S direction, and a smaller extension signal on the NW-SE component (Figure 1b). We compare the $dv/v$ time series to the strain data and strain rate estimates [Rivet et al., 2011]. For clean estimates we limit the analysis to 55-day and 40-day-long time windows that end after the occurrence of the peak strain amplitudes following the two earthquakes. The 55 days range after the EMC event stops just before the emergent onset in the $\varepsilon_{yy}$ and $\theta$ data (Figure 6a). This excursion seen on one strainmeter at 56 days has been traced to an instrumental problem. We then fit a fifth-order polynomial to these daily sampled strain data segments. These are the colored line segments in Figure 6a. Strain rate is obtained by differentiating these functions. The choice of the order five is not important for our conclusions. Figures 6b and 6c show the scaled absolute values of the obtained strain and strain rate functions. Strain on the individual components is maximum shortly before the end of the considered 55 days range after the EMC event. In contrast, volumetric strain peaks 30 days earlier due to the different $\varepsilon_{xx}$ and $\varepsilon_{yy}$ behavior and the opposite signs.

The $dv/v$ time series in Figure 6d are the results for the $0.3-1.5$ Hz (bottom, blue) and the $0.5-2.5$ Hz (top, lightblue) range using the $20-50$ s window, and the lapse time dependent $0.3-1.5$ Hz results ($20-40$ s darkred to $40-60$ s orange). The data and the colors correspond to Figure 5. We show the original data that have been smoothed in Figure 5 to highlight the daily resolution of the time series. The low-pass time series—again a polynomial fit using the same order five as for the strain analysis—should facilitate the comparison with the strain data. We emphasize that these east zone $dv/v$ results are characterized by a lower SNR compared to the west zone results (Figure 5).
The principal robust observation from the comparison of strain data (Figure 6a) and $dv/v$ time series (Figure 6d) is that two deformation episodes correspond to the proposed deep slip transients. In both cases the positive $\theta$ changes indicate volumetric extension that are associated with a reduction in seismic velocities followed by a recovery. That is, the timing and the mechanism are consistent across the sequence. This consistency advocates joint analyses of the two data types for improved detection capabilities and for improved monitoring and better resolution of slip episodes in active fault zone regions.

To determine further whether peak velocity reductions correspond better to the volumetric strain or the strain rate functions we indicate time windows around the maxima in the respective strain signals across Figures 6b–d using light blue and gray patches, respectively. The results differ between the EMC and the Collins Valley events responses. The maximum velocity reductions following the EMC earthquake fall mostly inside the peak volumetric strain range (blue patch). The exception to this rule is the dark red low frequency, early lapse time curve, where the minimum shows a better correlation with the strain rate. Note that the lapse time dependent delay in the timing of the peak reduction is consistent with the observations from the west and center zones discussed in Section 5.2.

The results associated with the Collins Valley event are clearer. Here, too, we discern a slight shift in the peak time with $\tau$. However, all obtained velocity change curves have a minimum early after the earthquake and the assumed onset of the slow slip event, and correlate thus systematically better with the period of high strain rate in the first 15 days. The clear upward trajectories of most curves indicate a recovery process during the light blue-colored high volumetric strain period around 30 days after the earthquake, which suggests a decreased sensitivity to this deformation mode compared to the preceding EMC event response.

7 Discussion

Using ambient noise data we resolve transient seismic velocity changes in the region around the Anza seismic gap in the central section of the San Jacinto fault zone. Analysis of wavefield and correlation quality markers indicate that the observed $dv/v$ variations in the 70 and 30 days following the El Mayor-Cucapah (EMC) and Collins Valley earthquakes are not governed by aftershock induced changes in the noise properties. Constraints on the lateral velocity change distribution are provided by the triangulation approach. The most striking observation is the significantly different response in the area to the west of the fault compared to the area to the east between the fault and the high-plateau hosting the PFO strainmeter. The
overall consistency of the $dv/v$ amplitude patterns in response to seasonal loading and to the
proposed creep episodes highlights the mechanical role of the PFO batholith and the material
and topographic variations across the Anza seismic gap region.

The velocity change episodes are overall compatible with the earthquake interaction sce-
nario proposed by Inbal et al. [2017] based on strain data from six PBO stations and local seis-
micity patterns. According to this scenario the EMC earthquake triggered aseismic slip on the
San Jacinto fault near Anza. Data from the first 10 days after the EMC event constrained the
triggered aseismic slip primarily to depths larger than 10 km. The accumulated slip corresponds
to a moment magnitude of 5.9. The associated stressing of the surrounding fault segments is
thought to have contributed to the nucleation of the Collins Valley event, which triggered an-
other aseismic slip episode with moment magnitude 5.8. The microseismicity that was used
to constrain properties of the slow slip episodes and the creep events were assumed to occur
on the primary San Jacinto fault interface. Follow-up work indicates that small events in the
area nucleate primarily off-fault while moderate and large events involve slip on the main faults
that tend to dip to the northeast below $\sim$10 km depth [Ross et al., 2017; Trugman and Shearer,
2017; Cheng et al., 2018]. The distribution of the microseismicity in broad damage zones also
likely explains the scarcity of repeating events [Jiang and Fialko, 2016] that occur on asper-
ities situated in creeping regions along more mature fault zone segments such as the central
San Andreas fault [Khoshmanesh et al., 2015]. Approximating the volumetric deformation pro-
cesses to occur on a single plane may lead to erroneous inferences concerning the duration,
size, or position of the slip events. In any case, the existence of tens-of-days-long deforma-
tion episodes in the focus area following the EMC and the Collins Valley earthquakes is sub-
stantially supported by the consistency of our seismic observations and the strain signals at
PFO and the PBO stations [Inbal et al., 2017]. We thus continue to work with the hypothe-
sis that the observed $dv/v$ variations reflect the spatio-temporal medium response to creep episodes.

Inbal et al. [2017] inferred three main slow slip patches on the San Jacinto fault in re-
sponse to the EMC event. These are visible from the along-fault amplitude distributions in Fig-
ure 8a that are separated in a shallow (<8 km) and deep (>8 km) average. The coincidence
of the obtained $dv/v$ pattern to the west of the fault and the deep profile is highlighted by the
compatibility of the along-strike position of the maximum velocity reduction with the posi-
tion of the largest slip at depth. Faint shallow creep is confined to a small region in the cen-
ter. Nevertheless, the distribution of the small amplitudes along the seismic gap section co-
incides with the position of red $dv/v$ areas next to the fault. These relations are another strong
indicator of a causal relationship between the velocity change and the proposed slow slip phenomena. Evidence in favor of the significance of deep over shallow slip is compatible with the overall absence of shallow creep on the Anza section [Lindsey et al., 2014], which is again in contrast to the behavior of segments along more mature faults such as the San Andreas fault [Khoshmanesh et al., 2015], the North Anatolian fault [Rousset et al., 2016], or the Haiyuan fault [Jolivet et al., 2015] that all exhibit shallow creep.

The duration and amplitude of the velocity change associated with the proposed creep event triggered by the Collins Valley earthquake (Figure 7) are significantly shorter and smaller compared to the changes after the EMC earthquake, although the cumulative moment magnitudes and the slip patch locations and sizes are overall similar [Inbal et al., 2017]. The different \( \frac{\delta v}{v} \) patterns suggest an evolution of the material properties in response to successive deformation episodes.

Relating the spatial velocity change distributions to the tomographic images of seismic velocities or velocity ratios obtained by Allam and Ben-Zion [2012], Zigone et al. [2015], or Fang et al. [2016] is complicated by the depth variations of the imaged properties. Are the lateral \( \frac{\delta v}{v} \) variations shown in Figures 4f, 7, and 8a controlled by the response at depths above 10 km where the sensitivity is maximum, or by the material properties below 10 km around the proposed deep slip? A visual comparison of our velocity change patterns and the images by Fang et al. [2016] reveals that the region of the largest \( \frac{\delta v}{v} \) amplitudes to the west of the fault coincides with an area of reduced \( v_P \) that is bound by material with larger velocities in the fault-parallel northwest direction. This pattern on the west side of the fault is seen over the entire 3–16 km depth range. Systematically more competent material to the east that correlates with the clear change in the response across the seismic gap can only be inferred in the top 7 km from tomographic images [Fang et al., 2016]. The fault-normal profile 6 of Allam and Ben-Zion [2012] at the seismic gap indicates high (low) \( v_P/v_S \) ratios to the west (east) of the fault in the top 5–7 km. The indicated mechanical differences can also contribute to the observed variations, linking relatively smaller shear wave speeds to an increased susceptibility. These conclusions are supported by the coincidence of the fault perpendicular shear modulus \( \mu \) distribution across the Anza segment that Lindsey et al. [2014] compiled from the tomographic model of Allam and Ben-Zion [2012]. It shows systematically reduced values of \( \mu \) to the west of the fault in the top 5–10 km where we observe the largest negative \( \frac{\delta v}{v} \) amplitudes, and high \( \mu \) values up to the surface in the area to the east where our \( \frac{\delta v}{v} \) signals are smallest. Recall that the implied smaller local \( \frac{\delta v}{v} \) values to the east are sensed by
low-frequency surface waves that are characterized by a 20% smaller sensitivity $K_{sw}$ in this area, leading to the observed small $dv/v$ values. The $\mu$ profile indicates an extensive compliant fault zone with reduced elastic shear modulus several kilometers deep and away from the fault that is consistent with trapped waves observations in the area [Li and Vernon, 2001; Lewis et al., 2005]. The position of the peak-$dv/v$ reductions away from this densely instrumented zone supports the notion that the observations are not dominated by shallow fault processes.

Our observations provide a direct comparison between regional tomographic images, inferred variations in the rock type and competence, and deformation amplitudes.

Brenguier et al. [2014] found significantly above-average velocity changes after the 2011 Tohoku earthquake in highly susceptible volcanic regions across Japan. The existence of hot springs in the study area (Figure 1b) implies the possibility that the low $\mu$ values and the high susceptibility leading to the observed large $dv/v$ amplitudes in the area between the San Jacinto and the Elsinore fault are associated with elevated water saturation.

The $dv/v$ results for several years show that the transients in 2010 are not governed by some systematic annually repeating environmental effect. Considering the large seasonal variation, the filter employed [Moreau et al., 2017] and the MWCS based inversion approach [Brenguier et al., 2014] are crucial for the resolution of the transient signals in 2010. This focus on resolution comes at the cost of outstripping the available resources for the computation of equivalent multi-year time series for potentially better filtering, fitting, or decomposition [Hobiger et al., 2012; Wang et al., 2017].

A comparison of PFO strainmeter and east zone $dv/v$ data (Figure 6) reveals a sequence of volumetric extension transients following the EMC and the Collins Valley events that correlate with episodes of reduced seismic velocities. Maximum velocity reductions after the Collins Valley event coincide better with peak strain rates compared to periods of elevated strain. The behavior following the EMC event is more ambiguous which is likely related to the fluctuations that bias the resolution of the transient onset. The strain rate related observations are compatible with the seismic velocity change pattern in the middle crust induced by the 2006 M7.5 slow slip event on the Guerrero segment of the Mexican subduction zone [Rivet et al., 2011], where the $dv/v$ signal also closely follows the inferred strain rate. The compatibility of the PFO strain rate signals and the $dv/v$ observations in the east zone suggests that nonlinear failure and recovery mechanics play a role in the observed behavior [Johnson and Jia, 2005; Lyakhovsky et al., 2009].
Overall, the consistent timing of proposed creep events, extensional strain, and reduced seismic velocities demonstrate the feasibility of seismic velocity monitoring along a hazardous fault segment to complement seismicity and geodetic observation techniques for improved resolution of the fault slip and deformation spectrum, and for in-situ probing of stress redistribution and relaxation processes. The quality and resolution of our results suggest that time-dependent material properties and key geometrical and kinematic source parameters of slow slip episodes can be estimated in future source inversions based on spatio-temporal $dv/v$ variations.

8 Conclusions

Our analysis of continuous seismic data from a regional network around the Anza seismic gap of the San Jacinto fault yields first-time observations of a complex pattern of episodic, tens-of-days-long seismic velocity variations $dv/v$ along a continental strike-slip fault. The studied section of the San Jacinto fault exhibits a relatively shallow geodetic locking depth around 10 km [Lindsey et al., 2014; Jiang and Fialko, 2016], compared to the 14–18 km maximum depth of the background seismicity that occurs predominantly off the main faults [Ross et al., 2017; Trugman and Shearer, 2017]. The potential of intermittent creep bursts in the transition zone below the locked fault [Wdowinski, 2009] is suggested by unusually elongated aftershock zones around moderate events [Meng and Peng, 2016; Ross et al., 2017]. The spatial and temporal velocity variations observed in this work together with the long base strainmeter signals likely reflect strain transients that follow two deep creep episodes that are triggered by the remote EMC and the local Collins Valley earthquakes [Inbal et al., 2017]. Our data suggest evolving nonlinear medium response because properties of the inferred slip events are more similar to each other than the amplitude and duration of the associated $dv/v$ patterns imply. The results extend observations related to interaction mechanisms between different slip and deformation modes along faults that are typically obtained from satellite, geodetic, and seismicity data [Linde et al., 1996; Donnellan et al., 2014; Khoshmanesh et al., 2015; Jolivet et al., 2015; Rousset et al., 2016; Meng and Peng, 2016; Inbal et al., 2017; Khoshmanesh and Shirzaei, 2018]. Noise based estimates of the induced velocity changes complement the array of observation techniques that can be used to study the coupling behavior along strike and down dip the seismogenic zone and relaxation processes in the surrounding crust. Future joint inversions of GPS, satellite, seismic velocity change, and seismicity and tremor data for key parameters of deformation events can improve the constraints on evolving crustal and fault rheology and provide
new insights on fault and earthquake interactions, and triggering, stress redistribution, and feedback processes between stable and unstable slip modes.

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Hauksson, E., W. Yang, and P. M. Shearer (2012), Waveform Relocated Earthquake Cat-


Figure 1. (a) Map of the study area. Grey circles indicate 2010 seismicity. Red circles indicate seismicity outside the two blue boxes in the five days after the El Mayor-Cucapah (EMC) event that lead to the spike at $t_0$, the day of the EMC event, in the red data in Figure 2b. Stars indicate $M \geq 4$ events. Their occurrence is indicated in Figures 2–5 with the correspondingly colored vertical lines. Blue, cyan, and red stars indicate the M7.2 EMC, M5.7 Ocotillo, and M5.4 Collins Valley earthquakes. Black lines are mapped fault traces. The black lines and crosses in the EMC box show the trace and grid points of the EMC finite source model [Wei et al., 2011b; Mai and Thingbaijam, 2014]. The ellipse around the Ocotillo event indicates the Yuha Desert region. Shallow slip has been triggered on faults in the orange areas. Deep slow slip has been inferred along the yellow indicated fault segment that extends to the pink star. The green patch indicates the Salton Sea Geothermal Field. SJFZ: San Jacinto fault zone. SAF: San Andreas fault. EF: Elsinore fault. ASG: Anza seismic gap. (b) Enlarged study area. Black and blue triangles indicate surface and borehole stations. Orange circles indicate seismicity in the SJF box in (a) in the five days after the EMC event that lead to the spike at $t_0$ in the black data in Figure 2b. PFO, KNW, and PMD are stations referred to in the text. The tripod indicates the configuration of the long base strainmeter at the PFO site. Lightblue dots indicate hot springs. TA: Trifurcation area.
Figure 2. (a) Temporal evolution of the magnitude dependent seismicity in the EMC box in Figure 1a. The ordinate is clipped. (b) Black data are events in the SJF box in Figure 1a. Red data correspond to the gray and red seismicity in Figure 1a that is not in the SJF box and not in the EMC box. The histograms in (a) and (b) show the number of events in 8-hour intervals. (c) $H$ and $V$ energy estimates in three frequency bands. Thin lines are 3-point median filtered 8-hour data. Thick lines are 5-day running averages. (d) $H/V$ energy ratios in 8-hour intervals in three frequency bands. Black lines are the network median, the gray lines are the 0.2 and 0.8 quantiles. In (c) and (d) data are on the same scale but offset for clarity. The dashed lines indicate the median level in the 50 days prior to $t_0$. Below average levels at later times reflect a seasonal excitation effect. The vertical lines indicate the occurrence of the correspondingly colored events in Figure 1a.
Figure 3. (a) Scaled $P$ values of stretching results obtained with the average-waveform reference approach. Lower $P$ levels prior to $t_0$, and the higher levels after 100 days are a seasonal effect. Black and orange lines correspond to the cases in (b) and (c). (b) Daily histograms of network and component average $dv/v$ estimates. The thin black line is the amplified thick-line average. (c) The same as in (b) for Wiener-filtered correlations. (d)–(f) The same as in (a)–(c) using the MWCS technique. (g) Statistics of how often a $dv/v$ estimate based on a correlation function from a given day is used in the inversion. These data show how many $dv/v$ pairs including the correlations functions from a given day have $cc > 0.85$. Black and orange lines correspond to the cases in (h) and (i). Data are scaled by the peak value of the Wiener-filtered data. (h) Inversion based network and component average $dv/v$ estimates using the MWCS technique (black). Gray data indicate network-average results obtained with any one of the nine component pairs. (i) The same as (h) using Wiener-filtered correlations. The vertical lines indicate the occurrence of the correspondingly colored events in Figure 1a. All results are from nine-component correlations, $0.2 - 2$ Hz.
Figure 4. (a) Network average 300-day seismic velocity changes in the 0.2 — 1 Hz range for the years 2008, 2009, 2011—2015 (multi-year, gray). The long red line is the corresponding 2010 result. The short red line is the 2010 result obtained with the nine-component correlation data. (b) Same as in (a) for stations in the west zone for 0.3 — 1.5 Hz data. The black line indicates the seasonal average that is subtracted from the red 2010 data to yield the blue deseasonalized 2010 results. Frequency dependent multi-year results for the west zone (c), center zone (d), and east zone (e). The associated group of stations is shown in the insets. (f) Spatial distribution of the multi-year average 0.3 — 1.5 Hz peak-to-peak seasonal $dv/v$ changes.
Figure 5. Frequency dependent (a) and lapse time dependent (b) temporal evolution of the $dv/v$ changes for the three west, center, east zones indicated in Figures 4c–e. The lapse time window for the results in (a) is 20 − 50 s. The frequency range used in (b) is 0.3 − 1.5 Hz. The vertical lines indicate the occurrence of the correspondingly colored events in Figure 1a. The daily sampled time series are smoothed with a 3-point running average.
Figure 6. (a) Records from the PFO laser strainmeter (gray). Colored line segments are fitted polynomials. \( \varepsilon \) indicates horizontal strain components, and \( \theta \) denotes volumetric strain. (b) Normalized fits to the data in (a). (c) Differentiated and normalized data from (b). (d) Relative velocity change data from Figure 5. Colored lines are the daily resolved \( dv/v \) curves, and the gray lines are polynomial fits to illustrate trends. Colors are the same as in Figure 5. Lightblue: 0.5 – 2.5 Hz, 20 – 50 s; orange: 0.3 – 1.5 Hz, 40 – 60 s; red: 30 – 50 s; darkred: 20 – 40 s; blue: 20 – 50 s. All data are from the east zone and on the same scale but offset for clarity. Vertical lines indicate the occurrence of the correspondingly colored earthquakes in Figure 1a. Grey and lightblue patches indicate periods associated with peak strain rate and peak strain.
Figure 7. Temporal evolution of the $dv/v$ change pattern obtained from 0.3 – 1.5 Hz station-triplet results. Numbers in the upper left denote days since the El Mayor-Cucapah event. Circles mark station locations. Dots in the first frame indicate triangle center locations.
Figure 8. (a) Spatial distribution of the peak $dv/v$ reduction during the 80 days after the El Mayor-Cucapah (EMC) earthquake. Gray lines indicate mapped faults. Gray circles show seismicity in the 80 days after the EMC earthquake. The Anza seismic gap is clearly visible. The crosses denote the sites for which the $K_{sw}$ kernels in (b) have been estimated. The black dots indicate the fault parametrization of Inbal et al. [2017]. The black-and-white lines show the cumulative slip distribution of the post-EMC asismatic event in two depth sections along this track. Smaller (larger) amplitudes correspond to slip above (below) 8 km depth. (b) The gray line is the cumulative slip distribution along the Anza seismic gap section. Orange lines are Rayleigh wave kernels $K_{sw}$ from the two sites to the west and east of the fault for two frequencies. The red lines are the body wave kernels $K_{bw}/\tau$ with units 1/km. The solid (dashed) lines correspond to the low (high) frequency solutions. For $K_{sw}$: $f = 0.2$ Hz and $f = 0.6$ Hz. For $K_{bw}$: $l^* = 50$ km and $l^* = 10$ km. $K_{sw}$ and $K_{bw}$ are on the same scale. Blue circles indicate the timing of the peak $dv/v$ reduction in the two areas (Figure 5), and the circle size is proportional to the $dv/v$ amplitude. The vertical position of a circle is associated with the $K_{sw}$ kernel-peak of the lower bound of the frequency range.