RESEARCH ARTICLE

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Key Points:

- Strain and triggered seismicity in Anza are spatiotemporally correlated
- The M_w 7.2 El Mayor-Cucapah main shock triggered a several monthlong transient in Anza
- The M_w 5.4 Collins Valley main shock was most likely triggered by aseismic slip induced by the remote El Mayor-Cucapah main shock

Supporting Information:

- Supporting Information S1

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Locally and remotely triggered aseismic slip on the central San Jacinto Fault near Anza, CA, from joint inversion of seismicity and strainmeter data

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Abstract We study deep aseismic slip along the central section of the San Jacinto Fault, near the Anza Seismic Gap, in southern California. Elevated strain rates following the remote M_w 7.2, 4 April 2010 El Mayor-Cucapah and the local M_w 5.4, 7 July 2010 Collins Valley earthquakes were recorded by Plate Boundary Observatory borehole strainmeters near Anza and were accompanied by vigorous aftershock sequences. We introduce a method to infer the distribution of triggered aseismic slip from combined seismicity and geodetic data, based on a rate-and-state friction model that maps observed changes in seismicity rates into stress changes. We invert for the cumulative slip in the 10 day period following each main shock. Synthetic tests show that the effect of aftershock interactions on the inferred slip distribution is negligible. The joint data set is more consistent with a model in which aseismic slip on a principal fault triggers seismicity on adjacent faults than with one in which aseismic slip and seismicity are coplanar. Our results indicate that aseismic slip primarily occurs along the rim of two seismicity clusters adjacent to Anza Gap, as well as beneath the Anza Gap itself, at depths larger than 10 km. The triggered aseismic slip generated by the two main shocks has little overlap, a pattern also found in sequences of large earthquakes occurring on a same fault. Stresses inferred from seismic activity leading to the Collins Valley main shock suggest that this earthquake was triggered by stresses imposed by the El Mayor-Cucapah remote-triggered aseismic slip, which persisted for more than 2 months.

1. Introduction

Continental strike-slip faults are mostly seismically active above 10 to 15 km depth and exhibit predominantly aseismic behavior at larger depths. Due to the limited sensitivity of surface monitoring systems, the mechanical properties of the deep seismic-aseismic transition zone are not well resolved. One approach for probing the downward extent of crustal faults is to analyze the accelerated deformation following large main shocks, which is typically characterized by rapid afterslip and numerous aftershocks. It has been long recognized that aftershocks can be driven by the stressing imposed by aseismic slip [e.g., Perfettini and Avouac, 2004; Lengliné and Marsan, 2009; Hsu et al., 2006; Pritchard and Simons, 2006; Chlieh et al., 2007; Perfettini and Avouac, 2007; Ozawa et al., 2012; Lengliné et al., 2012]. However, a joint analysis of the two phenomena is challenging, mainly because it requires a sensitive geodetic network capable of detecting deformations that marginally exceed the signal due to nontectonic sources, as well as a complete and accurate seismicity catalog. Designing an approach that will simultaneously satisfy both the geodetic and aftershock data sets is also complicated by the fact that aftershocks trigger their own aftershock sequences and redistribute stresses in their vicinity. Since the relative importance of earthquake interaction and aseismic fault slip for triggering of seismicity is unknown and since aseismic slip and aftershock rates show similar temporal decay [e.g., Perfettini and Avouac, 2004; Wang et al., 2010], separating their effects in the observed seismicity is difficult.

Details of the space-time evolution of seismicity are therefore rarely used to constrain geodetic slip inversions. Instead, they have been used as a posterior test of consistency of the inversion solution with the notion that earthquakes are triggered at locations of increased Coulomb stress [e.g., King et al., 1994]. The results of such tests are not always positive. For example, Ziv [2012] found that most of the aftershocks in the first day following the 2004 M_w 6.0 Parkfield earthquake occurred in areas where a geodetic-only slip inversion, which fits well the 1 day static GPS displacement data, predicted reduced Coulomb stresses.

Figure 1. Location map. Triangles and squares indicate seismic stations and PBO strainmeters, respectively. Black lines are fault traces. The red dashed line (A-A') indicates our assumed fault model. Red, green, and white stars indicate the locations of the M_w 5.4 Collins Valley main shock, the two $M > 4$ earthquakes of 13 June 2010, and the M_w 7.2 El Mayor-Cucapah main shock, respectively. Blue curves outline the polygons considered in the relocation procedure. HSC: Hot Springs Cluster, TC: Trifurcation Cluster, AG: Anza Gap.

Some authors have explored the possibility of using aftershocks to constrain the slip distribution under the assumption that the stress change due to slip is the dominant mechanism for aftershock triggering. One possibility is to optimize the correlation between Coulomb stress increase and aftershock locations [Seeber and Armbruster, 2000; Bennington et al., 2011; Wang et al., 2012]. Another approach is to relate observed variations in seismicity rates to variations in stressing rate via the seismicity evolution model introduced by Dieterich [1994], which incorporates time-dependent earthquake nucleation and laboratory-motivated rate-and-state friction. This approach was applied, for example, to the 1992 Landers aftershock sequence [Gross and Kisslinger, 1997], to infer stresses associated with volcanic deformation [Dieterich et al., 2000], aseismic transients [Lohman and McGuire, 2007], dike intrusions [Segall et al., 2006], and loading due to rain, thermoelastic strain, and solid earth tides [Hainzl et al., 2013]. Inversely, Dieterich [1994] seismicity model can be used to image the deformation inducing seismicity. In this approach, inferred stresses and observed surface displacements are jointly

inverted in an attempt to recover deformation at depth. This approach has been used in only a few studies. Ziv [2012] imaged the spatial distribution of coseismic (main shock) slip by joint inversion of geodetic data and first day aftershock rates. Segall et al. [2013] introduced a time-dependent scheme for jointly analyzing stresses and displacements that resulted from dike intrusions. The main advantage of these approaches over the geodetic-only approach is the substantial increase in the sensitivity to deep fault slip [Ziv, 2012]. Similar to the study of Ziv [2012], our formulation of the seismicity evolution model establishes a linear relation between logarithmic changes in seismicity rates and applied stresses, which allows us to apply a linear inversion scheme to the joint data set. This differs from the formulation presented by Segall et al. [2013], in which earthquake counts are a nonlinear function of deformation-induced stresses. Both the approach presented here and the approach taken by Segall et al. [2013] are suitable for studying long-term aseismic transients occurring over timescales of weeks.

We analyze transient slip events and seismicity along the central San Jacinto Fault (SJF) near Anza, southern California, following the regional 2010 M_w 7.2 El Mayor-Cucapah and the local 2010 M_w 5.4 Collins Valley main shocks. Both earthquakes were followed by numerous aftershocks near Anza and by deformation transients well recorded by Plate Boundary Observatory (PBO) borehole strainmeters operative since 2008 (Figures 1 and 2). Dynamic triggering of shallow creep has been observed on several fault segments following the El Mayor-Cucapah earthquake [Wei et al., 2011, 2015]. In contrast, the Anza segment does not exhibit any detectable shallow creep [Lindsey et al., 2014]. Thus, the strain transients there interpreted as resulting primarily from deep-triggered creep on the SJF. Our interpretation is based on strain changes that are coherent across the Anza PBO network and that can be correlated with observed changes in the spatiotemporal distribution of seismicity (Figure 2). Following the approach of Dieterich et al. [2000], we map the observed seismicity rates to stress changes on secondary faults near the main SJF strand. We use the cumulative postseismic stresses and strains to jointly invert for the distribution of triggered aseismic slip in the first 10 days following each main shock and to constrain the mode of stress transfer between creeping fault segments to seismically active clusters.

Figure 2. Strain data and pre-, co-, and post-El Mayor-Cucapah relocated SJF seismicity. (top) Cumulative number of events (solid line) and earthquake rate (vertical bars) as a function of time during 2010. Red, blue, and green dashed curves indicate the time of the $M_w 7.2$ El Mayor-Cucapah, $M_w 5.4$ Collins Valley, and two $M > 4$ foreshocks, respectively. Grey bands indicate periods of analysis around both main shocks shown in the rows below. (bottom) Left and right columns are strain data from the El Mayor-Cucapah and Collins Valley sequences, respectively. Second, third and fourth rows show the differential extensional, areal and shear strain components, respectively. Horizontal dashed lines indicate preseismic and postseismic mean strain levels. Vertical bars indicate cumulative postseismic strains at the end of the analysis periods (grey bands) and their respective uncertainties.

The remainder of this manuscript is organized as follows. The seismotectonic setting of the study region is presented in section 2. The strain and aftershock data sets and data processing are described in section 3. An overview of the space-time patterns of seismicity and strain is provided in section 4. The joint inversion approach is introduced, and the inversion results are presented in section 5. Key aspects of the results are discussed in section 6. The evolution of seismicity and stresses leading to the Collins Valley earthquake is analyzed in section 6.1. The relative importance of earthquake interaction and secondary aftershocks is assessed in sections 6.2 and 6.3. The ratio of seismic to aseismic strain release during transient fault behavior in Anza is discussed in section 6.4. The results of joint inversions under the alternative assumption that triggered aseismic slip and aftershocks occur on the same main SJF strand, as in models of asperities embedded in creeping faults commonly invoked to explain repeating earthquake sequences [Nadeau and McEvilly, 1999] are reported in Appendix B. This approach was found less effective at reconciling the two data sets than the one presented in the main text.

2. Seismotectonic Background

The SJF is the most seismically active fault in the Southern San Andreas Fault system. Between its northernmost section near San Bernardino to its southern termination near Borrego Springs, this 250 km long fault zone may be roughly divided into three sections, the most active of which is the central section near Anza. Long-term deformation along that section is mainly accommodated by the Clark Fault (Figure 1), whose geologic slip rate is 10–15 mm/yr [Rockwell et al., 1990; Salisbury et al., 2012]. Southeast of Anza, the main strand of the Clark Fault branches into several active faults. The transition is characterized by a 25 km long section almost devoid of seismicity, which is known as the Anza Seismic Gap [Sanders and Kanamori, 1984] (AG). It is also marked by the strongest velocity anomaly along the central SJF, expressed by a 5 km wide region of low V_p/V_s ratio that extends to a depth of about 8 km [Allam and Ben-Zion, 2012; Allam et al., 2014], and a well-developed, ~400 m thick damage zone which is well resolved in seismic data sets [Yang et al., 2014; Zigone et al., 2015]. Although nine $M > 6$ events occurred along the central SJF in the past 120 years [Sanders and Kanamori, 1984], the AG has not been ruptured by any $M > 7$ earthquake in over 200 years [Salisbury et al., 2012; Rockwell et al., 2015]. With a contemporary slip rate of about 19 mm/yr [Lindsey and Fialko, 2013], the 25 km long Anza section may be capable of producing events with $M > 6$ in the future, thus posing a major threat to nearby communities.

The interseismic strain accumulation along the central SJF is characterized by strong fault-normal gradients. To fit such high strain rates, geodetic inversions assuming a dislocation model require the Anza segment to be locked from the surface to a relatively shallow depth of 10.4 ± 1.3 km [Lindsey et al., 2014]. In contrast, the maximum depth of seismicity, which we define as the depth above which 95% of the earthquakes occur, is 16.5 km [Smith-Konter et al., 2011]. A possible mechanism that may contribute to the high strain rates observed at the surface, and thus to reconcile the seismic and geodetic observations in Anza, is transient deep fault creep [Wdowinski, 2009]. According to this view, the deep transition between fully locked and steady slipping portions of the fault consists of aseismically creeping patches, capable of sustaining transients, interspersed by seismogenic asperities whose dynamic failure results in microseismicity. Physical models of earthquake behavior suggest that the topology of the asperities and the heterogeneity of frictional properties within the transition zone account for the statistics of the earthquake catalog [Dublanche et al., 2013a, 2013b; Inbal et al., 2016]. As has been observed in a number of subduction zones [Schwartz and Rokosky, 2007], aseismic release of strain accumulated in the transition zone beneath the SJF could manifest itself by deep intermittent creep and tectonic tremors. Since the sensitivity of surface deformation to deep slip is small, it is expected that tectonic tremors, if they occur, will provide useful constraints on transient deformation in the transition zone.

Moderate ($M > 5$) events along the Anza segment tend to nucleate at the base of the seismogenic zone and are accompanied by numerous aftershocks located as far as 50 km from the main shock. The large, several weeklong increase in seismicity rates at remote sites suggests that physical processes promoting long-range earthquake interactions may be operating along this fault. Felzer and Kilb [2009] studied aftershock sequences triggered by two $M \approx 5$ Anza main shocks that occurred in 2001 and 2005. They concluded that the distribution of aftershocks density as a function of distance to these main shocks followed the typical inverse power law decay observed in southern California and argued that aseismic slip is not required to explain the spatial extent of seismicity. There are, however, other observations that point to the possibility of aseismic transients. Meng and Peng [2016] conducted a systematic study of 10 aftershock sequences triggered by $M \approx 4$ –5 events that occurred between 2001 and 2013 near Anza. They found that the size of the aftershock zone scales with

the main shock depth and that the aftershock expansion rate is logarithmic and concluded that aseismic creep at the base of the seismogenic zone is driving deep aftershock expansion in Anza. Several of these events have also caused an increase in local strain rates identified by a long-baseline strainmeter installed in the Piñon Flats Observatory [Agnew and Wyatt, 2005; Agnew *et al.*, 2013]. This observation suggests that creep and seismicity are indeed spatiotemporally correlated. Similar to the slip transients near the Parkfield and San Juan Bautista sections of the San Andreas Fault [e.g., Linde *et al.*, 1996; Murray and Segall, 2005], the Anza creep transients occur near the edges of a locked fault segment (the AG). However, unlike the Parkfield segment, where the fault sections adjacent to the locked segment creep at the plate rate near the surface, the segments bordering the AG do not exhibit shallow fault creep [Lindsey *et al.*, 2014]. This suggests that some fraction of strain accumulated along the deeper portions of the Anza section is released in deep episodic creep events. In this study, we test this hypothesis by performing a joint analysis of strain and seismicity data. Next, we present the geodetic and seismicity data sets and analyze their spatiotemporal distribution during the 2010 El Mayor-Cucapah and the 2010 Collins Valley aftershock sequences.

3. Data

3.1. Strain Data

We analyze continuous strain measurements from six out of eight PBO borehole strainmeters located in Anza (Figure 1). PBO stations B082 and B089 are omitted due to poor signal-to-noise ratio caused by a nearby water pumping station. The strainmeters consist of four collocated horizontal extensometers, which were installed at depths of 140 to 240 m. The data provided by PBO are downsampled to a rate of one sample per 5 min, calibrated, corrected for instrumental noise, and converted into areal and shear strain components. The transformation matrices from the measurements at each extensometer to the horizontal components of the strain matrix are routinely computed from the predicted tidal response [Hodgkinson *et al.*, 2013]. Since the analysis yields only three components of the strain field, the problem is overdetermined and may be solved using measurements at only three extensometers. For most of the analyzed time interval data from all four extensometers are available. When only three extensometers are operative, we convert the raw data to the areal and shear strains using a three-gauge calibration matrix (K. Hodgkinson, personal communication, 2014).

Recently, several studies pointed to potential errors associated with the response of PBO strainmeter instruments. Langbein [2015] found that the coseismic response of stations installed near the San Francisco Bay to the *M*6.4 South Napa earthquake deviates significantly from model predictions. He attributes the discrepancies mainly to the influence of tidal model errors on the strain calibration scheme. As was noted by Hodgkinson *et al.* [2013], the Anza network is sufficiently removed from the coast so that tidal model errors should exert little influence on the calibration matrices. Barbour *et al.* [2015] also found poor agreement between observed and predicted coseismic offsets in Anza PBO strainmeter data during the El Mayor-Cucapah earthquake and better agreement for the Collins Valley earthquake. These discrepancies were mainly attributed to mechanical shaking induced by seismic waves and thus should have negligible effect on the postseismic records we analyze here.

Strain measurements are contaminated by various sources, which include barometric effects, ocean and solid earth tides, and borehole deformation. The contribution of the latter appears as a linear trend in our records. The signal-to-noise ratio at several epochs is low such that the tectonic signal is obscured by the ambient noise. To recover the tectonic signal, we adopt the approach of Hawthorne and Rubin [2010] and estimate empirical correction terms from time intervals around the period of interest. For each station and for each strain component, the time series can be approximated by the sum of the tectonic offset and the contribution from borehole deformation, barometric effects, tidal loading, and noise

$$\epsilon(t) = c_1 + c_2 t + c_3 p(t) + c_4 F(t - t_{\text{tra}}) + \sum_k^{17} m_k \cos(2\pi t/T_k) + l_k \sin(2\pi t/T_k) + c_5 H(t - t_{\text{ms}}) + n(t), \quad (1)$$

where p is the recorded barometric pressure, n is measurement noise, and t_{ms} and t_{tra} are main shock time and the end time of the transient of interest, respectively. The constants m_k and l_k are associated with a tidal signal composed of periods listed in Table S1 (supporting information). The fourth term in equation (1) accounts for finite strain accumulated due to postseismic deformation, which contributes to a total offset, c_4 , that we model in the inversion procedure. The transient time function F is such that $F(t) = 0$ if $t < t_{\text{ms}}$ and $F(t) = 1$ if $t > t_{\text{tra}}$ and has arbitrary shape in the interval $[t_{\text{ms}}, t_{\text{tra}}]$. The term associated to a step function $H(t)$ represents the coseismic strain of the main shock.

We apply the following processing steps to the data from each station and strain component. We first estimate the coseismic offset, c_5 , which appears as a step in the strain records, by removing from the postseismic strain records the difference between averaged strains in 1 h intervals immediately before and after the main shock. We then determine t_{tra} by visual inspection, as an interval in which strain rates in most of the stations have returned to premain shock levels. Next, we fit equation (1) to the data in two 10 day long windows, one before t_{ms} and one after t_{tra} . We finally evaluate the correction terms (i.e., all but the c_4 term) for the entire time period, subtract them from the observations, and smooth the results using a 24 h median window (T_{smooth} in Table S1) to obtain the processed strain time series shown in Figure 2. We found that records obtained via this empirical approach contain less noise than the archived PBO processed strain data. Several of the processed strain time series contain steps and short-term ramps, which result in strains that are comparable to the offset during the transients (e.g., sites B087 and B093 for the El Mayor-Cucapah and Collins Valley episodes, respectively). To minimize the effect of unmodeled noise on the inversion procedure, we use as data for our inversion the cumulative transient strain defined on each channel as the difference between 3 day averages before and after the transient window (dashed lines in Figure 2).

We adopt an empirical approach for estimating the noise term $n(t)$. We select one hundred 30 days long segments that did not contain any detectable transients. We process these data in the same fashion as for the windows containing the transients, after setting the value of the fitting coefficients c_4 and c_5 equal to zero. We filter the residuals between the observed and modeled strains with a 3 day median window. For each 30 day segment of smoothed residuals, we compute the difference between all data points separated by 10 days interval. We use the standard deviation of the distribution of these differences as a measure of uncertainty on the strain data. These values range between a few nanostrain to up to about 50 nanostrain in the noisiest records.

3.2. Aftershock Data

3.2.1. Earthquake Rates and Stress Inference

Dieterich [1994] modeled the temporal evolution of seismicity rate of a population of earthquake nucleation patches subjected to an arbitrary stress history in the framework of rate-and-state friction. In this model, aftershocks occur on rate-weakening fault patches that are already accelerating toward failure. The seismicity rate, $\dot{N}$, is related to a state variable of the fault population, γ , proportional to its interevent time, by

$$\dot{N} = \frac{\dot{N}_{\text{bg}}}{\gamma \dot{\tau}_{\text{tect}}}, \quad (2)$$

where $\dot{N}_{\text{bg}}$ is the background seismicity rate and $\dot{\tau}_{\text{tect}}$ is the background tectonic stressing rate. The evolution of the seismicity state variable is related to the stressing history by the following equation:

$$\dot{\gamma} = \frac{1}{a\sigma}(1 - \gamma\dot{\tau}), \quad (3)$$

where σ is the effective normal stress (normal stress minus fluid pressure), a is a constitutive parameter quantifying the sensitivity of the fault friction to logarithmic changes of the sliding velocity, and $\dot{\tau}$ is the time-varying Coulomb stressing rate. Coulomb stress change on a fault is here denoted as $\Delta\tau$ and defined as $\Delta\tau = \Delta\tau' - f\Delta\sigma$, where τ' is fault shear stress, in the direction of slip, σ is effective normal stress, and f is the friction coefficient.

Solving equation (3) for a stress history consisting of a Coulomb stress step, $\Delta\tau$, applied in the middle of a time interval of duration Δt gives [*Dieterich et al.*, 2000]:

$$\Delta\tau = a\sigma \ln \left[\frac{\gamma_i + \frac{\Delta\tau}{2a\sigma}}{\gamma_{i+1} - \frac{\Delta\tau}{2a\sigma}} \right], \quad (4)$$

where γ_i and γ_{i+1} are the values at the beginning and end of the interval, respectively. Values of γ as a function of time are derived using equation (2) from estimates of seismicity rates on cells that contain seven or more aftershocks. The cells have a size of 2.0 km along strike, 1.0 km vertically, and 6.0 km in the fault-normal direction. Earthquake rates are assumed to be uniformly distributed within each cell.

During most of the aftershock sequence, aftershock decay rate is roughly proportional to $1/t$, and therefore, rates computed on logarithmic time windows are very sensitive to the duration of the window. Additionally, the time window scheme may suffer from an awkward situation in which a window does not contain any

event. To avoid these issues, we adopt the following approach for computing the seismicity rate. We compute the rate for an initial window containing five earthquakes. Next, we slide the window by one event and increase the window length by one event. This step is repeated until a predefined window length of 10 events is reached, after which the number of events in each window remains constant. The initial window size, incremental shift, and maximum size are referred to as N_o , ΔN , and $N_{\max}$, respectively (Table S1 and Figure S2). The corresponding time (with respect to the main shock) of the instantaneous rate measured in the interval $\Delta t = t_2 - t_1$ is [Rubin, 2002]

$$t' = \Delta t / \ln(t_2/t_1), \quad (5)$$

where t_1 and t_2 are the times of the first and last events in the window, respectively. By varying the window length we are able to capture slight temporal variations in the seismicity rates early in the aftershock sequence as well as the decay to the background rate late in the sequence. Note that sliding the window by a single event results in large overlap between windows and hence relatively smooth seismicity rate curves. Since we are only using the final stress change as input for the joint inversion procedure (section 5.1), this degree of overlap between neighboring windows has little effect on our final results. The procedure we use to estimate the uncertainties on the computed rates is presented in Appendix B. This procedure takes into account the number of events used to compute the seismicity rates, and hence, the errors on cells with small number of data are larger than on cells which contains tens of aftershocks.

The robustness of our slip inversion increases with the number of cells on which stresses may be evaluated. In order to increase the number of cells containing earthquake data, we compute the rates of earthquakes found in the catalog of Meng and Peng [2014]. This catalog was compiled by applying a waveform-matched filter technique and hence contains many events that were previously unlisted in the Southern California Seismic Network (SCSN) catalog. The matched filter-derived catalog is complete down to $M \sim 0$, about 1.5 magnitude units less than the magnitude of completeness of the relocated seismicity catalog of Hauksson et al. [2012] for the study area. Meng and Peng [2014] assign the template location to the newly detected events, which, for a dense aftershock sequence, might result in spurious increase of the stresses we infer. To eliminate this bias, we have relocated the newly detected events. Our approach for earthquake relocation is described in section 3.2.2.

The incorporation of the more complete, precisely located catalog allows us to compute earthquake rates (and hence stresses) at a large number of sites. Nonetheless, multiple sites contain too few earthquakes to reliably estimate their rate. The distribution of cells with well-resolved and poorly resolved earthquake rates is presented in Figures 3c and 3d. Note that here we assume that all events are occurring on the SJF plane, while in reality many events occur within the volume surrounding the fault. We account for the three-dimensional distribution of seismicity in the inversion procedure by perturbing the locations of seismically active cells with respect to the SJF fault plane (section 5).

Rates on each cell that contains between one and six aftershocks (target cell) are estimated from the three nearest cells in which rates are well resolved (reference cells). For this purpose, we resample the observed rates of reference cells in 1 h interval, stack them, and scale them by the ratio of number of events in the target cell to the number in the reference cells evaluated at the time of the last event in the target cell. This procedure allows us to estimate seismicity rates in areas with very few earthquakes. The stress change on cells whose background rates are well resolved (>50 events in 4 years preceding the El Mayor-Cucapah main shock, or in 3 months preceding the Collins Valley main shock) but do not contain any aftershock is set to zero. The real stress change might be positive or negative in these cells, but because of their large uncertainties, their overall contribution is small.

The regional catalog magnitude of completeness in Anza was reduced by 0.6–0.8 magnitude units between 2001 and 2010 with the deployment of 28 SCSN stations within about 100 km from the Collins Valley epicenter. A significant improvement in detection capabilities followed the deployment of the PBO borehole seismometers in 2006. To reduce the sensitivity of the inferred stresses to a possible bias in the background rates due to the reduction in the catalog magnitude of completeness, we evaluate the background seismicity rate prior to the El Mayor-Cucapah and the Collins Valley main shocks for events that occurred between 1 January 2006 and 3 April 2010 and between 14 April 2010 and 6 July 2010, respectively. For a small number of cells the background rate prior to the Collins Valley main shock is not well resolved (Figure 3b). These cells are assumed to maintain the level of activity estimated from the 4 year period prior to the El Mayor-Cucapah main shock

Figure 3. Background seismicity rate and spatial distribution of model cells for input stress calculation along profile A-A' (Figure 1). Vertical bars at the top indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). (a) Background seismicity rate between 1 January 2001 and 3 April 2010, spatially smoothed with a median filter of 6 km horizontal width and 4 km vertical width. Dashed polygons indicate areas that contain more than 50 events. (b) Same as Figure 3a but for the time interval preceding the Collins Valley main shock (14 April to 6 July 2010). Blue rectangles indicate cells that contained more than two events during that time interval. All other cells are assigned the same background rate as in Figure 3a. (c) Model cells with available background seismicity rates and >6 (black), 1–6 (grey), and 0 (brown) aftershocks in the 14 day period following the El Mayor-Cucapah main shock. (d) Same as Figure 3c for the Collins Valley aftershocks.

less active between 2006 and 2010. We analyze the spatiotemporal evolution of seismicity during the interval preceding the Collins Valley main shock in section 6.1.

Because of the logarithmic dependence of $\Delta\tau$ on the seismicity state parameter γ , the sensitivity of our results to the value of $\dot{\epsilon}_{\text{tect}}$ is not apparent from the formulation of *Dieterich's* seismicity model (equations (2)–(4)).

(the assumption is not critical because it involves very few cells). The background seismicity rates prior to the El Mayor-Cucapah main shock are computed using the relocated catalog of *Hauksson et al.* [2012], whose magnitude of completeness is about 0.5 magnitude units larger than the relocated version of *Meng and Peng's* [2014] catalog (section 3.2.2). To account for missing events, we complete the former catalog with the number of events that would have been present if it had the same magnitude of completeness as *Meng and Peng's* [2014] catalog, by assuming a Gutenberg-Richter distribution of event sizes with $b = 1$. In practice, we multiply the rates computed for the time period between 1 January 2006 and 3 April 2010 by a constant equal to 3.162. The spatial distribution of background rates used in equation (2) is obtained from earthquake counts between 2006 and 2010 smoothed using a median filter whose width is 6 and 4 km in the horizontal and vertical directions, respectively (Figures 3a and 3b). The parameters used for seismicity rate calculations are listed in Table S1.

Note that, similar to the manner in which seismicity rates during the 2010 transients are estimated, the background rates are computed for cells that extend out to 3 km on each side of the fault (Figure S2). Within each cell background rates are assumed to be uniformly distributed. In order for the background rates in the fault-normal direction to be considered representative, it is required that the spatial distribution of background seismicity is not significantly different from the distribution of earthquakes triggered during the 2010 transients. While this appears to be true for the Collins Valley sequence, the spatial distribution of earthquakes (given by their distance to line A-A' in Figure 1) that were triggered by the El Mayor-Cucapah earthquake is different from the distribution of events preceding that main shock (Figure S3). Presumably, that main shock also induced seismicity and aseismic slip on subsidiary faults that were

However, an analytical solution of equation (3), which provides a direct estimate of $\Delta\tau$ at time t for a transient starting a time $t = t_0$, gives [Helmstetter and Shaw, 2009, equation 55]:

$$\Delta\tau = a\sigma \ln \frac{\dot{N}(t)}{\dot{N}(t_0)} + \frac{\dot{\tau}_{\text{tect}}}{\dot{N}_{\text{bg}}} N. \quad (6)$$

This equation, valid at any time t , consists of two terms. The first is a nonlinear term due to the direct effect of rate-and-state friction. The second term is linear and prevails when nucleation is governed by Coulomb failure instead of rate-and-state friction, leading to a seismicity rate that is proportional to the stressing rate.

The sources of uncertainty on the inferred stresses can be readily appreciated from equation (6). If $a\sigma$ and $\dot{\tau}_{\text{tect}}/\dot{N}_{\text{bg}}$ are assumed, the major source of error is the uncertainties on $\dot{N}$. However, if $a\sigma$ is small such triggering is essentially instantaneous, the uncertainties on $\dot{N}$ have little impact and the major source of uncertainty on stress is the uncertainty on the estimated $\dot{\tau}_{\text{tect}}/\dot{N}_{\text{bg}}$. In this ratio the most uncertain term is $\dot{\tau}_{\text{tect}}$, which depends on assumptions about the way seismic coupling tapers off at depth. Uncertainties on N arise from earthquake location uncertainties and are small if the size of the seismicity cells is set larger than the location uncertainties. We present a method for approximating the errors associated with misestimated $\dot{N}$ in Appendix A. Our approach for incorporating uncertainties in the inferred stresses due to absolute event location errors is presented in section 5.1.

The tectonic stressing rate we assume in this study is obtained from an analytical solution [Segall, 2010] for the stresses induced due to strike-slip motion along an infinitely long vertical fault that is locked between 0 and 16 km and slips at a rate of 19 mm/yr [Lindsey and Fialko, 2013] below this depth. The actual value of $\dot{\tau}_{\text{tect}}$ is poorly constrained and hence strongly dependent on the assumptions regarding the form of loading that motion beneath the transition zone exerts on the seismogenic zone. We chose to model stresses that result from this motion with the least smooth model, one which contains a stress singularity at the locking depth. In reality, however, frictional heterogeneity within the transition zone may help regularize such high stress concentrations.

Our approach for estimating the stresses from observed seismicity rates assumes that the stress history during the time interval Δt is composed of a constant stressing rate, followed by a stress step, followed by a return to a constant (prestep) stressing rate. A different approach was taken by Segall et al. [2013], who assumed that during the interval Δt the stressing rate is a linear function of time. Under some conditions, and especially when the stressing rate varies smoothly as a function of time, Segall et al.'s [2013] approach may be more suitable for estimating stresses. Both approaches, however, should provide similar estimates if the duration of Δt is sufficiently small compared to T_γ , the typical timescale of fluctuations of the seismicity state parameter γ

$$T_\gamma \approx \gamma / \dot{\gamma}. \quad (7)$$

We can approximate γ and $\dot{\gamma}$ as follows:

$$\gamma \approx (\gamma_i + \gamma_{i-1})/2 \quad (8)$$

and

$$\dot{\gamma} \approx (\gamma_i - \gamma_{i-1})/\Delta t. \quad (9)$$

Thus, the requirement $\Delta t \ll T_\gamma$ leads to the following practical condition:

$$|\gamma_i - \gamma_{i-1}| \ll (\gamma_i + \gamma_{i-1})/2. \quad (10)$$

We find that the condition in equation (10) is met 95% of the times. As the stress changes in the remaining 5% of the times that generally do not exceed 10–20% of the maximum inferred stresses, we consider Δt to be sufficiently small such that it reasonably captures fluctuations in γ .

3.2.2. Earthquake Relocation

We use seismic data recorded by local SCSN, ANZA, and PBO stations in Anza (Figure 1). Our relocation approach follows the method of Got et al. [2002]. We divide the study area into polygons that delimit the SJF trace but do not cut through dense clusters of seismicity. The polygons overlap to allow for earthquake clusters to extend beyond a boundary. In each polygon, for each event pair and for each channel, we compute the waveform coherency in the 1–12 Hz frequency band for 2.56 s long windows centered on the P and S wave arrivals. Whenever phase data are not available (about 97% of the arrival times), we compute theoretical travel times using a 1-D local velocity model (E. Hauksson, personal communication, 2014) and run an

automatic picking algorithm to determine the arrival times [Shearer *et al.*, 2006]. The picking algorithm is based on a standard short-time average to long-time average approach for determining the phase arrival time. We visually reviewed the picks of tens of waveforms to ensure that the *P* and *S* wave trains are contained within the time window used for cross correlation. Note that the relocation procedure is based on differential travel times from many stations. Hence, as long as the correlation window contains the main *P* or *S* wave energies and as long as the arrival times are not systematically biased, the relocation results should not be sensitive to the absolute arrival time. In addition, our approach for stress calculations (section 3.2.1) is based on cells whose shortest dimension is 1 km, much larger than the error on the relative event locations.

Time delays are derived from the slope of the best linear fit (in the least squares sense) to the cross-phase spectrum of the shifted waveforms. This method generally yields time delay estimates with subsample precision. Once the polygon's coherency matrix is obtained, the events are grouped into clusters. We use a *k*-means clustering algorithm [Press *et al.*, 1992] with a coherency threshold of 0.8. These clusters contain from 10 to a few hundred events and span up to a few hundred meters.

Event relocation is performed using an iterative procedure, which weights the input in each iteration according to its deviation from the median delay time residuals. An implicit assumption in relocation techniques is that within each cluster, the interevent distances are much smaller than the average distance to the receiver. In order for this assumption to remain valid, the relocation is done sequentially. We start with clusters that contain up to 40 events and add an event to that cluster if its average coherency with the other members in the cluster exceeds 0.7. We then relocate the new cluster while considering delay times for pairs whose coherency is larger than 0.8. Our resulting catalog contains 7079 events divided into clusters whose dimensions are up to a few hundred meters.

4. Space-Time Analysis of the $M_w 7.2$ El Mayor-Cucapah and the $M_w 5.4$ Collins Valley Earthquake Sequences and Recorded Strain

In this section we analyze the space-time evolution of seismicity during the El Mayor-Cucapah and Collins Valley sequences. Our relocated version of Meng and Peng's [2014] catalog (section 3.2.2) allows us to study the response of the fault to perturbations from the local and remote main shocks at fine temporal and spatial scales. To do that, we compare the space-time evolution of seismicity immediately following the El Mayor-Cucapah and the Collins Valley earthquakes in Figures 4 and 5. The temporal evolution on a longer timescale is presented in Figure 2.

Seismicity is mainly concentrated along two segments located to the southeast and northwest of the AG. The AG is clearly visible in Figures 4 and 5 as an area mostly devoid of seismicity below 10 km depth, which extends between 25 and 40 km along fault strike. We refer to the two active segments that extend between 0–25 km and 40–60 km as the Hot Springs Cluster (HSC) and the Trifurcation Cluster (TC) [e.g., Zigone *et al.*, 2015], respectively. We identify migration of seismicity along fault strike during the Collins Valley sequence. We find weak evidence of migration following the El Mayor-Cucapah main shock. The small number of early relocated aftershocks triggered by the El Mayor-Cucapah earthquake makes difficult asserting whether the spatiotemporal evolution in Figure 4 is substantially different from what would be the result of a random fluctuation in the background seismicity rate. The long-term average background rate is higher in the TC than in the HSC (Figure 3a), which increases the likelihood of observing an apparent migration from the TC toward the HSC in a given time window.

The spatiotemporal distribution of the Collins Valley aftershocks indicates that their migration speed is inversely proportional to time since the main shock. We added a line indicating a logarithmic migration trend to Figure 5. Similar rapid initial expansion of the aftershock zone during the Collins Valley sequence, as well as during sequences triggered by several other moderate Anza main shocks, was also observed by Meng and Peng [2016]. Of the events recorded by the Anza strain network, only the El Mayor-Cucapah and the Collins Valley earthquakes generated strong, coherent signals at the borehole strainmeters, which allowed us to perform the comprehensive analysis presented in this paper.

We compare the cumulative number of aftershocks in the TC and HSC as a function of time since the main shocks (Figures 4b and 5b). This analysis allows us to identify abrupt changes in seismicity rates. The instantaneous response of seismicity in the TC to the local (Collins Valley) and remote (El Mayor-Cucapah) stress perturbation is similar. In both cases cumulative event numbers exhibit approximately a logarithmic

Figure 4. Space-time seismicity diagrams for the El Mayor-Cucapah sequence. (a) Aftershock locations projected along profile A-A' (Figure 1) as a function of time since the main shock. The symbol color and size indicate depth and magnitude, respectively. Vertical bars at the top indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). (b) Cumulative event count as a function of time since the El Mayor-Cucapah earthquake. Blue and red curves are for the TC and HSC, respectively.

with time for the remaining 2 weeks. The temporal evolution of the post-El Mayor-Cucapah events in the HSC differs significantly: their rate accelerates immediately following the main shock then undergoes a few days of quiescence and accelerates again about 6 days after the main shock. This behavior is similar to the activity in the TC during the El Mayor-Cucapah sequence. We note that the magnitude of events in the HSC during the quiescence periods does not exceed $M = 1.5$. This observation, together with high strain rates observed at the surface, suggests that loading due to nearby fault creep is the likely cause for the rate increase in the HSC 6 and 4 days after the El Mayor-Cucapah and Collins Valley earthquakes, respectively. Quiescence in nearby areas may be the result of stress unloading at the tail of the passing creep front. The lack of a strong main shock and the increase in strain rates presented below suggest that the increase in seismicity rates after the quiescence is due to additional loading from creep occurring later in that sequence.

We examined the strainmeter data for evidence of changes in surface strain rates associated with the fault slip changes inferred from the analysis of the seismicity rates presented in Figures 4b and 5b. The correlation between strainmeter and seismicity data is not obvious. Out of the six available strainmeters, only two or three stations exhibit temporal behavior consistent with our interpretation. For example, the rotation of principle strain directions at stations B081 and B088 is consistent with creep propagating along the SJF strike during both sequences, and the acceleration of seismicity on the sixth day following the El Mayor-Cucapah main shock is manifested by a gradual increase in strain rates observed only at stations B087 and B086. In addition, the records contain small offsets that are not correlated between stations, which are most likely the result of local unmodeled noise near the stations. Quantifying the degree of temporal correlation between creep and seismicity in Anza is beyond the scope of this paper study as it would require sufficient resolution to determine fault slip from the strainmeter data only. Instead, we attempt to infer the cumulative slip distribution that gives rise to deep seismicity and surface strain and thus proceed by jointly inverting the two data sets using the scheme presented in the next section. This analysis allows us to test the hypothesis that static stress transfer from fault patches slipping aseismically triggers seismicity along the SJF. The potential for time-dependent slip inversion will be examined in future work.

dependence on time since the main shock, indicative of an Omori-type aftershock sequence. However, while the Collins Valley aftershocks in the TC continue for about 2 weeks, the post-El Mayor-Cucapah events in that cluster take only 3–4 days to decay. Note that according to Dieterich [1994] seismicity model, the characteristic time for decay of seismicity following a stress step is not dependent on the size of the imposed stress change. In the framework of this model, a decrease of the product $a\sigma$ (for example due to pore pressure changes) or an increase in the stressing rate during the El Mayor-Cucapah sequence may have induced a shorter aftershock sequence following the El Mayor-Cucapah earthquake. The post-El Mayor-Cucapah rate increase is followed by a short quiescence and then by an increase on the sixth day to a new rate that is comparable or higher than the premain shock seismicity rate. Anomalous high seismicity rates in Anza persist for several months leading to the Collins Valley main shock (Figure 2). We present a detailed spatiotemporal analysis of seismicity leading to the Collins Valley earthquake in section 6.1.

The immediate effect of the Collins Valley main shock is to reduce the seismicity rates in the HSC cluster. The rate increases about 4 days after the main shock and then decays logarithmically

Figure 5. Space-time seismicity diagrams for the Collins Valley sequence. Dashed curve indicates migration velocity that decays as $1/t$, where t is time since the main shock. Same as in Figure 4 for details.

stresses we infer for the Collins Valley sequence on cells located near the main shock hypocenter. However, as we show below, the peak slip we infer from the joint data set is over 40 cm, much larger than the expected coseismic slip during the M_w 5.4 Collins Valley main shock (between 1 and 10 MPa assuming a standard circular crack [Eshelby, 1957]). We therefore ignore the contribution of coseismic stress changes and attribute the inferred stresses to postseismic slip.

The inversion procedure is set up to minimize the following cost function:

$$C = C_1 + \beta^2 C_2 = \|W(Au - d)\|^2 + \beta^2 \|S\Delta u\|^2, \quad (11)$$

where $\|\cdot\|$ denotes the L_2 norm of a vector. The term C_1 is a joint misfit function, in which d is a vector containing the observed strains and inferred stresses from seismicity rates at the end of the transient (as described in section 3.2.1), u is the solution vector comprising the slip distribution at the end of the transient, A is a model matrix relating model parameters to data, and W is a diagonal matrix of weights given to each data set. The term C_2 is a regularization term introduced to stabilize the inversion by imposing spatial smoothness on the slip distribution. S is a smoothing matrix, Δ is the Laplacian operator, and β is a smoothness coefficient that controls the importance of the regularization term.

The weights are computed according to the errors of each data set:

$$W_i = 1/s_i \quad (12)$$

for the strain data and

$$W_i = \alpha_\beta/s_i \quad (13)$$

for the stress data, where s_i are the standard deviation of the strain residuals (section 3.1) and of the estimated error on the inferred stresses (Appendix A). The stress errors range between 10% and 20% of the inferred stress. The constant α_β is chosen such that we obtain equal misfits to the joint data set [e.g., Shirzaei and Bürgmann, 2013]. For each value of β , we find the corresponding α_β by 1-D line minimization [Press et al., 1992] of the cost function with the constrain that the difference between the weighted misfits to each data set are smaller than 1% of the average of the misfit to each data set. For most of the inversions, the value of α_β ranges between 0.1 and 10. We found that after a few iterations the differences between the normalized misfits match our predefined threshold.

5. Joint Inversion of Strain and Seismicity Data

5.1. Method

We model the seismicity and the strain variations following the El Mayor-Cucapah and the Collins Valley earthquakes assuming that they were driven by triggered aseismic slip on the SJF. The observed strains and inferred stresses are therefore jointly inverted to recover the best fitting distribution of slip. In this work we focus on the total slip produced by each transient rather than on the time dependency of slip. The data for our inversion thus comprise the measured strains and the seismicity-derived stress changes accumulated during each transient. Here we try to explain these observations as a result of triggered aseismic slip on the SJF. In reality, the stresses we infer also contain the static coseismic stress change due to the main shock. Note that the coseismic strain change was removed from the strain data. Because of the rapid decay of stress with distance away from the main shock, our approach may bias the

We define the matrix A as follows:

$$A = (G_{ij}) \quad (14)$$

for the strain only inversion and

$$A = \begin{pmatrix} G_{ij} \\ K_{kj} \end{pmatrix} \quad (15)$$

for the joint inversion. G_{ij} and K_{kj} are elastic kernels [Okada, 1992], which relate strain at location i , ϵ_i , and Coulomb stress change at location k , $\Delta\tau_k$, to unit slip at location j , u_j . Slip is constrained to be right lateral and positive and is tapered by overweighting the Laplacian on the edges relative to cells in the interior of the model. Model cells measure 2.0 and 1.0 km in the along-strike and along-depth directions, respectively.

Inherent differences in the sensitivity of the geodetic and aftershock data to fault slip are to be taken into account in order to obtain a realistic slip distribution satisfying both data sets. While the geodetic data are most sensitive to the long-wavelength component of slip, the aftershock data are most sensitive to a local, short-wavelength slip variation, probably of similar scale as the aftershock clusters. To account for these differences, it is useful to weigh the model smoothness according to the resolution power of the joint data set. We downweight poorly resolved areas in the model by defining the matrix S as follows:

$$S = \frac{\text{diag}(A^T A)}{\max(A^T A)}. \quad (16)$$

Applying these weights to the smoothing operator adjusts the correlation length between cells according to the resolution power of the data, thus providing compact slip distributions in well-resolved areas and vice versa [Ortega Culaciati, 2013].

Seismicity southeast of the Anza seismic gap is distributed over several fault branches that compose a complex fault zone, where multiple active secondary faults are located within some distance from the main fault strand. We assume that slip on the main fault triggers seismicity on these secondary faults via static stress transfer. Seismicity rates and Coulomb stress histories are computed on 3-D cells that surround each slip cell as described in section 3.2.1. The Coulomb stress history obtained by solving equation (4) is taken to be the representative value at a reference point inside the cell that has the same on-fault position as the midcell point but can be at a distance to the main fault. We incorporate the uncertainties associated with the absolute location of events in the catalog by coupling the least squares algorithm with a Monte Carlo scheme, such that in each realization we perturb the distances between the reference points of seismic cells and the main fault plane. The distances are drawn from a normal distribution whose standard deviation (S_{norm} in Table S1) is equal to 2 km. Using a distribution with $S_{\text{norm}} > 2$ km requires unrealistically large fault slip to fit the seismicity data, which significantly reduces the fit to the strain data. However, due to the shape of the distribution function, a few cells are perturbed to distances that exceed 3 km. An alternative approach for incorporating the absolute location uncertainties would be to perturb earthquake locations while keeping fixed the reference stress estimation point. However, since the absolute location errors are much larger than the relative ones, perturbing the entire population might eliminate the (real) spatial clustering in the catalog. Our approach conforms to that spatial clustering. By first grouping events in cells whose dimensions are larger than the relative location uncertainties and then perturbing each cell's location, we effectively maintain the spatial structure that results from the earthquake relocation procedure.

Note that, unlike shear stresses, the Coulomb stresses due to slip on the main fault are not symmetrically distributed on both sides of the fault but may have opposite signs depending on the location relative to the dislocation tip. Thus, randomizing the distances to the fault plane may result in a situation in which a cell with (inferred) positive stress change is put in an area in which the model predicts negative stresses and vice versa. This may lead to drastically different slip distributions between consecutive iterations, which will not allow us to draw statistically significant conclusions from the ensemble of slip distributions. To mitigate this issue, we perturb the cell in the direction that matches the average location of the earthquakes contained in that cell relative to the fault plane (see Figure S2 for further details). We repeat the inversion 100 times (for each value of α and β) and report the results from the 50 best fitting inversions.

Equation (11) is solved using the least squares algorithm with positivity constraints of Lawson and Hanson [1974]. The static friction coefficient, f , is assumed constant, and its value is set to 0.6. The stresses on off-fault sites are related to slip as in equation (15). To reduce the number of unknowns in the problem, we assume

Figure 6. Inversion results. (a, b) The square root of the regularization term C_2 as a function of the square root of the joint misfit C_1 . (c, d) Square root of C_1 as a function of the constitutive parameter a . The color scale indicates the value of the smoothing coefficient β . Figures 6a and 6c: El Mayor-Cucapah. Figures 6b and 6d: Collins Valley. Green circles indicate the preferred solutions.

that the orientation of secondary faults is identical to that of its nearest slip cell on the main fault plane. This assumption is consistent with the observation of a rather homogeneous distribution of focal mechanisms along the Anza section of the SJF [Bailey *et al.*, 2010]. We assume Poisson's ratio and Young's modulus equal to 0.25 and 30 GPa, respectively. The value of the constitutive parameter a is assumed to be spatially uniform. Models that extend to the free surface require several centimeters of shallow slip to fit the El Mayor-Cucapah strain data, and thus, we only model slip at depths larger than 2 km. This maintains the same fit to the strain data and is more consistent with the lack of observable surface creep during the 2010 transients [Lindsey *et al.*, 2014]. The inversion scheme is presented graphically in the flow chart and sketch in Figures S1 and S2.

5.2. Results: Slip Distribution and Static Stress Transfer to Seismic Cells

We apply the inversion procedure outlined in section 5.1 to strain and aftershock data from the 2010 El Mayor-Cucapah and Collins Valley main shocks. Figures 6a and 6b present, for each value of the smoothness coefficient β , the value of the regularization term C_2 as a function of the joint misfit C_1 . The decrease in the data misfit with an increase in the roughness of the model (smaller values of β) is a well-known attribute of geophysical inversions and represents a trade-off between model resolution and variance. We adopt an L curve approach to choose a model that is sufficiently rough without overfitting the data: we select the β value corresponding to the inflection point of the solid blue curve in Figures 6a and 6b. Note that, according to the conventional L curve approach, a somewhat rougher model should have been selected ($\beta \sim 1 \times 10^{-5} - 1 \times 10^{-4}$ in Figures 6a and 6b). Such rough models contain significant slip close to the surface. Because previous studies failed to observe significant surface creep along the Anza section during the 2010 transients, we consider such rough slip distributions with significant shallow creep to be less likely.

A central assumption in our inversions is that static stresses due to triggered aseismic slip on a main fault strand are transferred to secondary seismically active faults. Under this assumption we obtain a satisfactory fit to the strain and aftershock data. In Appendix B, we test an alternative mode of stress transfer in which seismicity occurs on asperities embedded in the main fault, driven by stresses imposed by creep in the surrounding fault regions. This assumption is common in models of repeating earthquake sequences

Figure 7. Inferred aseismic slip distributions along profile A-A' (Figure 1). Triangles indicate the along-strike location of PBO strainmeters. Star indicates location of the Collins Valley main shock. Vertical bars at the top indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). Dashed red curves indicate seismically active segments within the HSC and TC. (a) El Mayor-Cucapah. (b) Collins Valley. (c) Slip contours of aseismic slip triggered by the El Mayor-Cucapah (blue) and Collins Valley (magenta) main shocks.

[e.g., *Chen and Lapusta*, 2009]. We find that this model provides a significantly poorer fit to the joint data set.

Figure 7 presents the slip distributions of our preferred models of aseismic slip triggered by the El Mayor-Cucapah and Collins Valley main shocks. The observed and modeled strains are shown in Figure 8, and the input, residual, and modeled on-fault stresses are shown in Figure 9. Note that input stresses in Figures 9a and 9d are estimated at sites located off the modeled fault, while the output in Figures 9c and 9f shows the on-fault stresses induced by the inverted aseismic slip. Figure 8 also presents the computed GPS displacements at several sites in Anza for the slip distributions in Figure 7. For most sites the predicted displacements do not exceed 2.5 mm, a value comparable to the uncertainty of GPS data. Because the sensitivity of the GPS network to deep (> 5 km) fault slip is small, we do not expect the GPS instruments to detect the deep slip imaged via our joint inversion approach.

The moment magnitude of triggered aseismic slip in a 10 day interval is 5.9 and 5.8 for the El Mayor-Cucapah and Collins Valley sequences, respectively. To estimate the robustness of this result, we have conducted synthetic tests in which we invert only the strain data for slip in scenarios with uniformly distributed slip around the Collins Valley main shock with moment magnitude 6.2. We found that the strain-only inversion

recovered the input moment but, as expected, was not able to recover the slip distribution satisfactorily. We therefore conclude that the estimated moment is robustly constrained by the inversion procedure.

In both sequences, the modeled slip is spatially anticorrelated with aftershock locations. This feature results from our assumption that the sign of aftershock rates change matches the sign of the Coulomb stress change due to slip on the main fault. Because of this anticorrelation, the output slip distribution is nonsmooth like the spatial aftershock distribution.

In our preferred models, triggered aseismic slip occurs primarily below a dense cluster of events located south-east of the AG and extends to the northwest below the Anza segment. The amplitude and location of small slip patches around the TC and HSC are mainly constrained by the stress distributions inferred from aftershock activity, while the overall distribution of slip (at length scales larger than about 1 km) is determined primarily by the strain data. In order to obtain a reasonable fit to the data set, slip must occur below the AG itself.

The triggered aseismic slip distribution in the patch lying directly to the SE of the HSC segment (20–30 km along profile A-A') is different between the two sequences. It has larger amplitude and is more compact during the Collins Valley sequence than during the El Mayor-Cucapah sequence. A qualitative difference in the temporal evolution of seismicity in the HSC between these two sequences was also demonstrated in section 4. The preferred slip model for the Collins Valley sequence allows us to fit the inferred stresses in the TC reasonably well but is not capable to fit the seismicity data from the HSC (Figures 9d and 9e). Because the number of cells with poorly resolved rates in that cluster is large compared to the TC, data from this cluster are downweighted

Figure 8. Observed and modeled strains and surface displacements. (a, c) El Mayor-Cucapah. (b, d) Collins Valley. Dashed black line indicates the modeled fault trace. Observed and modeled strains are indicated by red and blue crosses, respectively (Figures 8a and 8b). Location of observed strains are offset for clarity. The dashed polygons indicate the area covered by Figures 8c and 8d. Observed and predicted (using slip models in Figure 7) surface displacements at nearby GPS sites are indicated by red and blue vectors, respectively (Figures 8c and 8d). 1σ uncertainties are indicated by the red circles.

in the inversion. It is also possible that static stress transfer due to earthquake interaction, a mechanism unaccounted for by our approach, is more important for triggering seismicity in the HSC during the Collins Valley sequence. We analyze the relative contribution of static stress transfer due to earthquake interactions and triggered aseismic slip in section 6.2.

We find that the overlap between concentrations of peak-triggered aseismic slip following the Collins Valley and the El Mayor-Cucapah main shocks is small. (Figure 7c). This is reminiscent of the behavior of earthquakes rupturing fault gaps left unbroken by previous earthquakes and initiating in highly stressed areas at the edges of past ruptures [e.g., Wei *et al.*, 2013; Ye *et al.*, 2014; Sokos *et al.*, 2016]. It is, however, difficult to draw a strong conclusion based on these results as they are dependent on our modeling assumptions and regularization. Because the degree of smoothing that we apply to the model varies with location according to the resolution of our model (section 5.1), slip beneath the AG is relatively smoothed compared to shallower patches near the HSC and TC. In general, rougher models ($\beta < 1 \times 10^{-3}$) maintain the level of spatial overlap as in Figures 7a and 7b for the shallower concentrations of slip (10–13 km depth) but tend to become less complementary than the preferred models in the deepest segments beneath the AG.

The stress changes that we infer are proportional to the product $a\sigma$ (equation (4)), and thus, the inversion procedure should provide some constraints on this parameter. In this paragraph we assume that the effective normal stress σ is hydrostatic with crustal rock density of 2700 kg/m^3 . The value of a quantifies the sensitivity of friction to logarithmic changes in the sliding velocity of the receiver faults and therefore provides further insight into the mechanical properties of the transition zone. In practice, however, our inversions cannot resolve the product $a\sigma$. Figures 6c and 6d present the joint misfit as a function of a (with σ unchanged)

Figure 9. Observed and modeled stresses along profile A-A' (Figure 1). Vertical bars at the top of each panel indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). (a–c) El Mayor-Cucapah. (d–f) Collins Valley. Star indicates the Collins Valley main shock hypocenter. Figures 8a and 8d, 8b and 8e, and 8c and 8f show the inversion input, stress residuals (using the models in Figure 7) and output on-fault stresses, respectively.

for inversions with $\beta = 1 \times 10^{-3}$. We performed inversions with a in the range $1 \times 10^{-5} - 1 \times 10^{-1}$ and find that they provide equal fit to the joint data set.

The values we report for $a\sigma$ are likely sensitive to our inversion scheme. Our method only uses the final stress and strain changes at the end of the time interval containing aftershock activity and rapid strain changes and ignores temporal variations of these parameters. If the characteristic time for relaxation of stresses at seismically active areas is much shorter than the typical timescale for seismicity rate changes, the inversion may lose resolution on the value of $a\sigma$. Note that in Dieterich's aftershock model the product $a\sigma$ controls the characteristic duration of the aftershock sequence. When the value of $a\sigma$ is very small, seismicity rate is proportional to the stressing rate, as is predicted by Coulomb failure theory (equation (6)). In this situation, aftershock nucleation is essentially instantaneous, and rate-and-state theory is not needed in order to explain the observations. However, our observations suggest that aftershock rates actually decay approximately as $1/t$. If triggering of seismicity is instantaneous, such dependency may be the result of stress relaxation in the creeping segments [Perfettini and Avouac, 2004], in which case the constitutive parameters of the model (velocity strengthening) fault will control the spatiotemporal evolution of seismicity.

6. Discussion

6.1. Aseismic Slip and Seismicity Triggered by the El Mayor-Cucapah Earthquake Leading to the Collins Valley Earthquake

The El Mayor-Cucapah earthquake triggered an aftershock sequence on the SJF (Figure 4), whose rate decayed within 4–5 days to about 10 times the premain shock rate (Figure 2, top). The new rate persisted through the months leading to the Collins Valley earthquake. This sustained period of elevated seismicity rate includes two $M > 4$ events near the impending Collins Valley hypocenter, 24 days before that main shock (Figure 1). The time between these two events is anomalously short compared to the average interval of 2.16 ± 1.80 years between $M4-5$ events in the TC since 1981. No significant transient is observed in the borehole strainmeter data between 10 days after the El Mayor-Cucapah and the Collins Valley main shocks.

The elevated seismicity rates that persisted during the months leading to the Collins Valley main shock and the spatial complementarity between the aseismic slip triggered by the El Mayor-Cucapah and Collins Valley main shocks (Figure 7c) are intriguing observations. Stress release due to aseismic slip triggered by the El Mayor-Cucapah main shock should have been followed by a gradual return to the background seismicity rates, or even lower, especially given the large amount of postseismic slip. The average fault slip rate immediately following the El Mayor-Cucapah earthquake increased by a factor larger than 100 relative to the long-term slip rates. A period of increased slip rate must be compensated later by slip rates lower than

Figure 10. Earthquake rates and stresses along profile A-A' (Figure 1). Vertical bars at the top of each panel indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). (a) Seismicity rates as a function of distance along fault strike, from 7 to 14 days (red) and from 14 to 94 days (blue) after the El Mayor-Cucapah earthquake. (b) Cumulative shear stresses inferred from seismicity rates as a function of distance along fault strike, from 0 to 14 days (red) and from 14 to 94 days (blue) after the El Mayor-Cucapah earthquake. Calculations were done using $a = 1 \times 10^{-3}$. Blue and black stars indicate the location of the 7 July 2010, M_w 5.4 Collins Valley and the 13 June 2010 $M > 4$ earthquakes, respectively. Seismicity rates and stresses are averaged between 12 and 15 km depth.

the long-term average and hence lead to seismic quiescence as reported in aftershock sequences following small ($M < 2$) earthquakes [Ziv *et al.*, 2003] and, more rarely, following large ($M > 6$) main shocks [Marsan, 2003]. The spatial complementarity between the two events suggests that strong aseismic slip induced by the El Mayor-Cucapah redistributed stresses in a manner that promoted the occurrence of the Collins Valley main shock.

We propose that the El Mayor-Cucapah main shock initiated a several monthlong transient on the SJF, too slow to be detected by the borehole strainmeters. Analysis of the more sensitive laser strainmeter installed in the Piñon Flats Observatory confirms that a several monthlong transient on the SJF initiated with the El Mayor-Cucapah main shock [Agnew *et al.*, 2013]. The observed strain rates indeed decay 10–14 days following the El Mayor-Cucapah earthquake. According to this scenario, the deep-triggered aseismic slip that we imaged using the first 10 days of seismicity and strain data following the El Mayor-Cucapah earthquake (Figures 7a and 7c) continued to accumulate, more slowly, for at least another 80 days. Deep slip during this period may have increased loading on nearby secondary faults, thus triggering seismicity in the TC and eventually culminating in the Collins Valley earthquake.

In order to test the hypothesis of a deep transient extending up to 94 days following the El Mayor-Cucapah earthquake, we estimate stresses from earthquake rates between 14 April and 7 July 2010. This time window begins when the surface strain signal due the El Mayor-Cucapah aseismic slip becomes negligible (Figure 2) and ends on the day of the Collins Valley earthquake. The duration of this interval is much longer than the intervals we considered in previous sections for the analysis of aseismic slip induced by the two main shocks. This complicates the strain data processing and makes our joint approach difficult to implement. We therefore only use stresses inferred from seismicity using the approach outlined in section 3.2.1 to crudely characterize the evolution of fault slip.

Seismicity and inferred stresses are consistent with a scenario in which the aseismic slip triggered by the El Mayor-Cucapah main shock continued to grow in amplitude but remained stationary in space, thus triggering seismicity in nearby segments. Figure 10 presents the earthquake rates in the weeks leading to the Collins Valley earthquake and the inferred cumulative shear stresses in the region extending out to 3 km on both sides of the SJF strand, between approximately 10 and 15 km depth. In particular, Figure 10a shows that the spatial distribution of seismicity rates is similar in the periods of 7–14 days and 14–94 days following the El-Mayor Cucapah earthquake. As a result, the highly stressed areas on both sides of the AG inferred for days 0–14 and 14–94 following the El Mayor-Cucapah earthquake are spatially correlated (Figure 10b). If triggered aseismic slip had continued migrating during the months leading to the Collins Valley earthquake, it should have resulted in stress concentrations farther away from the peak slip locations inferred in the initial 14 days (Figure 7a). In contrast, we find that high stress areas are located on segments directly adjacent to the zone of peak aseismic slip triggered by the El Mayor-Cucapah earthquake, while low-stress areas reside within that zone. This spatial pattern resembles the stress field that would have resulted from continuous slip in the segment hosting the peak aseismic slip triggered by the El Mayor-Cucapah earthquake. Moreover, if the

amplitude of triggered aseismic slip had continued to increase significantly, it should have resulted in a distinguishable strain signal at the surface, which we do not observe. The stationarity of the late El Mayor-Cucapah aseismic slip is further supported by the lack of observed migration of seismicity in that time interval (Figure 4).

Stresses to the SE of the peak aseismic slip triggered by the El Mayor-Cucapah earthquake strongly encourage the Collins Valley main shock. From the location of peak aseismic slip, the pre-Collins Valley earthquake shear stresses increase up to about 0.5 MPa near the Collins Valley hypocenter and are also high near the SE edge of the fault, where two $M \approx 4$ earthquakes occurred 25 days before that event. Such stress levels are extremely high compared to the ambient tectonic stresses acting on the SJF: they are 50 to 100 times larger than the stresses at the Collins Valley hypocentral depth resulting from uniform steady slip below 16 km at 19 mm/yr during 90 days.

This analysis suggests that the El Mayor-Cucapah earthquake triggered a long aseismic transient at the bottom of the seismogenic zone, which was accompanied by elevated seismicity rates. The case for sustained seismicity induced by deep-triggered aseismic slip due to coseismic stress redistribution is most clearly exemplified by the Collins Valley sequence.

6.2. Importance of Earthquake Interactions

Earthquakes are known to change the stresses in adjacent regions and trigger their own aftershock sequences and postseismic slip. Since the stress change in the vicinity of a given aftershock may greatly exceed the stresses imposed in that location by the more distant main shock or by aseismic slip, it is not readily apparent which is the dominant mechanism driving seismicity at remote sites. To measure the extent to which earthquake interactions are important in triggering seismicity along the Anza segment, it is instructive to compare the static stress changes transferred by aftershocks to the stresses transferred by aseismic slip. Next, we compare the cumulative stresses in the TC and HSC to the output stress distributions in each of the sequences and show that during both studied sequences, earthquake interactions play a negligible role in aftershock triggering compared to the loading due to aseismic slip.

We start by analyzing the activity in the TC. Given its distance from the El Mayor-Cucapah main shock, dynamic stresses, which decay slower than static stresses with distance to the main shock, may have been dominant in triggering aseismic slip in the TC. The expected amplitude of dynamic stresses in Anza due to the passage of seismic waves generated by the El Mayor-Cucapah main shock is up to several tens of kilopascal [e.g., Hill, 2012]. For comparison, our aseismic slip distribution implies stress changes of the order of several megapascal near the TC (Figure 9e). Thus, it is more likely that triggered aseismic slip was the dominant driver of the sustained seismicity in that cluster during the El Mayor-Cucapah sequence. For the Collins Valley sequence, the large aseismic slip moment (four times larger than the coseismic moment) suggests also that aseismic slip is the dominant driver of aftershocks.

Earthquake triggering in the HSC is more challenging to explain. Events in that cluster may have been dynamically triggered by seismic waves from the El Mayor or Collins Valley earthquakes or statically triggered due to long-range elastic interactions following the Collins Valley earthquake, and so one alternative would be that aseismic slip in the HSC was triggered by events occurring in that cluster.

To examine this alternative, we compare the static stress changes induced by aftershocks and by the estimated triggered aseismic slip. The slip of each microearthquake is inferred from its seismic moment assuming a standard circular crack *Eshelby* [1957] with stress drop of 3 MPa, a typical value in southern California. In most cells, stresses induced by a microearthquake are computed using Okada's stress kernels. In cells containing a microearthquake, the induced stresses are highly nonuniform and we compute the stress change at the center of the cell using the following approximate relation for a circular crack [Dieterich, 1994]:

$$\Delta\tau = \begin{cases} \Delta\tau_e \left[\left(1 - \frac{R^3}{x^3} \right)^{-\frac{1}{2}} - 1 \right], & x > R \\ -\Delta\tau_e & x < R \end{cases}, \quad (17)$$

where x is the distance between the cell center and the crack center, $\Delta\tau_e$ is the stress drop within the crack, and R is the crack radius. This relation does not incorporate any azimuthal dependency in the stress distribution but provides a first-order estimate of the stress decay with distance. To account for the uncertainty in the value of x , we randomly perturb the catalog locations and report the average values from 50

Figure 11. Amplitude of shear stress differences between cumulative stresses due to Collins Valley aftershocks and stresses due to the Collins Valley aseismic slip along profile A-A' (Figure 1). Vertical bars at the top of each panel indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). Differences are normalized by the stresses due to aseismic slip. Contours are for the aseismic slip distribution in Figure 7b. Star indicates the Collins Valley main shock hypocenter.

Monte Carlo simulations. In addition, to prevent stress singularities from exceeding the elastic strength of the medium, we cap the stresses at the crack tip at a value equal to $0.6 \times \sigma$.

We compare the static stresses due to aftershocks to the stresses from aseismic fault slip during the Collins Valley sequence. Since the number of HSC aftershocks is larger and slip along the HSC segment is smaller during the Collins Valley sequence than during the El Mayor-Cucapah sequence (Figure 7), using the Collins Valley data set to test static stress transfer is more conservative. Figure 11 presents the ratio between the sum of stresses due to Collins Valley aftershocks and the stresses due to cumulative aseismic slip in a 10 day period as a function of position. That ratio is small; it does not exceed 2% in the HSC or the TC. We there-

fore conclude that stresses due to long-range static earthquake interactions can explain only a small fraction of seismicity in the HSC and that most of the events in the TC are triggered by a creep transient initiated near the main shock. The relative contribution of static stress transfer due to earthquake interactions in the HSC, which are not modeled in our inversion, may help explain the poor fit of the Collins Valley-triggered aseismic slip model to the stress data in that cluster (Figures 9c and 9d).

Creep near the HSC during the Collins Valley sequence could have been triggered dynamically due to body waves or statically due to aseismic slip extending below the AG from the source region. To examine these two alternatives, we inverted the aftershock and strain data for the cumulative slip distribution in the first day following the main shock. Due to large errors of the stresses inferred from the rates of aftershocks occurring during the first day following the main shock (Figure A1), the modeled slip is less reliably determined than the one we obtained using 10 days of strain and aftershock data. Accounting for these uncertainties, we find that the stress change near the HSC due to static slip occurring in the first day following the Collins Valley main shock is of the order of about 1 MPa, much larger than the expected dynamic stress change due to body waves at this site. This implies that aseismic slip in the HSC is mainly triggered by static stresses transferred to the area from creep extending from the TC.

6.3. Importance of Secondary Aftershocks

A major assumption in Dieterich's aftershock model is that nucleation sites do not interact: stress perturbations caused by aftershocks are neglected. These additional stresses can trigger secondary aftershocks, which may occur in the days and weeks following the main shock and at distances of several rupture radii from the main shock [e.g., Ziv, 2006a]. A large fraction of secondary aftershocks relative to direct aftershocks may bias the inferred stresses and hence the slip model. To assess this potential bias, we quantify the importance of secondary aftershocks in the Collins Valley sequence, which contains more events than the El Mayor-Cucapah sequence and is thus more amenable to the statistical analysis we perform. We show that secondary aftershocks play a negligible role in transferring stresses during the first few days following the main shock, the period during which strain rates are highest.

Studies that estimate the fraction of secondary aftershocks usually classify individual events by declustering the seismic catalog. Since most declustering algorithms use somewhat arbitrary conditions to discriminate between main shocks, direct, and secondary aftershocks, we choose a different approach to the problem. We generate synthetic earthquake catalogs in which the rates of secondary and direct aftershocks are known and use these estimates to correct the observed rates for the rates of secondary aftershocks. This approach allows us to infer stresses from the rates of direct aftershocks in our observations. Next, we describe our approach and compare the slip distribution from an inversion in which stress data are based on the rate of direct aftershocks to the distribution presented in section 5.2.

We generate synthetic catalogs by using an epidemic-type aftershock model (ETAS) [e.g., Ogata, 1999; Felzer and Kilb, 2009]. Such statistical models are often used to mimic aftershock distributions by assuming several

Figure 12. Temporal evolution of direct and secondary aftershocks. (a) Observed and modeled aftershock counts for the Collins Valley aftershock sequence. (b) The fraction of secondary aftershocks as a function of time from ETAS simulations.

empirical relations. Here aftershock rate decays with time since the main shock according to the modified Omori law [Utsu, 1961]:

$$\dot{N} = \frac{k}{(t+c)^p}, \quad (18)$$

where k , c , and p are fitting coefficients that we obtain from modeling the cumulative event counts. The value of the aftershock decay constant, p , is usually near one. For short aftershock sequences, such as the Collins Valley one, it is advantageous to use the cumulative form of equation (18) with $p = 1$ [Ziv, 2006b]:

$$N = \int \dot{N} dt = k \ln(t+c) + N_0, \quad (19)$$

where N_0 is an integration constant. Figure 12 presents the observed and modeled cumulative number of aftershocks as a function of time since the Collins Valley main shock. The modified Omori law with $p = 1$ provides a good fit to the aftershock data. The second empirical relation describes the distribution of earthquake sizes by a Gutenberg-Richter magnitude-frequency distribution with $b = 1$,

which characterizes that distribution in Anza during the 2010 transients. The third empirical relation describes the aftershock density decay as a function of distance to the main shock by an inverse power law. The distance decay exponent is set to 1.9, which is more suitable for triggering of direct aftershocks [Marsan and Lengliné, 2010]. Tests with values as small as 1.7 [Felzer and Kilb, 2009] gave similar results.

We generate 500 synthetic catalogs by performing 500 ETAS simulations. For each catalog, rates of events with $M > 0.5$ are computed according to the scheme outlined in section 3.2. In ETAS simulations, secondary aftershocks can be readily distinguished from direct aftershocks. Figure 12b presents the rate of secondary aftershocks as a function of time since the main shock, averaged over the 500 synthetic catalogs. In the first few days, the fraction of secondary aftershocks is less than 30%. We use the results presented in Figure 12 to correct the observed rates for the rate of secondary aftershocks. In each time step, the expected fraction of secondary aftershocks is removed from the observations to obtain the rates of direct aftershocks. Using the methodology described in section 3.2, these rates are used to infer stresses, which are then used as input for the joint inversion (section 5.1).

Figure 13 presents the fractional differences between the slip distribution presented in Figure 7 and the slip distribution obtained from inversion in which the input stresses were computed from the rate of direct aftershocks, using the procedure outlined in section 3.2.1, with $\alpha = 1 \times 10^{-2}$ and $\beta = 1 \times 10^{-3}$. The value of α_β is determined iteratively as described in section 5.1. In most areas of the model, and in particular near patches that accumulate substantial postseismic slip, the difference between the two distributions is less than 10%. This small difference validates Dieterich's assumption of noninteracting nucleation sites for the Collins Valley sequence. As shown in section 6.2, it is the spatiotemporal distribution of triggered aseismic slip that dictates the dynamics of the system during aftershock sequences in Anza. Since the amplitude of slip decays rapidly with time, our slip estimates are mostly sensitive to early seismicity rates, in periods where the fraction of secondary aftershocks is small.

6.4. Seismic and Aseismic Strain Release Along the Anza Segment

The cumulative moment due to aseismic fault slip in the 10 days following the El Mayor-Cucapah and the Collins Valley earthquakes is remarkably large compared to the coseismic moment of the Collins Valley earthquake or the total seismic moment of earthquakes during the first 10 days of the El Mayor-Cucapah and Collins Valley aftershock sequences. For large ($M > 7$) earthquakes, equivalent moment of afterslip generally does not exceed 30% of the coseismic moment [Bürgmann et al., 2002; Chlieh et al., 2007; Lin et al., 2013]. However, most

Figure 13. Normalized difference between the aseismic slip distribution in Figure 7b and slip from joint inversion with stresses inferred from direct aftershock rates (Figure 12) computed along profile A-A' (Figure 1). Vertical bars at the top of Figure 13a indicate the approximate along-strike extent of the Hot Springs Cluster (HSC), the Anza Gap (AG), and the Trifurcation Cluster (TC). Differences are normalized by the best fit slip distribution (Figure 7). (a) Spatial distribution. Contours indicate inverted aseismic slip distribution. Star indicates the Collins Valley main shock hypocenter. (b) Histogram of normalized residuals.

geodetic studies exclude the first day of postseismic deformation and so may underestimate the moment associated with aseismic slip. Some studies have reported examples where afterslip following a smaller earthquake was much larger than the coseismic moment. For example, the moment due to afterslip following the 2004 M_w 6.0 Parkfield earthquake exceeded the coseismic moment after 9 months [Langbein *et al.*, 2006] and was twice as large as the coseismic moment after 5 years [Bruhat *et al.*, 2011]. Murray and Segall [2005] argued that the aseismic moment of episodic slip accompanying three $M \approx 4.5$ Parkfield main shocks was 10 times larger than the combined seismic moment of the triggering main shocks. Yagai and Ozawa [2013] found that moment due to afterslip over an 8 year interval following two events with M_w 6.8 and M_w 6.7 in the Hyuga-Nada area, southwest Japan, was about three times larger than the sum of the moments of the two main shocks. Thomas *et al.* [2014] studied seismic and aseismic slip along the southern segment of the Longitudinal Valley Fault (LVF) in Taiwan, which is known to deform mostly aseismically and can also produce $M > 6.5$ events. They found that moment due to afterslip that occurred in the first year around the rupture area of the 2003 M_w 6.8 Chengkung earthquake was equivalent to 80% of the coseismic moment. These sequences occurred on faults that are known to exhibit steady state creep interseismically (northwest of Parkfield or the southern segment of the LVF) or that accommodate large episodic slow-slip events (southwest Japan). If fault strength is rate-and-state dependent and if the frictional response of patches that slip aseismically is close to velocity neutral, one can expect moderate stress perturbations to trigger large amplitude aseismic slip [e.g., Perfettini and Ampuero, 2008].

The peak stress drops during the transient slip events in Anza are of the order of a few megapascal (Figures 9c and 9f), about a hundred times larger than stress drops during episodic slow-slip events in Cascadia and southwest Japan [Schmidt and Gao, 2010; Sekine *et al.*, 2010]. It is not surprising that our peak stress drops are larger than those inferred from geodetic-only inversions. The incorporation of seismicity data in the inversion results in rough slip distributions, larger slip, and hence larger stress drops than the ones obtained from a geodetic-only inversion. For example, Bartlow *et al.* [2014] used both a geodetic-only inversion and a joint inversion approach to constrain the slip distribution during a transient slow-slip event in the Hikurangi subduction zone. The peak stress drop that they obtained from the joint inversion was 2–3 times larger than the one they obtained from a geodetic-only inversion and 10 times smaller than the stress drops we obtain in this study. Additionally, because the static stress drop in the dislocation model is inversely proportional to the dislocation width, if the widths of regions thought to be hosting episodic slow slip in subduction zone have been systematically overestimated, then their real stress drops are actually higher. One other factor which likely contributes toward the high stress drops we observe concerns the stress level at which seismicity in Anza is driven. Triggering of seismicity in Anza seems to require stresses that are of the order of several megapascal, much larger than the stresses required to trigger tremor. The analysis of the M_w 7.2 El Mayor-Cucapah sequence suggests that events in Anza were triggered mainly due stresses transferred from nearby aseismically slipping fault segments rather than due to the (much smaller) dynamic stress changes associated with the surface waves of this event. The lack of observed correlation between ocean tides and seismicity in Anza (and in most other faults in southern California) indicates that the stress levels required to drastically change seismicity rates in that region are in the range of several hundreds kilopascal to a few megapascal. In the

presence of large scale aseismic slip, sufficiently high static stresses changes driving seismogenic faults are provided when the stress drop within the aseismically slipping region is large.

Aseismic slip triggered by the Collins Valley main shock peaks between 20 and 25 km and near 50 km along profile A-A' (Figure 7). This pattern is anticorrelated with the largest stress concentrations imposed by the early (10 days) postseismic slip due to the El Mayor-Cucapah earthquake (Figure 9c). One interpretation of this observation is that the stress field due to aseismic slip triggered by the El Mayor-Cucapah main shock determined the spatial extent of the aseismic slip triggered by the Collins Valley main shock. This is supported by the results presented in Figure 9c, which show that the early (10 days) aseismic slip triggered by the El Mayor-Cucapah main shock increased the stresses in the area that later hosted the Collins Valley hypocenter and its nearest aseismic slip and that the southeast termination (30 km along profile A-A' in Figure 1) of aseismic slip triggered by the Collins Valley main shock is adjacent to a patch on which the aseismic slip triggered by the El Mayor-Cucapah main shock left negative stresses. An alternative interpretation is that the spatial distribution of triggered aseismic slip is the result of strong heterogeneity of frictional properties within the transition zone.

7. Conclusions

The El Mayor-Cucapah earthquake initiated a several monthlong transient along the central section of the SJF near Anza. This is manifested by high seismicity rates and periods of elevated surface strain rates. We analyzed two periods of particularly intense seismicity and high surface strain rates. The first, which lasted approximately 10 days after the El Mayor-Cucapah earthquake, is associated with aseismic slip along the SJF with moment magnitude of 5.9. We observe only weak evidence for migration of seismicity during this sequence. The second initiated with the Collins Valley earthquake, which occurred 94 days after the El Mayor-Cucapah earthquake. The Collins Valley earthquake triggered aftershocks that migrated along the SJF strike and triggered aseismic slip with moment magnitude of 5.8.

The joint inversion of strain and seismicity data allows us to resolve slip on deep segments. We model the joint data set by assuming that slip on the main fault strand triggers seismicity on secondary faults. Our approach does not account for interactions between aftershocks. We find, however, that the effect of earthquake interactions on the slip distribution is small. This is most likely because the moment due to aseismic slip is much larger than the seismic moment, and thus, aseismic fault slip is the dominant driver of seismicity, especially in the more seismically active segment southeast of the Anza Seismic Gap.

The joint data set requires aseismic slip to occur at the edges of the Hot Springs and Trifurcation clusters, two large clusters of seismicity adjacent to the Anza Seismic Gap, and beneath the gap itself, at depths of 10 to 17 km. The distribution of triggered aseismic slip following the El Mayor-Cucapah and Collins Valley earthquakes is complementary, but the result is sensitive to our modeling assumptions. Because aseismic slip is more important than aftershocks for generating stresses around the Anza segment, this spatial pattern implies that stresses due to the El Mayor-Cucapah aseismic slip on the SJF triggered the Collins Valley earthquake and facilitated the large extent of its aseismic slip. Observations supporting this conclusion are that the Collins Valley earthquake initiated in the area of peak stress left by the El Mayor-Cucapah aseismic slip and that the Collins Valley aseismic slip is spatially correlated with stresses imposed by the first 10 days of El Mayor-Cucapah aseismic slip.

We find that aseismic slip generated by El Mayor-Cucapah earthquake propagated rapidly in the first few days and then more slowly, as evidenced by the decaying surface strain rates. The segment hosting the largest triggered aseismic slip continued slipping at rates that exceed the long-term slip rates, thus stressing nearby fault segments and producing seismicity at a rate 10 times larger than the long-term background seismicity rate. This loading most likely triggered the Collins Valley earthquake, which in turn fed back deep aseismic slip. The sequence as a whole illustrates the dynamics that can arise from coupling between seismic and aseismic slips. The whole sequence must have increased shear stresses on the locked portion of the SJF in the Anza Seismic Gap.

Appendix A: Error of Inferred Stresses

Uncertainties in aftershock rates (and hence inferred stresses) may result from several factors. One source of bias may be mislocated aftershocks contaminating the event counts within seismic cells. The location uncertainties can be separated into absolute and relative error. We account for absolute location uncertainty by

Figure A1. Stress errors from synthetic tests. (a) Earthquake rate as a function of time since the main shock. Grey curves are estimated from synthetic catalogs whose interevent times are drawn from a nonstationary Poissonian distribution with prescribed stress history. Blue and magenta curves are the prescribed rates computed with $\dot{N}_{bg} = 10^{-1}$ and $\dot{N}_{bg} = 10^{-2}$ earthquakes/day, respectively, and with $\dot{N}(0^+)/\dot{N}_{bg} = 10^6$ and $\alpha = 10^{-3}$. (b) Stress error computed as the mean difference between stresses inferred from synthetic catalogs and the actual stresses, normalized by the latter. Colors have the same meaning as in Figure A1a.

coupling the least squares algorithm with a Monte Carlo scheme in which the locations of seismic cells containing aftershocks are perturbed before each inversion. The relative location uncertainty is smaller than the dimension of the cells and, since the inferred stresses are assumed to be representative of the stress in the reference position of the cell, shuffling event locations within a given cell would not change the estimated stresses.

Two other sources of bias arise from our aftershock counting procedure. The first is the possible inclusion of secondary aftershocks, which are the result of earthquake interactions that are not accounted for in Dieterich's model. We show in section 6.2 that the effect of these secondary aftershocks on the output slip distribution is small. A second source of bias is related to the number of events used to evaluate seismicity rates in each time window. Our rate estimation scheme is based initially on only five aftershocks and later on up to 10 aftershocks. Our objective in this section is to estimate the error on rates computed with this technique. We propagate the error on rates in our stress computations (equations (2) and (4)) to obtain uncertainties on the inferred stress used in our inversion (section 5.1).

We model aftershocks as nonstationary Poissonian processes with time-dependent rate $\lambda(t)$. As a transient scenario we consider an initial constant stressing rate, $\dot{\tau}_{tect}$, followed by a stress step, $\Delta\tau$ at $t = 0$, and then

by a different constant loading rate, $\dot{\tau}$. The resulting evolution of seismicity rate in the rate-and-state model is [Dieterich, 1994]:

$$\dot{N}(t) = \frac{\dot{N}_{bg} \dot{\tau} / \dot{\tau}_{tect}}{\left[\frac{\dot{\tau}}{\dot{\tau}_{tect}} \exp\left(\frac{-\Delta\tau}{a\sigma}\right) - 1 \right] \exp\left(\frac{-t}{t_a}\right) + 1}, \quad (A1)$$

where $\dot{N}_{bg}$ is the initial (steady) seismicity rate, a is a constitutive parameter of rate-and-state friction, σ is the effective normal stress, and t_a is the characteristic duration of the aftershock sequence, which is close to 10 days for the Anza sequences. The ratio $\dot{\tau} / \dot{\tau}_{tect}$ is set to 10. The instantaneous change in seismicity rate, $\dot{N}(0^+) / \dot{N}_{bg}$, is related to the stress step by

$$\Delta\tau = a\sigma \ln\left(\frac{\dot{N}(0)}{\dot{N}_{bg}}\right). \quad (A2)$$

We use this equation to prescribe $\Delta\tau$ needed to achieve a target $\dot{N}(0^+) / \dot{N}_{bg}$ in our scenarios. We construct synthetic catalogs by simulating a nonstationary Poisson process with this prescribed rate (Figure A1a). To assess the sensitivity of the computed rates to the number of events in the simulated catalogs, we vary the value of $\dot{N}_{bg}$ between 0.01 and 1 earthquakes/day, similar to the range of observed background rates in the Anza catalog (Figure 3). Our estimate of the ratio $\dot{N} / \dot{N}_{bg}$ is taken as the median value of the observed rate in each cell, and the value of a spans the range $1 \times 10^{-6} - 1 \times 10^{-1}$ used in the inversion procedure. We estimate stress time histories from the rates of each simulated sequence, following the same procedure as for the real data. We then measure the final stress values and compare them to the exact stress step amplitude obtained via equation (A2).

Figure A1b presents the average stress error as a function of time since the main shock using 100 simulated catalogs. We find that uncertainties on stresses associated with our rate estimations are highly time dependent and are highest early in the aftershock sequence when the number of events in the rate estimation window is smallest. The stress error increases to up to about 12–15% of the cumulative stresses near the end of the period of analysis (10 days).

Note that stress errors are shown for synthetic tests in which $\dot{N}_{bg}$ is equal to 0.01 and 0.1 events per day. The uncertainty of the stress measurements (for a given value of $a\sigma$) is obtained by binning the seismic cells according to their background rates (Figure 3a) and the median value of the measured ratio $\dot{N} / \dot{N}_{bg}$, which is used to determine the stress change driving seismicity.

The analysis presented in section 4 suggests that the response at the HSC was delayed by about 1 day, which is approximately the time it takes for the propagating creep front to arrive near that area. To mimic this behavior, we conducted the same statistical analysis on synthetic catalogs in which the main shock time is delayed by 1 day. We found that delaying the main shock time had little effect on the errors presented in Figure A1b.

Appendix B: Alternative Model for Stress Transfer to Seismically Active Cell

The geodetic and seismic data sets allow us to assess whether aftershock seismicity is dominated by stress transfer to off-fault or to on-fault sites. The joint inversion of strain and aftershock data, whose results are presented in Figure 7, was performed under the assumption that transient slip on a main fault strand triggered seismicity at off-fault sites (i.e., secondary faults off the main strand). In this section we examine an alternative mechanism for static stress transfer to aftershock nucleation sites on the main fault strand.

On several well-studied faults, earthquakes tend to occur repeatedly in the same location. These so-called repeaters are composed of tightly clustered microearthquakes, with nearly identical waveforms. Among continental transforms, they are most commonly observed on the Parkfield section of the San Andreas Fault, where their locations well delineate the main fault strand along which they occur. A physical mechanism for explaining repeating earthquakes, which is capable of reproducing observed scaling relations between earthquake recurrence interval and moment, is one in which frictional instabilities nucleate on velocity-weakening patches that are loaded by creep in the surrounding, velocity-strengthening area [e.g., Chen and Lapusta, 2009]. Near the Anza segment, however, repeating earthquakes are not a common feature. Anza seismicity is diffuse and extends out to about 3 km on both sides of the fault (Figures 1 and S3). Such distribution may be the result of geometric irregularities that are an inherent feature of immature fault

Figure B1. Regularization term C_2 as a function of the joint misfit C_1 for inversions assuming on-fault aftershocks. Solid black and blue curves are for $\Omega/K_{kk} = 10$ and $\Omega/K_{kk} = 100$, respectively. Dashed black and blue curves are for $a = 10^{-5}$ and $a = 10^{-3}$, respectively.

zones [Powers and Jordan, 2010]. Nonetheless, we wish to exploit the available strain and aftershock data to test the hypothesis that seismic and aseismic patches share the same fault plane. This description is a priori admissible given that the absolute horizontal location uncertainty is about 2 km, which is in the range of distances between most of the aftershocks and the model fault plane (Figure S3).

We consider a model in which seismically active cells are coplanar with the fault surface. We assume that aseismic slip occurs over most of the cell's area and is the primary source of loading on frictionally unstable asperities contained within the cell. The average stress on an asperity with radius R increases due to aseismic slip u outside the asperity by an amount proportional to $\mu u/R$. The stress on the k th cell is the sum of the average stress on N locked asperities located within that cell and stresses due to slip on all other model cells:

$$\Delta \tau_k = \Omega u_k + \sum_{j \neq k} K_{jk} u_j, \quad (\text{B1})$$

where the stiffness of a cell containing N seismic asperities is

$$\Omega = \frac{\mu}{N} \sum_i \frac{1}{R_i} + K_{kk}, \quad (\text{B2})$$

where R_i is the radius of the i th asperity and μ is the shear modulus. The first term on the right-hand side of equation (B2) is positive. The second term, the self-stiffness K_{kk} of an asperity-free cell, is negative and scales as $-\mu/\Delta x$, where Δx is the shortest cell dimension. For the magnitude range in our catalog, assuming a circular crack model with 3 MPa stress drop, individual asperities are expected to be roughly an order of magnitude smaller than the shortest cell dimension. Since $R_i \ll \Delta x$, the first term dominates and Ω is positive. We determine a uniform value of Ω (the same for all cells) that minimizes the cost function defined by equation (11), through grid search. Setting $a = 10^{-2}$, we find that the best fitting model requires $\Omega \sim 10K_{kk}$ (Figure B1). Our estimate of Ω is consistent with our modeling assumptions.

For large values of Ω ($> 100K_{kk}$), slightly better results are obtained when the value of a is increased to 10^{-1} . Note, however, that the values of Ω and a cannot be arbitrarily large. In the Collins Valley models, using $\Omega > 100$ and $a > 1$ results in output stresses that exceed the shear strength of the media. With larger values ($\Omega \approx 100$ and $a = 10^{-1}$), the misfit is still larger than the misfit of our preferred model developed in the main text (Figure 6). This indicates that the off-fault aftershock model is more probable than the on-fault aftershock model.

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