

1                   **Rheologic constraints on the upper mantle from five years of**  
2                   **postseismic deformation following the El Mayor-Cucapah**  
3                   **earthquake**

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6                   **Key Points:**

- 7                   • Transient postseismic deformation can be observed following the El Mayor-Cucapah  
8                   earthquake at epicentral distances of up to 400 km  
9                   • Near-field postseismic deformation exhibits early transience that decays to a sustained  
10                  rate which is elevated above the preseismic trend  
11                  • Far-field postseismic deformation can be explained with a Zener or Burgers rheology  
12                  upper mantle

13 **Abstract**

14 We analyze five years of Southern California GPS data following the Mw=7.2 El Mayor-Cucapah  
 15 earthquake. We observed transient postseismic deformation which persists for three years at  
 16 epicentral distances greater than  $\sim 200$  km. In the near-field, rapid postseismic transience de-  
 17 cays to a sustained rate which exceeds its preseismic trend. We attempt to determine the mech-  
 18 anisms driving this deformation, where we consider afterslip at seismogenic depths and vis-  
 19 coelastic relaxation in the lower crust and upper mantle as candidate mechanisms. We find that  
 20 early, rapid, near-field deformation can be explained with afterslip on the fault that ruptured  
 21 coseismically, while the later, sustained, near-field deformation can be explained with either  
 22 continued afterslip in an effectively elastic lower crust, or lower crustal viscoelastic relaxation  
 23 with a steady-state viscosity of  $\sim 10^{19}$  Pa s. The trend in far-field deformation is best explained  
 24 with a transient viscosity of  $\sim 10^{18}$  Pa s in the upper mantle. We argue that a transient rhe-  
 25 ology in the mantle is preferable over a Maxwell rheology because it better predicts the de-  
 26 cay in postseismic deformation, and also because it does not conflict with the generally higher,  
 27 steady-state viscosities inferred from studies of geophysical processes occurring over longer  
 28 time-scales.

29 **1 Introduction**

30 Ground deformation in the years following a large ( $Mw \gtrsim 7$ ) earthquake can be used to  
 31 gain insight into the mechanical behavior of the crust and upper mantle. The interpretations  
 32 of postseismic deformation are not always conclusive because multiple postseismic deforma-  
 33 tion mechanisms, such as afterslip or viscoelastic relaxation in the lower crust and upper man-  
 34 tle, can have qualitatively similar surface expressions [e.g. Savage, 1990]. This non-uniqueness  
 35 complication can potentially be remedied if the postseismic deformation occurs in an area that  
 36 is sufficiently well instrumented with GPS stations [Hearn, 2003]. Owing to the dense geode-  
 37 metic network deployed throughout the 2000s as part of the Plate Boundary Observatory, the post-  
 38 seismic deformation following the April 4, 2010, Mw=7.2 El Mayor-Cucapah earthquake in  
 39 Baja California was observed at more GPS stations than any other earthquake in California  
 40 to date (see Hauksson *et al.* [2011] and Fletcher *et al.* [2014] for a detailed description of this  
 41 earthquake and its seismotectonic context). With such a large collection of data, we attempt  
 42 to discern the mechanisms driving the postseismic deformation.

43 Previous studies which have modeled postseismic deformation following the El Mayor-  
 44 Cucapah earthquake include Pollitz *et al.* [2012], Gonzalez-Ortega *et al.* [2014], Spinler *et al.*  
 45 [2015], and Rollins *et al.* [2015]. Of these studies, Gonzalez-Ortega *et al.* [2014] and Rollins  
 46 *et al.* [2015] have attempted to describe the postseismic deformation with afterslip in an elas-  
 47 tic half-space. Gonzalez-Ortega *et al.* [2014] described five months of postseismic deforma-  
 48 tion, observed by InSAR and GPS stations within  $\sim 50$  km of the rupture, with afterslip and  
 49 contraction on the coseismically ruptured fault. Gonzalez-Ortega *et al.* [2014] noted that their  
 50 preferred model underestimated the GPS displacements for stations  $\gtrsim 25$  km from the rup-  
 51 ture and suggested that it could be the result of unmodeled viscoelastic relaxation. Using only  
 52 continuous GPS stations, which are mostly north of the rupture zone, Rollins *et al.* [2015] found  
 53 that three years of postseismic deformation can be adequately explained by afterslip, albeit with  
 54 an implausibly large amount of slip inferred on the least constrained, southern-most fault seg-  
 55 ment. Here, we suggest the afterslip inferred by Rollins *et al.* [2015] may have been acting as  
 56 a proxy for distributed relaxation in the upper mantle.

57 Pollitz *et al.* [2012], Rollins *et al.* [2015] and Spinler *et al.* [2015] have explored viscoelas-  
 58 tic relaxation in the lower crust and upper mantle as a potential postseismic deformation mech-  
 59 anism. The rheology of the crust and mantle is largely unknown and so modeling postseis-  
 60 mic deformation with viscoelastic relaxation requires one to assume a rheologic model and  
 61 then find the best fitting rheologic parameters. The inference of these rheologic parameters is  
 62 a computationally expensive non-linear inverse problem which is typically approached with  
 63 a forward modeling grid search method. Consequently, a simplified structure for the Earth must

be assumed in order to minimize the number of rheologic parameters that need to be estimated. For example, it is commonly assumed that the lower crust and upper mantle are homogeneous, Maxwell viscoelastic layers, which may be too simplistic for postseismic studies [Riva and Govers, 2009; Hines and Hetland, 2013]. To further reduce the dimensions of the model space, it is also necessary to make simplifying assumptions about the behavior of afterslip. For example, one can assume a frictional model for afterslip and parametrize afterslip in terms of the unknown rheologic properties of the fault [e.g. Johnson *et al.*, 2009; Johnson and Segall, 2004]. One can also assume that afterslip does not persist for more than a few months and then model the later postseismic deformation assuming it to be the result of only viscoelastic relaxation [e.g. Pollitz *et al.*, 2012; Spinler *et al.*, 2015]. However, afterslip in similar tectonic settings has been observed to persist for decades following earthquakes [Cakir *et al.*, 2012; Cetin *et al.*, 2014]. Indeed, the preferred viscoelastic model from Pollitz *et al.* [2012] significantly underestimates deformation in the Imperial Valley, which could be indicative of unmodeled continued afterslip. Neglecting to allow for sustained afterslip as a postseismic mechanism could then lead to biased inferences of viscosities.

In this study, we perform a kinematic inversion for fault slip, allowing it to persist throughout the postseismic period, while simultaneously estimating the viscosity of the lower crust and upper mantle. We create an initial model of the fault slip and effective viscosity necessary to describe early postseismic deformation using the method described in Hines and Hetland [2016]. This method uses a first-order approximation of surface deformation resulting from viscoelastic relaxation which is only applicable to the early postseismic period. In this case, our initial model describes the first 0.8 years of postseismic deformation following the El Mayor-Cucapah earthquake. We then use the inferred effective viscosity structure from the initial model to create a suite of postseismic models to explain the five years of postseismic data available to date. Of the suite of models tested, we find that postseismic deformation following the El Mayor-Cucapah earthquake can be explained with a combination of afterslip on a fault segment running through the Sierra Cucapah and viscoelastic relaxation in a Zener rheology upper mantle with a transient viscosity on the order of  $10^{18}$  Pa s.

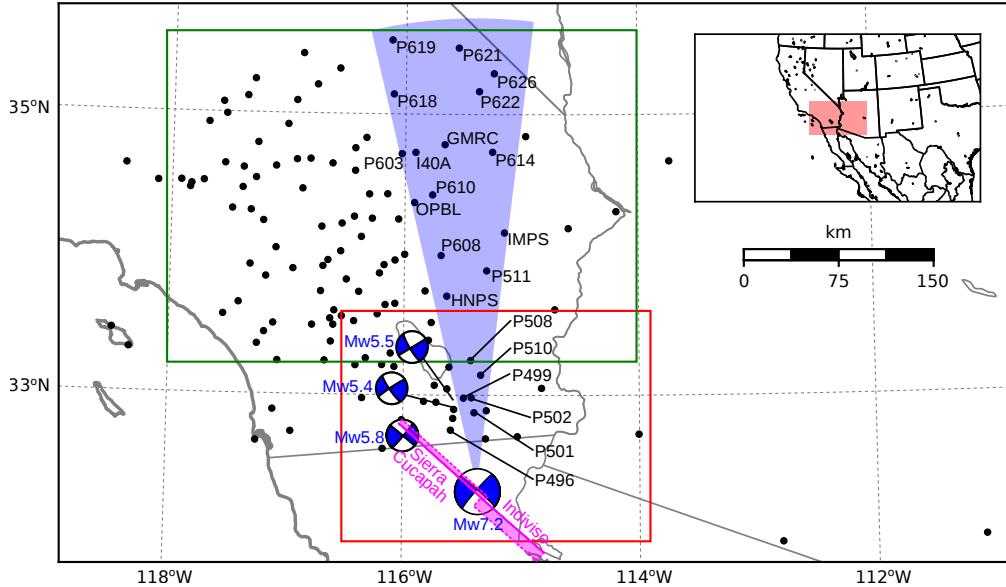
## 2 Data Processing

We use continuous GPS position time series provided by University Navstar Consortium (UNAVCO) for stations within a 400 km radius about the El Mayor-Cucapah epicenter. We collectively describe the coseismic and postseismic displacements resulting from the El Mayor-Cucapah earthquake as  $u_{\text{post}}(t)$ . We consider the GPS position time series,  $u_{\text{obs}}(t)$ , to be the combination of  $u_{\text{post}}(t)$ , secular tectonic deformation, annual and semi-annual oscillations, and coseismic offsets from significant earthquakes over the time span of this study. The June 14, 2010, Mw=5.8 Ocotillo earthquake and the Brawley swarm, which included an Mw=5.5 and an Mw=5.4 event on August 26, 2012 (Figure 1), are the only earthquakes that produced noticeable displacements in any of the time series. We treat the displacements resulting from the Brawley swarm as a single event because the daily solutions provided by UNAVCO cannot resolve the separate events. Although the Ocotillo earthquake had its own series of aftershocks [Hauksson *et al.*, 2011], neither the Ocotillo earthquake nor the Brawley swarm produced detectable postseismic deformation. We model displacements resulting from these events with only a Heaviside function,  $H(t)$ , describing the coseismic offsets. We then model  $u_{\text{obs}}(t)$  as

$$u_{\text{obs}}(t) = u_{\text{pred}}(t) + \epsilon, \quad (1)$$

where

$$\begin{aligned} u_{\text{pred}}(t) = & u_{\text{post}}(t)H(t - t_{\text{emc}}) + c_0 + c_1 t + \\ & c_2 \sin(2\pi t) + c_3 \cos(2\pi t) + c_4 \sin(4\pi t) + c_5 \cos(4\pi t) + \\ & c_6 H(t - t_{\text{oc}}) + c_7 H(t - t_{\text{bs}}). \end{aligned} \quad (2)$$



92 **Figure 1.** Map of the region considered in this study. The large focal mechanism is the GCMT solution  
 93 for the El Mayor-Cucapah earthquake, and the three small focal mechanisms are for the Ocotillo earthquake  
 94 and the two main shocks during the Brawley swarm. The black dots indicate the locations of GPS stations  
 95 used in this study. The fault geometry used in this study is shown in magenta where dashed lines indicate  
 96 buried edges of the fault segments. The green and red boxes demarcate the extent of the near-field and  
 97 far-field maps (Figures 4 and 5). Stations inside the blue sector, which highlights the area within  $10^{\circ}$  of the El  
 98 Mayor-Cucapah P-axis, are used in Figures 7 and 10.

115 In the above equations,  $t_{\text{emc}}$ ,  $t_{\text{oc}}$  and  $t_{\text{bs}}$  are the times of the El Mayor-Cucapah earthquake,  
 116 Ocotillo earthquake, and the Brawley swarm, respectively,  $c_0$  through  $c_7$  are unknown coef-  
 117 ficients, and  $\epsilon$  is the observation noise. We are using years as our unit of time which makes  
 118  $c_2$  through  $c_5$  the coefficients for annual and semi-annual oscillations. We only estimate jumps  
 119 associated with the Ocotillo earthquake and Brawley swarm for stations within 40 km of their  
 120 epicenters.

121 Stations which recorded displacements that clearly cannot be described by the aforemen-  
 122 tioned processes are not included in our analysis. This includes stations in the Los Angeles  
 123 basin, where anthropogenic deformation can be larger than the postseismic signal that we are  
 124 trying to estimate [Bawden *et al.*, 2001; Argus *et al.*, 2005]. In order to ensure an accurate es-  
 125 timation of the secular deformation, we only use stations that were installed at least six months  
 126 prior to El Mayor-Cucapah earthquake even though several GPS stations were installed after  
 127 the earthquake to get better coverage of the postseismic deformation field [Spiner *et al.*, 2015].  
 128 It would be possible to subtract secular velocities derived from elastic block models [e.g. Meade  
 129 and Hager, 2005] from velocities recorded at the newly installed stations to get an estimate  
 130 of postseismic velocities at those stations. However, estimating velocities from an already noisy  
 131 displacement time series can introduce significant uncertainties depending on exactly how the  
 132 estimation is done. We therefore use coseismic and postseismic displacements, rather than ve-  
 133 locities, in our inverse method described in Section 3. This choice prevents us from using the  
 134 newly installed stations for our analysis.

135 The October 16, 1999, Mw=7.1 Hector Mine earthquake, which occurred  $\sim$ 270 km north  
 136 of the El Mayor-Cucapah epicenter, produced transient postseismic deformation which we do  
 137 not wish to model, either mechanically or through empirical line fitting. We thus restrict our  
 138 analysis to deformation observed six years after the Hector Mine earthquake, which is when

139 postseismic velocities at sites near the Hector Mine epicenter are approximately constant [Savage  
 140 and Svare, 2009]. When appraising our model fit in Section 3, we see some systematic  
 141 residuals in the vicinity of the Hector Mine epicenter, which may be the result of errors in the  
 142 assumption that the trend in Hector Mine postseismic deformation is linear after six years.

143 Studies of postseismic deformation typically assume a parametric form for  $u_{\text{post}}(t)$ , such  
 144 as one with a logarithmic or exponential time dependence [e.g. Savage *et al.*, 2005]. However,  
 145 by assuming a logarithmic or exponential form of  $u_{\text{post}}(t)$  we run the risk of over fitting the  
 146 GPS time series and inferring a non-existent postseismic signal. We therefore do not assume  
 147 any parametric form for  $u_{\text{post}}(t)$  and rather treat it as integrated Brownian motion, so that

$$\dot{u}_{\text{post}}(t) = \sigma^2 \int_0^t w(s)ds, \quad (3)$$

148 where  $w(t)$  is white noise and the variance of  $\dot{u}_{\text{post}}(t)$  increases linearly with time by a factor  
 149 of  $\sigma^2$ . We use a Kalman filtering approach to estimate  $u_{\text{post}}(t)$  and the unknown parameters  
 150 in eq. (2). In the context of Kalman filtering, our time varying state vector is

$$\mathbf{X}(t) = [u_{\text{post}}(t), \dot{u}_{\text{post}}(t), c_0, \dots, c_7] \quad (4)$$

151 and eq. (2) is the observation function which maps the state vector to the GPS observations.  
 152 We initiate the Kalman filter by assuming a prior estimate of  $\mathbf{X}(t)$  at the first time epoch, denoted  
 153  $\mathbf{X}_{1|0}$ , which has a sufficiently large covariance, denoted  $\Sigma_{1|0}$ , to effectively make our  
 154 prior uninformed. For each time epoch,  $t_i$ , Bayesian linear regression is used to incorporate  
 155 GPS derived estimates of displacement with our prior estimate of the state,  $\mathbf{X}_{i|i-1}$ , to form  
 156 a posterior estimate of the state,  $\mathbf{X}_{i|i}$ , which has covariance  $\Sigma_{i|i}$ . We then use the posterior  
 157 estimate of the state at time  $t_i$  to form a prior estimate of the state at time  $t_{i+1}$  through the  
 158 transition function

$$\mathbf{X}_{i+1|i} = \mathbf{F}_{i+1} \mathbf{X}_{i|i} + \delta_{i+1}, \quad (5)$$

159 where

$$\mathbf{F}_{i+1} = \begin{bmatrix} 1 & (t_{i+1} - t_i) & \mathbf{0} \\ 0 & 1 & \mathbf{0} \\ \mathbf{0} & \mathbf{0} & \mathbf{I} \end{bmatrix} \quad (6)$$

160 and  $\delta_{i+1}$  is the process noise, which has zero mean and covariance described by

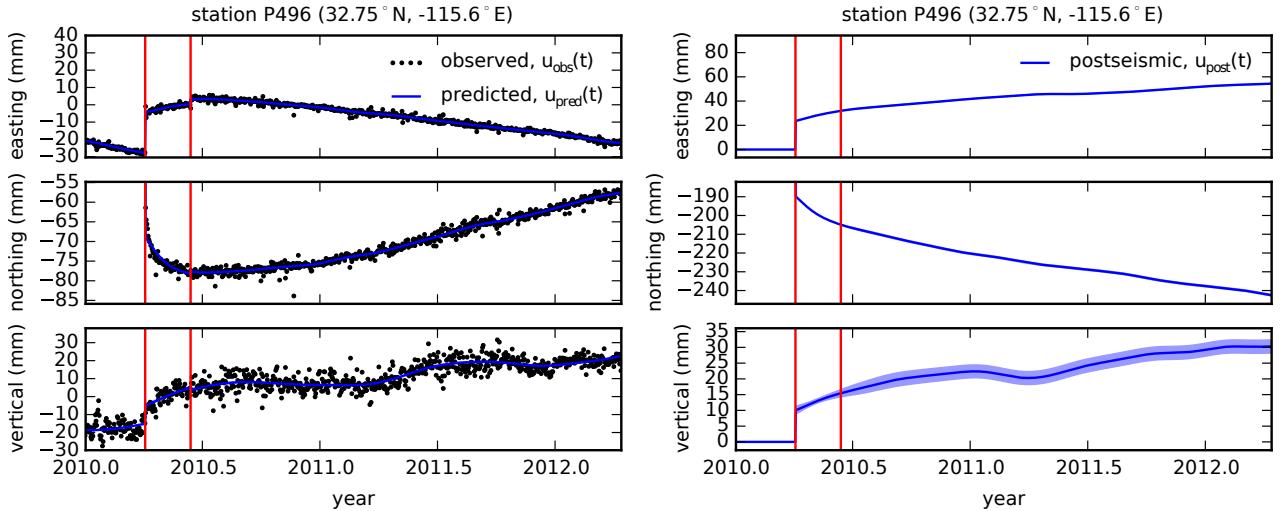
$$\mathbf{Q}_{i+1} = \sigma^2 \begin{bmatrix} \frac{(t_{i+1}-t_i)^3}{3} & \frac{(t_{i+1}-t_i)^2}{2} & \mathbf{0} \\ \frac{(t_{i+1}-t_i)^2}{2} & (t_{i+1} - t_i) & \mathbf{0} \\ \mathbf{0} & \mathbf{0} & \mathbf{0} \end{bmatrix}. \quad (7)$$

161 The covariance of the new prior state,  $\mathbf{X}_{i+1|i}$ , is then described by

$$\Sigma_{i+1|i} = \mathbf{F}_{i+1} \Sigma_{i|i} \mathbf{F}_{i+1}^T + \mathbf{Q}_{i+1}. \quad (8)$$

162 This process is repeated for each of the  $N$  time epochs at which point we use Rauch-Tung-  
 163 Striebel smoothing [Rauch *et al.*, 1965] to find  $\mathbf{X}_{i|N}$ , which is an estimate of the state at time  
 164  $t_i$  that incorporates GPS observation for all  $N$  time epochs. Our final estimates of  $u_{\text{post}}(t)$   
 165 are used in subsequent analysis, while the remaining components of the state vector are con-  
 166 sidered nuisance parameters. In the interests of computational tractability, we down sample  
 167 our smoothed time series from daily solutions down to weekly solutions.

168 The smoothness of  $u_{\text{post}}(t)$  is controlled by the chosen value of  $\sigma^2$ , which describes how  
 169 rapidly we expect postseismic displacements to vary over time. Setting  $\sigma^2$  equal to zero will

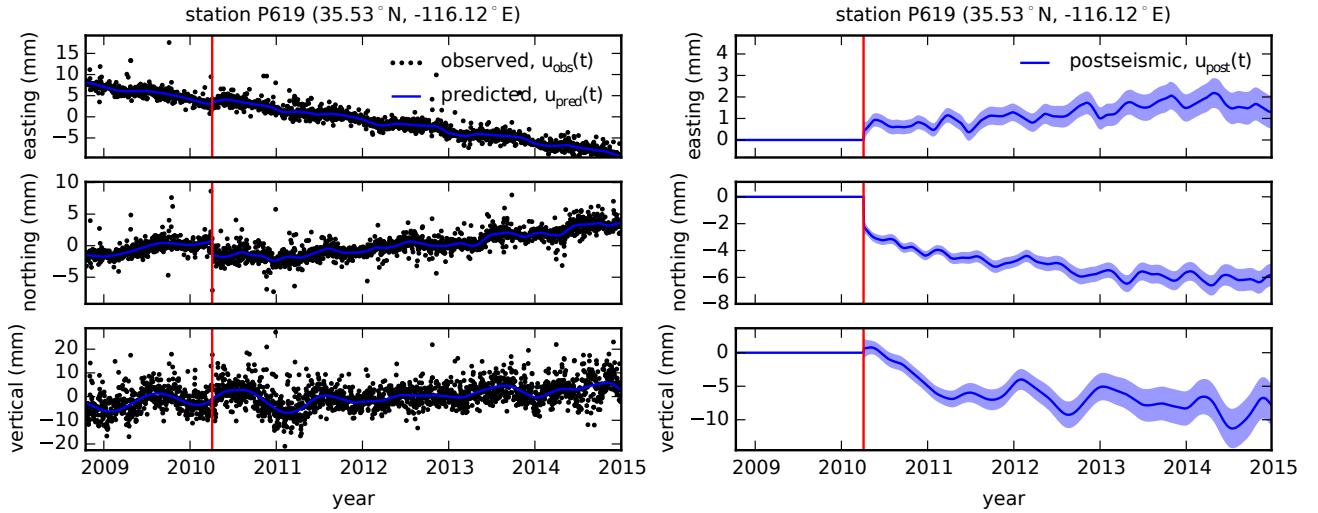


186 **Figure 2.** Left panels show GPS time series from UNAVCO (black) and the predicted displacement (blue)  
 187 from eq. (2) for a near-field station. Red lines indicate the times of the El Mayor-Cucapah and Ocotillo earth-  
 188 quake. The right panels show estimated coseismic and postseismic displacements,  $u_{\text{post}}$ , which are extracted  
 189 from the predicted displacements. The 68% confidence interval is shown in light blue.

170 effectively result in modeling  $u_{\text{post}}(t)$  as a straight line which is insufficient to describe the  
 171 expected transient behavior in postseismic deformation. The other end member, where  $\sigma^2$  is  
 172 infinitely large, will result in  $u_{\text{pred}}(t)$  over fitting the data. While one can use a maximum like-  
 173 lihood based approach for picking  $\sigma^2$  [e.g. Segall and Mathews, 1997], we instead take a sub-  
 174 jective approach and choose a value for  $\sigma^2$  that is just large enough to faithfully describe the  
 175 observed deformation at the most near-field station in our study, P496, which exhibits the most  
 176 rapid changes in velocity. This ensures that  $\sigma^2$  will be sufficiently large so that our estimate  
 177 of  $u_{\text{post}}(t)$  does not smooth out potentially valuable postseismic signal at the remaining sta-  
 178 tions. We find that using  $\sigma^2 = 0.05 \text{ m}^2/\text{yr}^3$  adequately describes all but the first week of post-  
 179 seismic deformation at station P496, which slightly increases our estimate of coseismic dis-  
 180 placements (Figure 2). We include an example of estimating  $u_{\text{post}}(t)$  for a far-field station,  
 181 P619, which is about 359 km north of the El Mayor-Cucapah epicenter (Figure 3). At station  
 182 P619, along with all the other stations in the Mojave region, there is a south-trending post-  
 183 seismic transience that persists for the first three years after the El Mayor-Cucapah earthquake.  
 184 Postseismic deformation that extends to these epicentral distances has also been observed af-  
 185 ter the Hector Mine earthquake [Freed et al., 2007].

191 It is important to note that the shown uncertainties in  $u_{\text{post}}(t)$  do not account for the non-  
 192 negligible epistemic uncertainty in eq. (2). For example, we assume a constant rate of secu-  
 193 lar deformation, which appears to be an appropriate approximation for all but perhaps the sta-  
 194 tions closest to the Hector Mine epicenter, as noted above. Also, our model for seasonal de-  
 195 formation in eq. (2) assumes a constant amplitude over time, which means that any yearly vari-  
 196 ability in the climatic conditions could introduce systematic residuals [Davis et al., 2012]. In-  
 197 deed, it would be more appropriate to consider the seasonal amplitudes  $c_2 - c_5$  in eq. (2) as  
 198 stochastic variables [Murray and Segall, 2005]. By using constant seasonal amplitudes, our es-  
 199 timate of  $u_{\text{post}}(t)$  seems to describe some of the unmodeled annual and semi-annual oscilla-  
 200 tions (e.g. Figure 3).

201 We show in Figures 4 and 5 the near and far-field postseismic displacements accumu-  
 202 lated over the time intervals 0–1 years, 1–3 years, and 3–5 years, as well as the coseismic dis-  
 203 placements. Stations at epicentral distances beyond  $\sim 200$  km have an elevated rate of defor-  
 204 mation for the first three years following the earthquake. This far-field deformation is trend-



**Figure 3.** same as Figure 2 but for a far-field station.

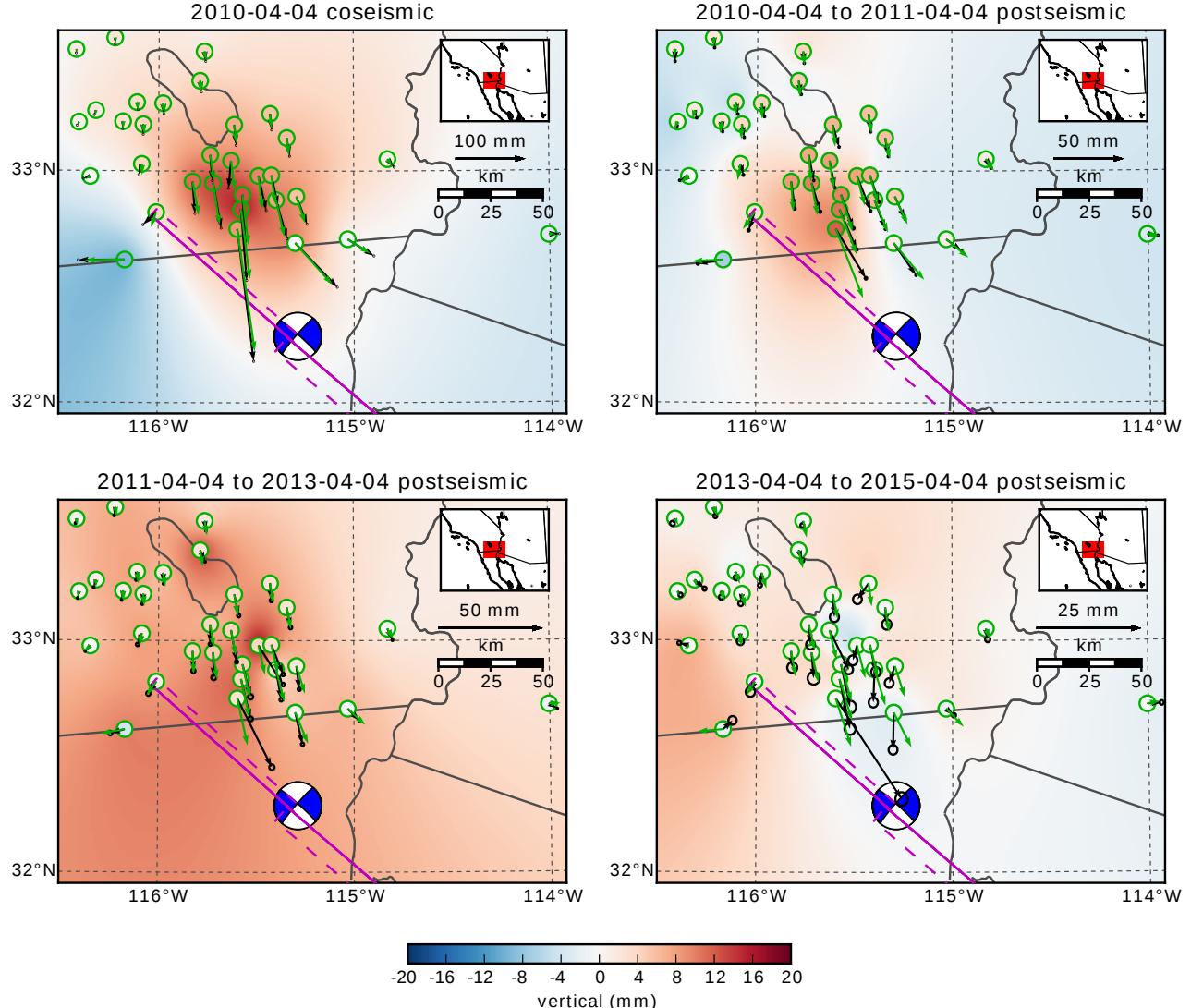
ing southward at a rate of a few millimeters per year along the direction of the El Mayor-Cucapah P-axis. A similar eastward trend can be seen in the few far-field stations in Arizona, located along the T-axis. After three years, the trend in far-field postseismic deformation is barely perceptible. Most far-field stations display an initial subsidence for the first year after the El Mayor-Cucapah earthquake followed by continued uplift. This trend in vertical deformation can be observed in all three of the quadrants where postseismic data is available, which means that the vertical deformation does not exhibit an anti-symmetric quadrant pattern, as would be expected for postseismic processes. Although we use vertical deformation in our analysis in Section 3, we do not put an emphasis on trying to describe the vertical deformation because it likely does not have postseismic origins.

The near-field postseismic deformation is notably sustained when compared to the far-field deformation. Namely, the station in this study which is closest to the El Mayor-Cucapah epicenter, P496, has a steady postseismic trend of  $\sim 1.5$  cm/yr to the south after about one year. Vertical postseismic deformation in the near-field does display a quadrant pattern which is consistent with the coseismic vertical deformation, suggesting that it is resulting from postseismic processes. However, the vertical postseismic signal is only apparent for the first year after the earthquake (Figure 4). As with the far-field deformation, there is a general trend of uplift in the near-field after about one year.

### 3 Postseismic Modeling

We seek to find the mechanisms driving five years of postseismic deformation following the El Mayor-Cucapah earthquake and we consider afterslip and viscoelastic relaxation as candidate mechanisms. Poroelastic rebound has also been used to model postseismic deformation [e.g. Jónsson *et al.*, 2003]; however, Gonzalez-Ortega *et al.* [2014] found that poroelastic rebound is unlikely to be a significant contributor to postseismic deformation following the El Mayor-Cucapah earthquake. Furthermore, we consider stations which are sufficiently far away from the rupture that poroelastic rebound should be insignificant.

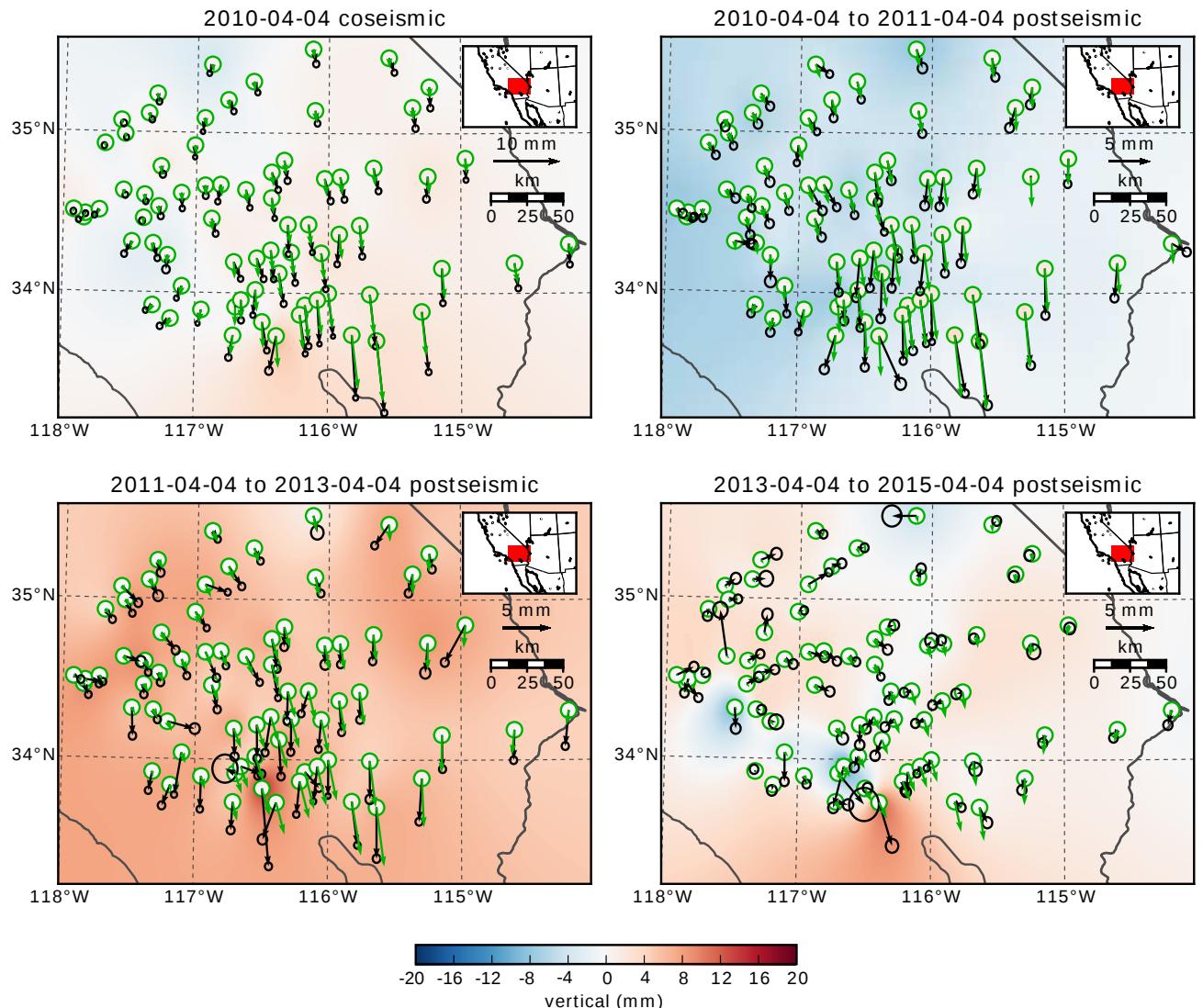
We estimate coseismic and time-dependent postseismic fault slip, both of which are assumed to occur on a fault geometry modified from Wei *et al.* [2011b]. Field studies [Fletcher *et al.*, 2014] and LIDAR observations [Oskin *et al.*, 2012] have revealed a significantly more complicated fault geometry than what was inferred by Wei *et al.* [2011b], especially within the Sierra Cucapah. However, we find that a relatively simple coseismic fault geometry based on



215 **Figure 4.** Near-field coseismic and cumulative postseismic displacements over the indicated time periods  
 216 (black) and predicted displacements for our preferred model from Section 3.3 (green). Observed vertical  
 217 displacements are shown as an interpolated field and predicted vertical displacements are shown within the  
 218 green circles. Note that the interpolant is not well constrained in Mexico where there is no data available.

depth (km)	$\lambda$ (GPa)	$\mu$ (GPa)	$\eta_{\text{eff}}$ ( $10^{18}$ Pa s)	$\mu_k/\mu$
0-5	24.0	24.0	-	-
5-15	35.0	35.0	-	-
15-30	42.0	42.0	44.3	0.0
30-60	61.0	61.0	5.91	0.375
60-90	61.0	61.0	1.99	0.375
90-120	61.0	61.0	1.31	0.375
120-150	61.0	61.0	1.10	0.375
150- $\infty$	61.0	61.0	1.07	0.375

229 **Table 1.** Assumed and estimated material properties.  $\lambda$  and  $\mu$  are assumed known *a priori* and are based  
 230 on the values used for the coseismic model by Wei *et al.* [2011b]. The values for  $\eta_{\text{eff}}$  are estimated in Section  
 231 3.2, and  $\mu_k/\mu$  are the optimal shear moduli ratios found in Section 3.3 for a Zener rheology upper mantle.

**Figure 5.** Same as Figure 4 but for far-field stations.

219

[Wei *et al.*, 2011b] is adequate because most of the stations used in this study are sufficiently far from the El Mayor-Cucapah rupture that they are insensitive to the details in the fault geometry found by Fletcher *et al.* [2014] and Osokin *et al.* [2012]. The fault geometry used in this study (Figure 1) consists of the two main fault segments inferred by Wei *et al.* [2011b], where the northern segment runs through the Sierra Cucapah up to the US-Mexico border and the southern segment is the Indiviso fault which extends down to the Gulf of California. Both segments extend from the surface to 15 km depth. We extend the northern segment by 40 km to the northwest, which is motivated by the clustering of aftershocks on the northern tip of the coseismic rupture zone [Hauksson *et al.*, 2011; Kroll *et al.*, 2013]. This extended fault segment was also found to be necessary by Rollins *et al.* [2015] and Pollitz *et al.* [2012] in order to describe the postseismic deformation.

### 3.1 Elastic Postseismic Inversion

We consider a variety of rheologic models for the lower crust and upper mantle. The simplest rheologic model is to consider them to be effectively elastic and isotropic. In such case, the rheologic parameters consist of the Lamé parameters,  $\lambda$  and  $\mu$ , which are reasonably well known and we use the same values used by Wei *et al.* [2011b] throughout this paper (Table 1). The only unknown is the distribution of fault slip, which can be easily estimated from postseismic deformation through linear least squares. Rollins *et al.* [2015] used a subset of the GPS stations considered in this study and found that three years of postseismic deformation following the El Mayor-Cucapah earthquake can be explained with afterslip on the coseismic fault plane without requiring any viscoelastic relaxation. We also perform an elastic slip inversion, but we use GPS stations within a larger radius about the El Mayor-Cucapah epicenter (400 km instead of  $\sim$ 200 km). Our forward problem describing predicted postseismic deformation,  $u_{\text{pred}}$ , in terms of time dependent fault slip,  $s$ , is

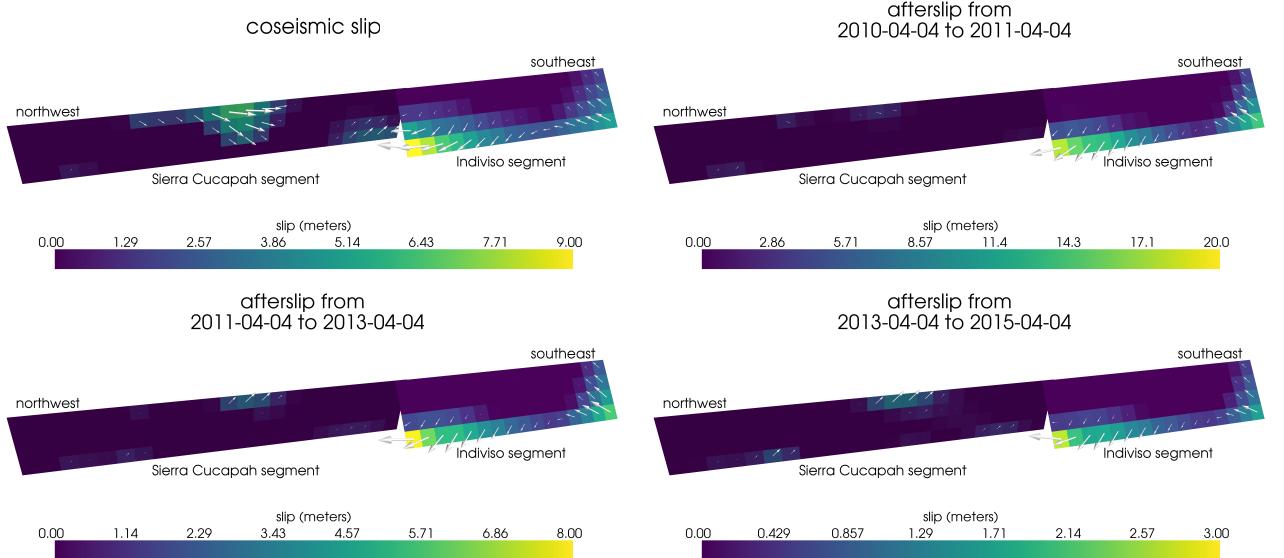
$$u_{\text{pred}}(x, t) = \int_F s(\xi, t) g(x, \xi) d\xi, \quad (9)$$

where  $F$  denotes the fault and  $g(x, \xi)$  is the elastic Green's function describing displacement at surface position  $x$  resulting from slip at  $\xi$  on the fault. We estimate coseismic slip and the rate of afterslip over the postseismic time intervals 0.0-0.125, 0.125-0.25, 0.25-0.5, 0.5-1.0, 1.0-2.0, 2.0-3.0, 3.0-4.0, and 4.0-5.0 years. Each fault segment is discretized into roughly 4 km by 4 km patches and we impose that the direction of slip and slip rate are within  $45^\circ$  of right-lateral. We also add zeroth-order Tikhonov regularization so that our solution for  $s$  satisfies

$$\min_s \left( \left\| \frac{u_{\text{pred}}(s) - u_{\text{post}}}{\sigma_{\text{post}}} \right\|_2^2 + \lambda_s \|s\|_2^2 \right), \quad (10)$$

where  $\sigma_{\text{post}}$  is the uncertainty on postseismic displacements and  $\lambda_s$  is a penalty parameter which is chosen with a trade-off curve. We use Pylith [Aagaard *et al.*, 2013] to compute the Green's functions for this inversion as well as for the remaining inversions in this paper.

Our coseismic slip and afterslip solutions are shown in Figure 6. Similar to Rollins *et al.* [2015], we find that a large amount of afterslip on the Indiviso fault segment is required to explain the observations. The potency of our inferred coseismic slip is  $3.2 \times 10^9 \text{ m}^3$ , equivalent to a Mw=7.28 earthquake when assuming a shear modulus of 32 GPa. The potency of our inferred cumulative five years of afterslip is  $6.1 \times 10^9 \text{ m}^3$ , equivalent to a Mw=7.46 earthquake, which is unrealistically large if we consider afterslip to be driven by coseismically induced stresses. Figure 7 shows the time series for the observed and predicted postseismic displacements at stations along the El Mayor-Cucapah P-axis. We show the radial component of displacements with respect to the El Mayor-Cucapah epicenter and we also rescale the displacements so that the difference between the minimum and maximum observed displacements are the same for each station. Our elastic slip model accurately describes near-field postseis-



294 **Figure 6.** Coseismic slip and cumulative afterslip over the indicated time intervals when assuming the crust  
295 and mantle are elastic. Color indicates the magnitude of slip while arrows indicate the motion of the hanging  
296 wall.

297 mic deformation and systematically underestimates postseismic deformation at epicentral dis-  
298 tances  $\gtrsim 150$  km. When the fault segments used in the inversion are extended down to 30 km  
299 depth, rather than 15 km, the systematic far-field residuals are smaller but remain apparent.  
300 Because an elastic model requires an unrealistic amount of afterslip and is unable to predict  
301 far-field deformation, we move on to consider viscoelastic models in the next section.

### 306 3.2 Early Postseismic Inversion

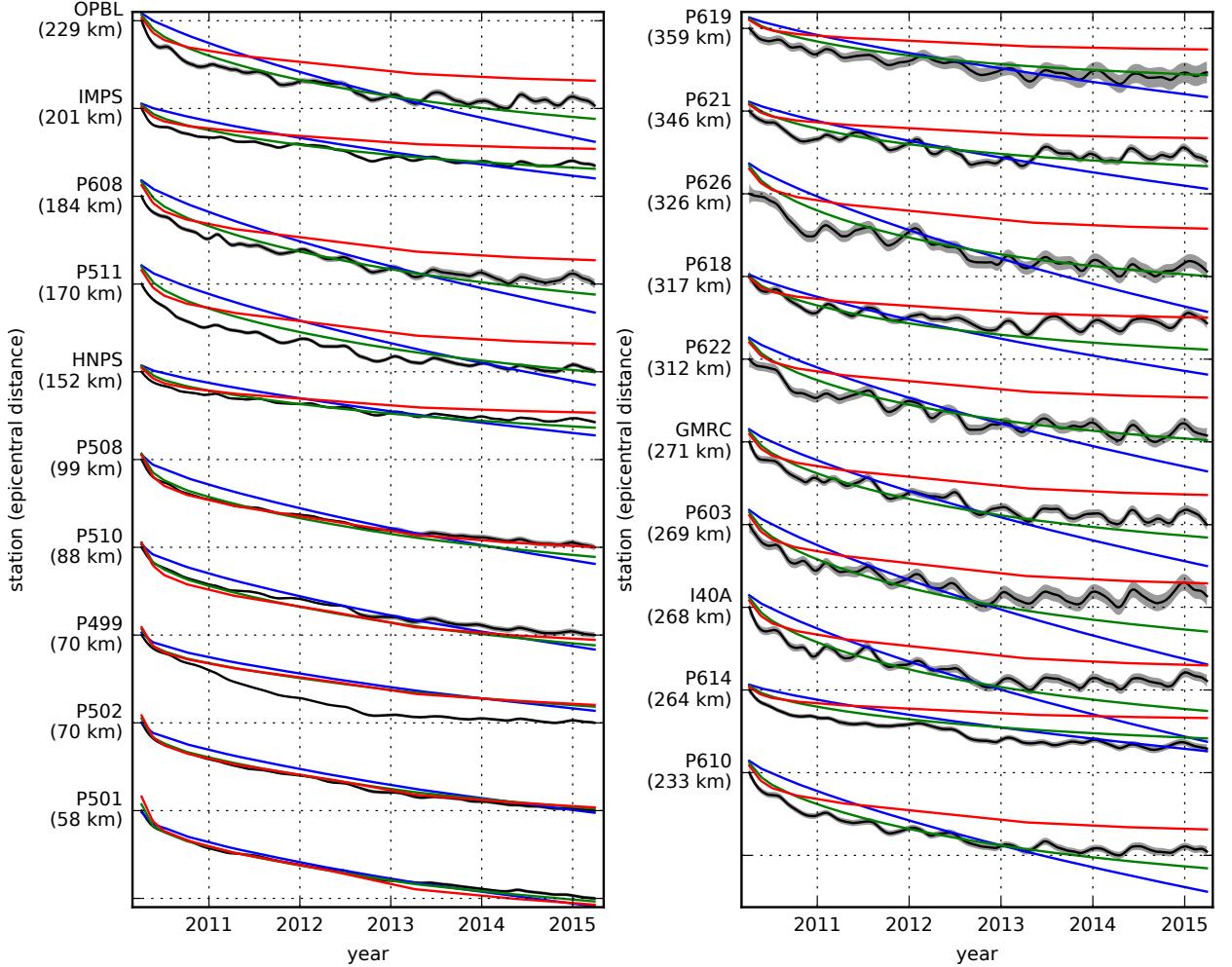
307 For any linear viscoelastic rheology of the crust and mantle, postseismic displacements  
308 resulting from time dependent fault slip can be described as

$$u_{\text{pred}}(x, t) = \int_F s(\xi, t) g(x, \xi) d\xi + \int_0^t \int_F s(\xi, \tau) f(t - \tau, x, \xi) d\xi d\tau, \quad (11)$$

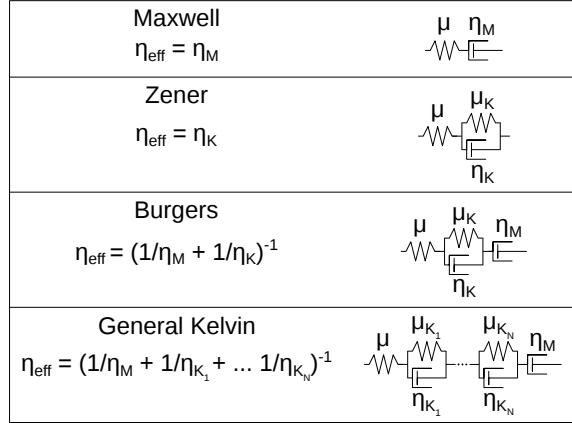
309 where  $f(t, x, \xi)$  describes the time-dependent velocity at  $x$  resulting from viscoelastic relax-  
310 ation of stresses induced by slip at  $\xi$ .  $f$  is a function of  $\lambda$ ,  $\mu$ , and any additional rheologic pa-  
311 rameters controlling the viscoelastic response, which are generally not well known. Schematic  
312 representations of the viscoelastic rheologic models considered in this study are shown in Fig-  
313 ure 8. We discuss these rheologic models and their use in geophysical studies in Section 4.

314 In order to greatly simplify the inverse problem, we use the method described in *Hines*  
315 and *Hetland* [2016] to constrain an initial effective viscosity structure from the early postseis-  
316 mic deformation. Our method uses the fact that coseismic stresses throughout the crust and  
317 upper mantle depend on the instantaneous elastic parameters and are independent of the vis-  
318 coelastic parameters which we wish to estimate. Immediately following an earthquake, each  
319 parcel will have a strain rate that is proportional to the coseismic stress and inversely propor-  
320 tional to the parcel's effective viscosity,  $\eta_{\text{eff}}$ . Using one-dimensional rheologic models, we de-  
321 fine the effective viscosity as

$$\eta_{\text{eff}} = \left. \frac{\sigma}{\dot{\varepsilon}} \right|_{t=0}, \quad (12)$$



297 **Figure 7.** Scaled radial component of postseismic displacements. Downward motion indicates that the  
 298 station is moving toward the El Mayor-Cucapah epicenter. Displacement time series are scaled so that the  
 299 minimum and maximum observed values lie on the grid lines. The observed postseismic displacements,  $u_{\text{post}}$   
 300 are shown in black with gray indicating the 68% confidence interval. The displacements predicted by the best  
 301 fitting elastic model are shown in red. The blue and green lines are the predicted postseismic displacements  
 302 for the models discussed in Section 3.3. The blue lines show the predicted displacements for the model with  
 303 a Maxwell viscoelastic lower crust and upper mantle. The green line shows the predicted displacements  
 304 for the model with a Maxwell viscoelastic lower crust and a Zener viscoelastic upper mantle. The effective  
 305 viscosities are the same for both models and are shown in Figure 12.



314 **Figure 8.** Schematic illustration of the rheologic models considered in this paper as well as their effective  
 315 viscosities.

324 where  $\sigma$  is an applied stress at  $t = 0$  and  $\dot{\varepsilon}$  is the resulting strain rate. Figure 8 shows how  
 325  $\eta_{\text{eff}}$  relates to the parameters for various linear viscoelastic rheologies. We can deduce that the  
 326 initial rate of surface deformation resulting from viscoelastic relaxation is a summation of the  
 327 surface deformation resulting from relaxation in each parcel, scaled by the reciprocal of the  
 328 parcel's effective viscosity. That is to say

$$f(0, x, \xi) = \int_L \frac{h(x, \xi, \zeta)}{\eta_{\text{eff}}(\zeta)} d\zeta, \quad (13)$$

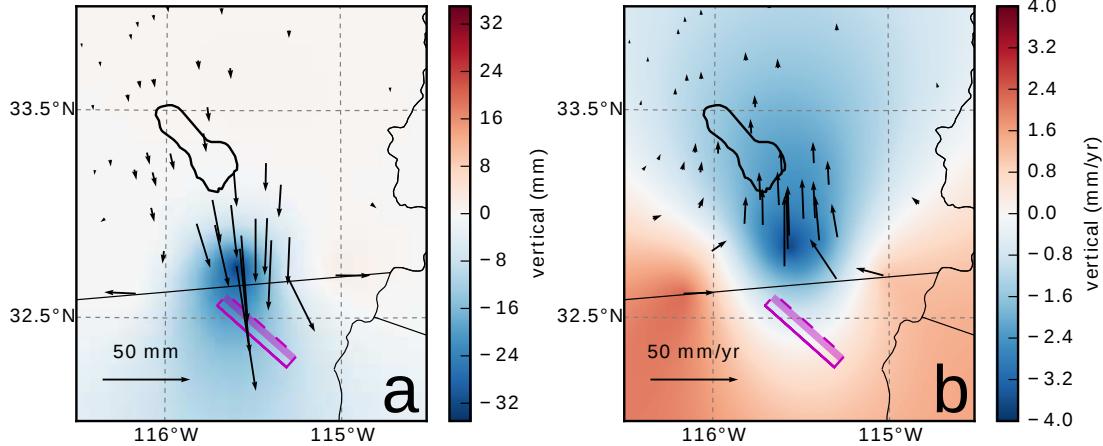
329 where  $L$  denotes the crust and mantle and  $h(x, \xi, \zeta)$  describes the initial rate of deformation  
 330 resulting from viscoelastic relaxation at  $\zeta$  induced by slip at  $\xi$ . We can combine eq. (13) with  
 331 eq. (11) to get a first-order approximation for early postseismic deformation,

$$u_{\text{pred}}(x, t) \approx \int_F s(\xi, t) g(x, \xi) d\xi + \int_0^t \int_F \int_L \frac{s(\tau, \xi)}{\eta_{\text{eff}}(\zeta)} h(x, \xi, \zeta) d\zeta d\xi d\tau, \quad (14)$$

332 which is valid for as long as the rate of deformation resulting from viscoelastic relaxation is  
 333 approximately constant. Although eq. (14) may only be valid for a short portion of the post-  
 334 seismic period, its utility becomes apparent when noting that  $g$  and  $h$  are only functions of  
 335 the fault geometry and instantaneous elastic properties,  $\lambda$  and  $\mu$ , and thus  $g$  and  $h$  can be com-  
 336 puted numerically as a preprocessing step. The forward problem in eq. (14) can then be rapidly  
 337 evaluated for any realization of  $s$  and  $\eta_{\text{eff}}$ . This is in contrast to evaluating the full forward  
 338 problem, eq. (11), numerically for each realization of  $s$  and the unknown rheologic proper-  
 339 ties.

340 Details on how eq. (14) is used to estimate  $s$  and  $\eta_{\text{eff}}$  from postseismic deformation can  
 341 be found in *Hines and Hetland* [2016]. A non-linear Kalman filter based inverse method can  
 342 also be used to estimate  $s$  and  $\eta_{\text{eff}}$  in a manner similar to *Segall and Mathews* [1997] or *McGuire*  
 343 and *Segall* [2003], in which we would not have to explicitly impose a time dependent parametriza-  
 344 tion of  $s$ . We have thoroughly explored Kalman filter based approaches, but we ultimately pre-  
 345 fer the method described in *Hines and Hetland* [2016] because of its relative simplicity. More-  
 346 over, we believe the piecewise continuous representation of slip with respect to time is suf-  
 347 ficiently general for the resolving power of these GPS data.

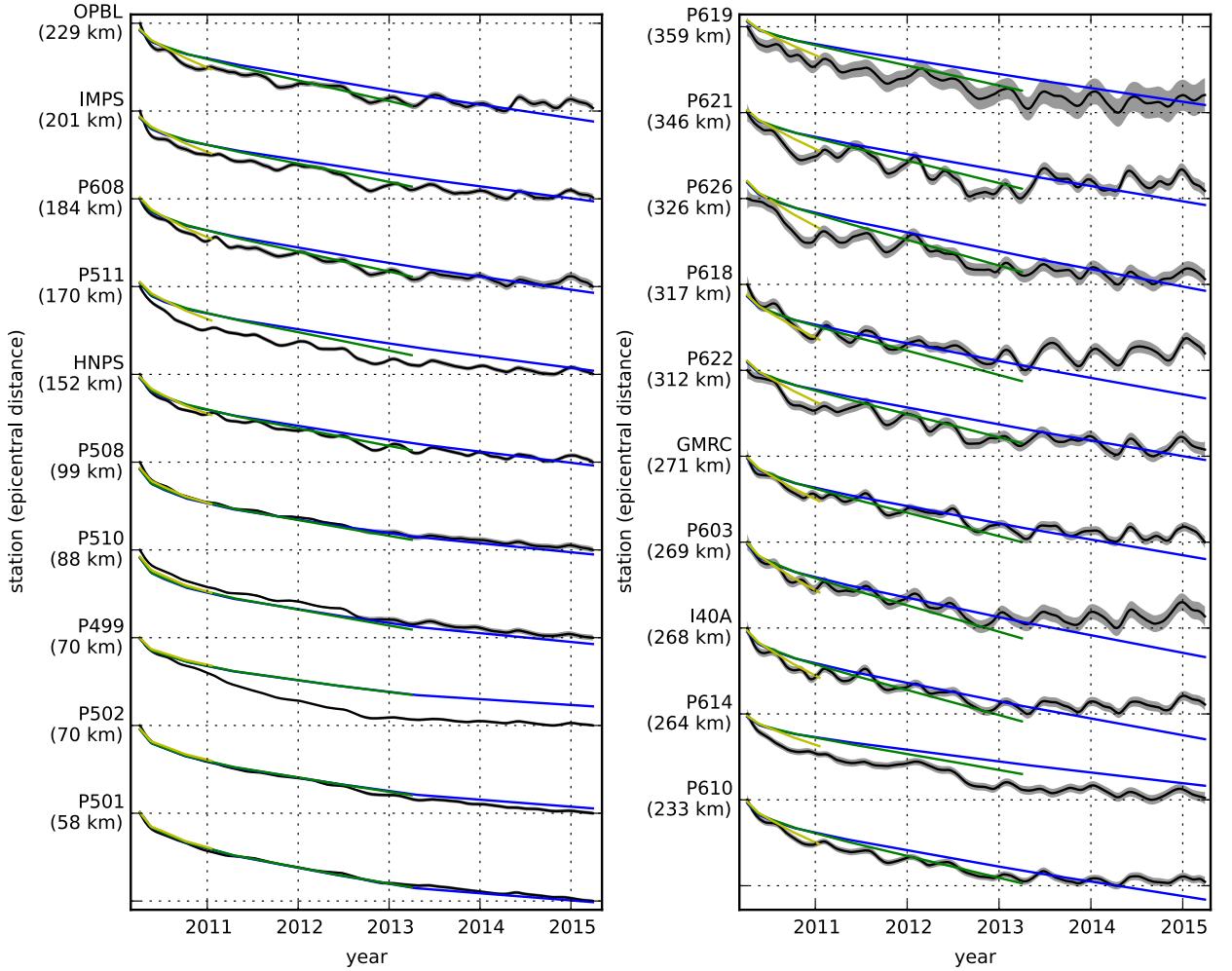
348 We estimate coseismic slip and afterslip with the same spatial and temporal discretiza-  
 349 tion as in Section 3.1. Simultaneously, we estimate  $\eta_{\text{eff}}$  within six vertically stratified layers  
 350 which have depths ranging from 15–30 km, 30–60 km, 60–90 km, 90–120 km, 120–150 km,



370 **Figure 9.** Displacements resulting from fault slip at lower crustal depths (a), and initial velocities resulting  
 371 from subsequent relaxation of a viscoelastic lower crust (b). The fault segment dips  $75^\circ$  to the north-east and  
 372 its surface projection is outlined in magenta. The highlighted area on the fault extends from 15 to 30 km depth  
 373 and indicates where 1 meter of right-lateral slip was imposed. The elastic properties of the crust and mantle  
 374 are the same as in Table 1, and  $\eta_{\text{eff}}$  is  $10^{18}$  Pa s in the lower crust. Vertical displacements are interpolated  
 375 between station locations.

351 and from 150 km to the bottom of our numerical model domain at 800 km. We again restrict  
 352 fault slip to occur between 0 and 15 km depth, which is done in order to help eliminate in-  
 353 evitable non-uniqueness in the inversion. It is well understood that fault slip at sufficiently great  
 354 depths can produce surface deformation that is indistinguishable from viscoelastic relaxation,  
 355 at least in two-dimensional earthquake models [Savage, 1990]. Additionally, we note that when  
 356 simultaneously estimating both afterslip and viscosity in the lower crust, the inverse problem  
 357 becomes particularly ill-posed. This ill-posedness is illustrated in Figure 9, which shows the  
 358 displacements resulting from a meter of slip on a fault extending from 15 to 30 km depth and  
 359 the initial velocity resulting from subsequent viscoelastic relaxation in the lower crust, which  
 360 is given a viscosity of  $10^{18}$  Pa s. In this demonstration, the viscoelastic relaxation is entirely  
 361 driven by the fault slip in the lower crust. The horizontal displacements from fault slip are in  
 362 the opposite direction as the displacements resulting from viscoelastic relaxation. This means  
 363 that surface displacements resulting from afterslip at lower crustal depths can be cancelled out,  
 364 at least partially, by a low viscosity lower crust. We eliminate this null space by allowing only  
 365 one mechanism in the lower crust, which we choose to be viscoelastic relaxation. This is not  
 366 to say that we do not believe deep afterslip to be a possibility; rather, we restrict slip to seis-  
 367 mogenic depths as a modeling necessity. Although, it has been noted that the pattern of ver-  
 368 tical postseismic deformation following the El Mayor-Cucapah earthquake indicates that a sig-  
 369 nificant amount of afterslip must be shallow [Rollins *et al.*, 2015].

376 We must determine at which point the early postseismic approximation breaks down, which  
 377 we will denote as  $t_{\text{bd}}$ . As noted, eq. (14) is valid for as long as the rate of deformation re-  
 378 sulting from viscoelastic relaxation is approximately constant. We can almost certainly assume  
 379 that deformation at the most far-field stations, which are  $\sim 400$  km away from the El Mayor-  
 380 Cucapah epicenter, is the result of viscoelastic relaxation. The approximation should then be  
 381 valid for as long as a linear trend adequately approximates the far-field deformation. Using  
 382 this logic, it would appear that  $t_{\text{bd}}$  is about one year after the El Mayor-Cucapah earthquake.  
 383 Another way to determine  $t_{\text{bd}}$  is to find the best fitting prediction of eq. (14) to observed de-  
 384 formation using increasing durations of the postseismic time series.  $t_{\text{bd}}$  should be the point  
 385 when eq. (14) is no longer capable of describing the observed deformation without incurring  
 386 systematic misfits. When using eq. (14) to fit the entire five years of postseismic displacements,



398 **Figure 10.** Observed postseismic displacements (black) and best fitting predictions of eq. (14) to 5.0 (blue),  
 399 3.0 (green), and 0.8 (yellow) years of the postseismic data.

387 we see that the near-field displacements (e.g., station P501) are accurately predicted. When  
 388 looking at displacements in the far-field (e.g., station P621), we see that eq. (14) overestimates  
 389 the rate of deformation in the later postseismic period and underestimates the rate of defor-  
 390 mation in the early period (Figure 10). Due to the low signal-to-noise ratios for far-field sta-  
 391 tions, it is difficult to determine at what point eq. (14) is no longer able to predict the observed  
 392 displacements; however, we settle on  $t_{bd} = 0.8$  years after the earthquake, while acknowl-  
 393 edging that the choice is subjective. As noted in *Hines and Hetland* [2016], overestimating  $t_{bd}$   
 394 will result in a bias towards overestimating  $\eta_{eff}$ , while picking a  $t_{bd}$  which is too low will not  
 395 necessarily result in a biased estimate of  $\eta_{eff}$ , although the uncertainties would be larger. We  
 396 can then consider inferences of  $\eta_{eff}$  to be an upper bound on the viscosity needed to describe  
 397 the far-field rate of deformation during the first 0.8 years of postseismic deformation.

400 We estimate coseismic slip, afterslip, and effective viscosities by solving

$$\min_{s, \eta_{eff}} \left( \left\| \frac{u_{pred}(s, \eta_{eff}) - u_{post}}{\sigma_{post}} \right\|_2^2 + \lambda_s \|s\|_2^2 + \lambda_\eta \|\nabla \eta_{eff}^{-1}\|_2^2 \right), \quad (15)$$

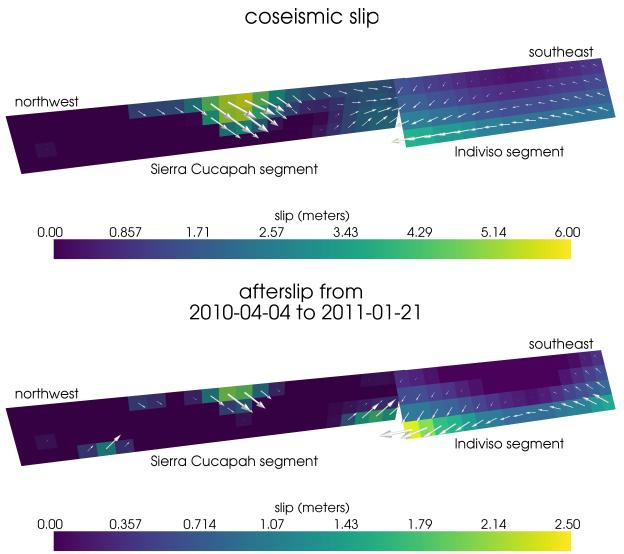
401 where  $u_{\text{post}}$  consists of the first 0.8 years of postseismic deformation and  $u_{\text{pred}}$  are the pre-  
 402 dicted displacements from eq. (14). Due to inherent non-uniqueness, we have added zeroth-  
 403 order Tikhonov regularization to estimates of  $s$  and second-order Tikhonov regularization to  
 404 estimates of effective fluidity  $\eta_{\text{eff}}^{-1}$ . The degree to which we impose the regularization on slip  
 405 and fluidity is controlled by the penalty parameters  $\lambda_s$  and  $\lambda_\eta$ , which are chosen with trade-  
 406 off curves (Figure S1). Our goal is to get a prior constraint on  $\eta_{\text{eff}}$  to minimize the amount  
 407 of searching we have to do when describing the postseismic deformation over the full five years,  
 408 which we do in Section 3.3. Estimates of  $s$  made here will not be used in Section 3.3, and  
 409 so the motivation behind adding regularization to  $s$  is to ensure that the slip driving viscoelas-  
 410 tic relaxation in eq. (14) is sensible.

411 Our initial estimate for coseismic slip and cumulative afterslip over the first 0.8 years  
 412 after the El Mayor-Cucapah earthquake are shown in Figure 11. Similar to our elastic slip model  
 413 from Section 3.1, a significant amount of right-lateral and normal coseismic slip is inferred  
 414 to be on the Sierra Cucapah segment. Our coseismic slip solution on the Sierra Cucapah seg-  
 415 ment is consistent with field studies [Fletcher *et al.*, 2014] and the model from Wei *et al.* [2011b].  
 416 Our inferred slip on the Indiviso fault segment differs from Wei *et al.* [2011b] because the GPS  
 417 data used in this study is not capable of resolving the spatial distribution of fault slip on that  
 418 segment (Figure S2). The potency of inferred coseismic slip is  $3.3 \times 10^9 \text{ m}^3$ , which is also  
 419 about the same as that inferred from Section 3.1. The present inference of afterslip on the In-  
 420 diviso fault is significantly less than what was found in the Section 3.1 where we did not ac-  
 421 count for viscoelasticity. When fault slip is simultaneously estimated with viscosity, the po-  
 422 tency of inferred afterslip over the first 0.8 years after the earthquake is  $0.85 \times 10^9 \text{ m}^3$ , com-  
 423 pared to  $3.5 \times 10^9 \text{ m}^3$  when we assume the crust and upper mantle are elastic. The signifi-  
 424 cant amount of afterslip inferred on the Indiviso fault in Section 3.1 seems to be compensat-  
 425 ing for unmodeled viscoelastic relaxation. The fact that there is still an appreciable amount  
 426 of afterslip inferred on the Indiviso fault raises the question of whether it is compensating for  
 427 viscoelastic relaxation that is more localized than what we allow for since we only estimate  
 428 depth dependent variations in viscosity.

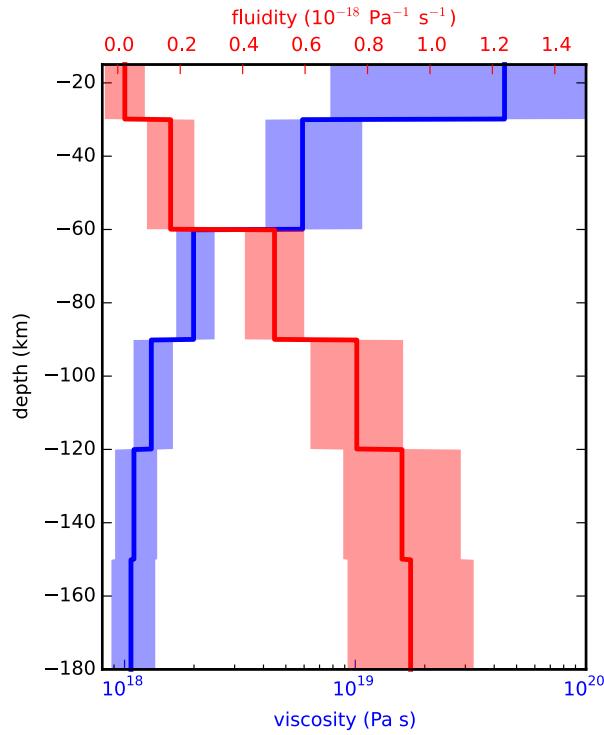
429 Our estimated effective viscosities, and corresponding fluidities, are shown in Figure 12.  
 430 Although fluidity is rarely used in geophysical literature, eq. (13) is linear with respect to flu-  
 431 idity and so the fluidity indicates the amplitude of the viscoelastic signal coming from each  
 432 layer. We use bootstrapping to find the 95% confidence intervals for our estimated effective  
 433 viscosities which are shown as shaded regions in Figure 12. It is important to remember that  
 434 the presented effective viscosities were estimated with a smoothing regularization constraint  
 435 and so the uncertainties are almost certainly underestimated [Aster *et al.*, 2011]. Indeed, many  
 436 viscosity profiles which are outside of the shown confidence intervals can just as adequately  
 437 described the first 0.8 years of postseismic deformation. Our solution in Figure 12 should be  
 438 interpreted as the smoothest effective viscosity profile which is capable of describing the data.  
 439 This means that any sharp viscosity transitions will be tapered out in the inversion, which we  
 440 demonstrate with a synthetic test in Figure S2. Nonetheless, a robust feature that we see is that  
 441 the largest jump in fluidity is at 60 km depth, which is consistent with the range of lithosphere-  
 442 asthenosphere boundary depths inferred by Lekic *et al.* [2011]. This transitional depth is also  
 443 consistent with the the viscosity structure required to explain far-field postseismic deforma-  
 444 tion following the Hector Mine earthquake [Freed *et al.*, 2007]. We find that the viscosity be-  
 445 low 60 km depth needs to be  $\sim 1 \times 10^{18} \text{ Pa s}$  to describe the early rate of postseismic defor-  
 446 mation at far-field stations while the lower crust and uppermost mantle need to be relatively  
 447 stronger. The viscosity of the lower crust has the largest uncertainties because there is no ev-  
 448 idence of relaxation in that layer, meaning that it is effectively elastic over the first 0.8 years  
 449 after the earthquake.

### 455 3.3 Full Postseismic Inversion

456 In the previous section, we used the inverse method from Hines and Hetland [2016] to  
 457 constrain the effective viscosity structure required to explain the first 0.8 years of postseismic



450 **Figure 11.** Coseismic slip and afterslip inferred by fitting eq. (14) to the first 0.8 years of postseismic  
451 displacements.



452 **Figure 12.** Effective viscosities and associated fluidities inferred by fitting eq. (14) to the first 0.8 years of  
453 postseismic displacements. 95% confidence intervals, estimated from bootstrapping, are indicated by shaded  
454 regions.

458 deformation. In this section, we use these effective viscosities as a prior constraint when searching  
 459 for models which are capable of describing the available five years of postseismic data,  
 460 where our forward problem is now eq. (11) rather than the approximation given by eq. (14).  
 461 We perform a series of fault slip inversions assuming a variety of rheologies for the lower crust  
 462 and upper mantle which are consistent with our findings from Section 3.2. We appraise each  
 463 model using the mean chi-squared value,

$$\bar{\chi}^2 = \frac{1}{N} \left\| \frac{u_{\text{pred}} - u_{\text{post}}}{\sigma_{\text{post}}} \right\|_2^2, \quad (16)$$

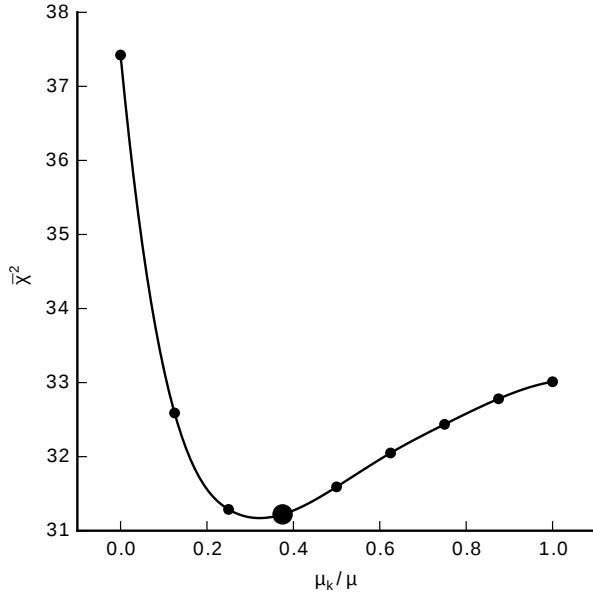
464 where  $N$  is the number of observations.

465 We first assume that the crust and mantle can be described with a Maxwell rheology,  
 466 and we set the steady-state viscosity,  $\eta_M$ , equal to our inference of  $\eta_{\text{eff}}$ . We compute  $f$  and  
 467  $g$  from eq. (11) using Pylith, and we use the same spatial and temporal discretization of  $s$  as  
 468 in Sections 3.1 and 3.2. We estimate  $s$  using linear least squares and find a misfit of  $\bar{\chi}^2 =$   
 469 37.4. For comparison,  $\bar{\chi}^2 = 35.3$  for the elastic model from Section 3.1. The Maxwell vis-  
 470 coelastic model has a larger misfit because it tends to overestimate the rate of deformation af-  
 471 ter about three years (Figure 7). Since our initial estimates of  $\eta_{\text{eff}}$  may be biased towards over-  
 472 estimating viscosities, we have also performed the slip inversion where we use uniformly lower  
 473 viscosities in the crust and mantle; however, decreasing the viscosity only increases the mis-  
 474 fit. Although, the viscosities used here are consistent with the successful Maxwell viscoelas-  
 475 tic models found by *Rollins et al.* [2015] and *Spinler et al.* [2015], which had mantle viscosi-  
 476 ties on the order of  $10^{18}$  Pa s and relatively higher lower crustal viscosities, we find that such  
 477 a model is incapable of describing the entire postseismic time series. *Pollitz et al.* [2001] sim-  
 478 ilarly recognized this deficiency in a Maxwell rheology, which then motivated their exploration  
 479 of a Burgers rheology upper mantle [*Pollitz*, 2003].

480 Instead of exploring a Burgers rheology mantle, which introduces two new parameters  
 481 that need to be estimated, the transient viscosity,  $\eta_K$ , and transient shear modulus,  $\mu_K$ , we first  
 482 consider a Zener rheology for the mantle, which only introduces one additional unknown pa-  
 483 rameter,  $\mu_K$ . We assume that the lower crust still has a Maxwell rheology. The steady-state  
 484 viscosity in the crust and the transient viscosity in the mantle are set equal to the inferred ef-  
 485 fective viscosities. We then estimate the ratio of shear moduli,  $\frac{\mu_K}{\mu}$ . We compute nine differ-  
 486 ent sets of Green's functions,  $f$  and  $g$ , where we assume values of  $\frac{\mu_K}{\mu}$  ranging from 0 to 1.  
 487 The former being a degenerate case where the Zener model reduces to the above Maxwell model.  
 488 We estimate coseismic slip and afterslip for each realization of  $\frac{\mu_K}{\mu}$ . We find that a shear mod-  
 489 uli ratio of 0.375 yields the best prediction to the observed postseismic displacements with a  
 490 misfit of  $\bar{\chi}^2 = 31.2$  (Figure 13). The improvement in the Zener model over the Maxwell model  
 491 can be seen in the fit to the far-field data (Figure 7) where the Zener model does a significantly  
 492 better job at explaining the transient rate of deformation throughout the five years considered  
 493 in this study. The rheologic parameters for our preferred Zener model are summarized in Ta-  
 494 ble 1.

497 Because we are able to adequately describe the available five years of postseismic de-  
 498 formation with a Zener model, we do not find it necessary to explore the parameter space for  
 499 a more complicated Burgers rheology. However, since the Zener model is a Burgers model with  
 500 an infinite steady-state viscosity, we can conclude that any Burgers rheology that has a tran-  
 501 sient viscosity consistent with that found in Section 3.2 and a steady-state viscosity  $\gtrsim 10^{20}$   
 502 Pa s, which is effectively infinite on the time scale of five years, would also be able to sat-  
 503 isfactorily describe the observable postseismic deformation.

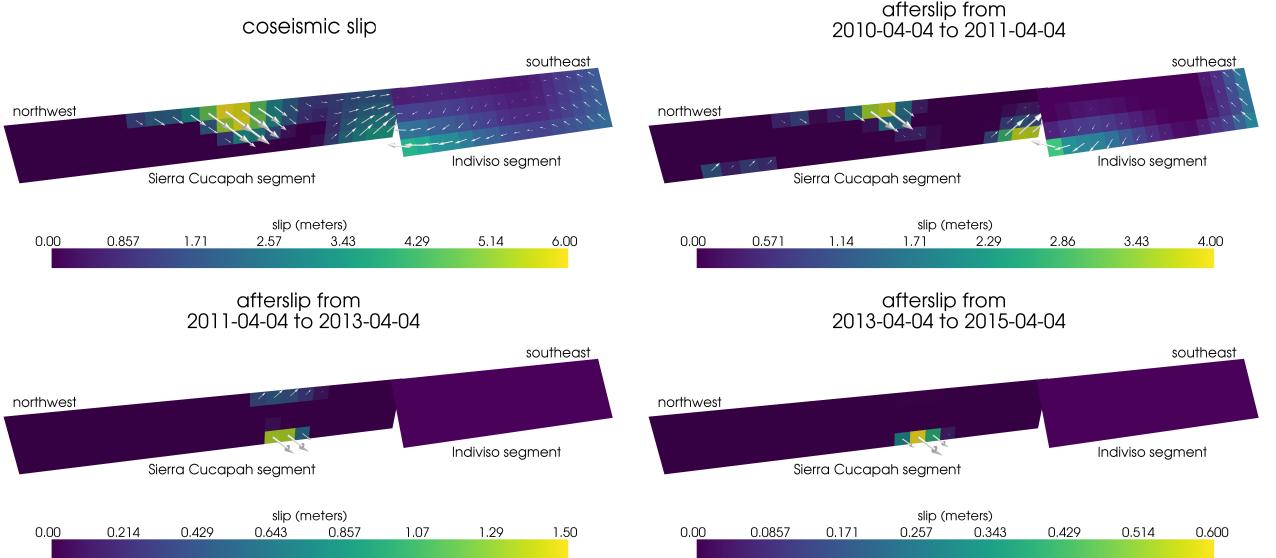
504 The regularized inference of coseismic slip and afterslip for our preferred Zener model  
 505 is shown in Figure 14. The inferred coseismic potency is  $3.0 \times 10^9$  m<sup>3</sup>, equivalent to a Mw=7.26  
 506 earthquake, where most of the slip is shallow and on the Sierra Cucapah fault segment. The  
 507 potency of five years of afterslip is  $1.1 \times 10^9$  m<sup>3</sup>. Most of the afterslip in our preferred model  
 508 occurs within the first year after the earthquake and coincides with the location of our inferred



495 **Figure 13.** Mean chi-squared value for the best fitting slip model as a function of the transient shear modu-  
 496 lus relative to the elastic shear modulus in a Zener upper mantle. Large dot indicates our preferred ratio.

509 coseismic slip. Inferred afterslip within the first year is accounting for the most rapid near-  
 510 field transient deformation. After one year, afterslip is inferred to be deeper down on the Sierra  
 511 Cucapah segment, which is describing much of the sustained near-field postseismic deforma-  
 512 tion. We emphasize, that the GPS station closest to where we infer afterslip, P496, is still about  
 513 30 km away, which is too far for us to conclusively argue for sustained localized deformation  
 514 rather than shallow distributed deformation. The deep afterslip inferred after one year could  
 515 potentially be describing deformation resulting from lower crustal flow. To test this, we have  
 516 modified our preferred model by decreasing the lower crustal viscosity from  $5.91 \times 10^{19}$  Pa  
 517 s to  $1 \times 10^{19}$  Pa s, which is still consistent with our viscosity inference from Section 3.2, and  
 518 we inverted for fault slip. We find that a model with a weaker lower crust adequately describes  
 519 the postseismic displacements without any afterslip after one year, while still requiring about  
 520 the same amount of afterslip over the first year. We do believe that the early, shallow after-  
 521 slip on the Sierra Cucapah segment is a robust feature in our preferred model, while we are  
 522 not confident in our inference of later deep afterslip.

523 The postseismic displacements predicted by our preferred Zener model are shown in Fig-  
 524 ures 4, 5 and 7. The largest misfit occur within the Imperial Valley and there does not appear  
 525 to be any systematic trend in the residuals. This suggests that the large errors are due to lo-  
 526 calized processes such as fault slip in the Imperial Valley triggered by the El Mayor-Cucapah  
 527 earthquake [Wei *et al.*, 2011a, 2015]. We do not see any pattern in the residuals that would sug-  
 528 gest a laterally heterogeneous viscosity structure, which has been explored by Pollitz *et al.* [2012]  
 529 and Rollins *et al.* [2015]. We do notice regionally uniform seasonal oscillations in the lateral  
 530 and vertical components of the residuals with an amplitude of 1-2 millimeters. This is the re-  
 531 result of our method for data processing which is not able to completely remove the seasonal  
 532 signal in the GPS data, which was discussed in Section 2. Additionally, we see systematic mis-  
 533 fit in the later postseismic period west of the Landers and Hector Mine earthquakes, which may  
 534 be the result of unmodeled postseismic deformation following those earthquakes. Lastly, there  
 535 are clear discrepancies between the observed and predicted vertical displacements following  
 536 the first year after the El Mayor-Cucapah earthquake. We observe a broad uplift throughout  
 537 Southern California which is inconsistent with any postseismic model.



538 **Figure 14.** Inferred coseismic slip and afterslip when assuming a Maxwell rheology in the lower crust and  
 539 a Zener rheology in the upper mantle. The transient viscosity  $\eta_K$  in the mantle and steady-state viscosity  $\eta_M$   
 540 in the crust are set equal to the effective viscosities from Figure 12. We set  $\frac{\mu_K}{\mu} = 0.375$  in the upper mantle.

## 541 4 Discussion

542 It has long been recognized that deep afterslip and viscoelastic relaxation following an  
 543 upper crustal earthquake can result in similar horizontal ground deformation at the surface [e.g.  
 544 *Savage*, 1990; *Pollitz et al.*, 2001; *Hearn*, 2003; *Feigl and Thatcher*, 2006]. The similarity of  
 545 the horizontal postseismic deformation results in a non-uniqueness in inferences of afterslip  
 546 or viscoelastic relaxation. In contrast, the spatial pattern of vertical postseismic deformation  
 547 has been proposed to be a discriminant between deep afterslip and viscoelastic relaxation [e.g.  
 548 *Pollitz et al.*, 2001; *Hearn*, 2003]. It is, however, important to note that patterns of vertical de-  
 549 formation are very sensitive to the depth-dependence of viscosity below the upper crust [*Yang* and  
 550 *Toksöz*, 1981; *Hetland and Zhang*, 2014]. The similarity between deformation resulting  
 551 from deep afterslip and viscoelastic relaxation of coseismic stresses is different from the ill-  
 552 posedness described in Section 3.2. In our method, any inferred afterslip will also mechani-  
 553 cally drive additional viscoelastic relaxation. The horizontal deformation resulting from deep  
 554 afterslip will generally be in the opposite direction as horizontal deformation resulting from  
 555 viscoelastic relaxation of subsequent stresses in the lower crust (Figure 9). As a result, there  
 556 is a trade-off between inferences of deep afterslip and lower crustal viscosity. In our synthetic  
 557 tests in *Hines and Hetland* [2016], we have found that inverting surface deformation for af-  
 558 terslip and viscosity within the same depth interval tends to result in overestimated afterslip  
 559 and an underestimated viscosity.

560 Most postseismic studies assume Maxwell viscoelasticity in the lower crust and upper  
 561 mantle [e.g. *Nur and Mavko*, 1974; *Pollitz et al.*, 2000; *Hetland*, 2003; *Freed et al.*, 2006; *John-*  
 562 *son et al.*, 2009; *Hearn et al.*, 2009], which is the simplest viscoelastic rheologic model. In South-  
 563 ern California, postseismic studies following the Landers [*Pollitz et al.*, 2000], Hector Mine  
 564 [*Pollitz et al.*, 2001], and El Mayor-Cucapah earthquake [*Spinler et al.*, 2015; *Rollins et al.*, 2015],  
 565 have assumed Maxwell viscoelasticity in the lower crust and upper mantle and have inferred  
 566 upper mantle viscosities on the order of  $10^{17}$  to  $10^{18}$  Pa s and lower crust viscosities  $\gtrsim 10^{19}$   
 567 Pa s. These postseismic studies are consistent with *Kaufmann and Amelung* [2000] and *Cav-*  
 568 *alié et al.* [2007], who found that an upper mantle viscosity of  $10^{18}$  Pa s and a crustal viscos-

569 ity  $\gtrsim 10^{20}$  Pa s are necessary to describe subsidence resulting from changes in loading from  
 570 Lake Mead. This isostatic adjustment is a process with similar spatial and temporal scales as  
 571 postseismic deformation, and thus the inferred viscosities of these two types of studies would  
 572 likely agree. While these studies found viscosities that are consistent with our effective vis-  
 573 cosities from Section 3.2, they are inconsistent with viscosity estimates made from geophys-  
 574 ical processes that occur over longer time scales. For example, *Lundgren et al.* [2009] found  
 575 that lower crust and upper mantle viscosities on the order of  $10^{21}$  and  $10^{19}$  Pa s, respectively,  
 576 are needed to describe interseismic deformation along the Southern San Andreas fault zone  
 577 in the Salton Sea region. An even higher mantle viscosity, on the order of  $10^{20}$  Pa s, is re-  
 578 quired to describe isostatic adjustment resulting from the draining of Lake Bonneville, which  
 579 occurs on the time scales of  $10^4$  years [*Crittenden*, 1967; *Bills and May*, 1987].

580 An additional deficiency with the Maxwell rheology is that it predicts a steady decay  
 581 in the rate of postseismic deformation over time, which fails to describe the commonly ob-  
 582 served rapid, early transience followed by a relatively steady rate of postseismic deformation.  
 583 One could explain the early transient postseismic deformation with fault creep and the later  
 584 phase with relaxation in a Maxwell viscoelastic lower crust and upper mantle [e.g *Hearn et al.*,  
 585 2009; *Johnson et al.*, 2009]. However, postseismic deformation at distances greater than  $\sim 200$   
 586 km from the El Mayor-Cucapah epicenter can only be attributed to viscoelastic relaxation [*Freed*  
 587 *et al.*, 2007] and we have demonstrated that the far-field deformation cannot be explained with  
 588 a Maxwell rheology (Figure 7).

589 We found that a Zener rheology in the upper mantle with a transient viscosity of  $\sim 10^{18}$   
 590 Pa s does a noticeably better job at predicting far-field postseismic deformation. A general-  
 591 ization of the Zener viscoelastic model, schematically represented as several Kelvin elements  
 592 connected in series, is commonly used to describe seismic attenuation [*Liu et al.*, 1976]. The  
 593 highest viscosity needed to describe seismic attenuation is on the order of  $10^{16}$  Pa s [*Yuen and*  
 594 *Peltier*, 1982] which has a characteristic relaxation time on the order of days. Even though  
 595 our inferred transient viscosity is orders of magnitude larger than that required for seismic at-  
 596 tenuation models, the two models are not incompatible. Rather, the delayed elasticity in seis-  
 597 mic attenuation models occurs on such short time scales that it can be considered part of the  
 598 instantaneous elastic phase of deformation associated with the preferred Zener model in this  
 599 study.

600 Of course, a Zener rheology provides an incomplete descriptions of the asthenosphere,  
 601 as it does not have the fluid-like behavior required to explain isostatic rebound or convection  
 602 in the mantle [*O'Connell*, 1971]. *Yuen and Peltier* [1982] proposed a Burgers rheology with  
 603 a low transient viscosity ( $\eta_K \approx 10^{16}$  Pa s) and high steady-state viscosity ( $\eta_M \approx 10^{21}$  Pa  
 604 s) to describe both seismic attenuation and long term geologic processes. The justification of  
 605 a Burger's rheology mantle is further supported by laboratory experiments on olivine [*Chopra*,  
 606 1997]. *Pollitz* [2003] sought to describe postseismic deformation following Hector Mine with  
 607 a Burgers rheology mantle and they found a best fitting transient viscosity of  $1.6 \times 10^{17}$  Pa  
 608 s and steady-state viscosity of  $4.6 \times 10^{18}$  Pa s. While the Burgers rheology was introduced  
 609 as a means of bridging the gap between relaxation observed in long and short term geophys-  
 610 ical processes, the inferred steady state viscosity from *Pollitz* [2003] is still inconsistent with  
 611 the Maxwell viscosities inferred from studies on the earthquake cycle and Lake Bonneville.  
 612 The transient viscosity inferred by *Pollitz* [2003] is constrained by the earliest phase of post-  
 613 seismic deformation following the Hector Mine earthquake. While *Pollitz* [2003] ruled out deep  
 614 afterslip as an alternative mechanism based on inconsistent vertical deformation, it is still pos-  
 615 sible to successfully describe all components of early postseismic deformation following the  
 616 Hector Mine earthquake with afterslip at seismogenic depths [*Jacobs et al.*, 2002]. It is then  
 617 possible that the preferred rheologic model from *Pollitz* [2003] was biased towards inferring  
 618 a particularly low transient viscosity by neglecting to account for afterslip. This is in contrast  
 619 to the present study, where we have inferred a viscosity structure simultaneously with after-  
 620 slip. We also argue that a transient rheology is necessary to explain postseismic deformation;  
 621 however, our preferred transient viscosity of  $\sim 10^{18}$  Pa s in the upper mantle is an order of mag-

622 nitude larger than the transient viscosity found by *Pollitz* [2003]. The transient viscosity in-  
 623 ferred here is consistent with the results of *Pollitz* [2015], who reanalyzed postseismic data  
 624 following the Landers and Hector Mine earthquake allowing the first few months of transient  
 625 deformation to be described by afterslip. Since a Zener model is able to describe the avail-  
 626 able postseismic deformation following the El Mayor-Cucapah earthquake, any Burgers rhe-  
 627 ology with a steady-state viscosity that is  $\gtrsim 10^{20}$  Pa s, effectively infinite over five years, would  
 628 also be able to describe the postseismic deformation. Such a Burgers model might then be con-  
 629 sistent with the steady-state viscosities necessary for lake loading, interseismic deformation,  
 630 and mantle dynamics.

## 631 5 Conclusion

632 We have extracted a smoothed estimate of postseismic deformation following the El Mayor-  
 633 Cucapah earthquake from GPS displacement time series. Our estimated postseismic deforma-  
 634 tion reveals far-field (epicentral distances  $\gtrsim 200$  km) transient deformation which is undetectable  
 635 after about three years. Near-field deformation exhibits transience that decays to a sustained,  
 636 elevated rate after about one or two years. We found that near-field transient deformation can  
 637 be explained with shallow afterslip, and the sustained rate of near-field deformation can ei-  
 638 ther be explained with continued afterslip or viscoelastic relaxation in the lower crust. Far-field  
 639 transient deformation can be more definitively ascribed to viscoelastic relaxation at depths greater  
 640 than  $\sim 60$  km. Beneath that depth, a transient viscosity of  $\sim 1 \times 10^{18}$  Pa s is required to de-  
 641 scribe the rate of far-field deformation throughout the five years considered in this study. By  
 642 describing the available postseismic deformation with a transient rheology in the mantle, our  
 643 preferred model does not conflict with the generally higher steady-state viscosities inferred from  
 644 geophysical processes occurring over longer time scales.

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