Geodetically Inferred Coseismic and Postseismic Slip due to the M 5.4 31 October 2007 Alum Rock Earthquake

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Abstract On 31 October 2007 the M 5.4 Alum Rock earthquake occurred near the junction between the Hayward and Calaveras faults in the San Francisco Bay Area, producing coseismic and postseismic displacements recorded by 10 continuously operating Global Positioning System (GPS) instruments. The cumulative postseismic displacements over the four months following the earthquake are linearly related to the cumulative number of aftershocks and are comparable in magnitude to the coseismic displacements. The postseismic signal suggests that, in addition to afterslip at seismogenic depths, localized right-lateral/reverse slip occurred on dipping shallow fault surfaces southwest of the Calaveras. The spatial distribution of slip inferred by inverting the GPS data is compatible with a model in which moderate Calaveras fault earthquakes rupture locked patches surrounded by areas of creep, afterslip, and microseismicity (Oppenheimer *et al.*, 1990). If this model and existing Calaveras fault slip rate estimates are correct, a slip deficit remains on the 2007 Alum Rock rupture patch that may be made up by aseismic slip or slip in larger earthquakes. Recent studies (e.g., Manaker *et al.*, 2005) suggest that at depth the Hayward and central Calaveras faults connect via a simple continuous surface illuminated by the Mission Seismic Trend (MST), implying that a damaging earthquake rupture could involve both faults (Graymer *et al.*, 2008). If this geometry is correct, the combined coseismic and postseismic slip we infer for the 2007 Alum Rock event predicts static Coulomb stress increases of ~ 0.6 bar on the MST surface and on the northern Calaveras fault ~ 5 km northwest of the Alum Rock hypocenter.

Online Material: Data assessment, model geometry, and results of modeling using a more realistic fault geometry.

Introduction

Postseismic deformation is commonly observed following moderate and large earthquakes and may include a combination of poroelastic response (Jónsson *et al.*, 2003; Fialko, 2004), viscoelastic relaxation in the lower crust and mantle (Savage and Prescott, 1978; Thatcher, 1983; Deng *et al.*, 1998; Freed *et al.*, 2006), and afterslip on the fault plane (Bürgmann *et al.*, 2002; Hsu *et al.*, 2002, 2006). The moment release due to afterslip is commonly significantly less than that from coseismic slip (e.g., Hsu *et al.*, 2002; Freed *et al.*, 2006). However, some earthquakes on faults that creep interseismically have been observed to produce afterslip that releases moment equivalent to or greater than that of the mainshock (e.g., Oppenheimer *et al.*, 1990; Langbein *et al.*, 2006). The cumulative number of aftershocks and the cumulative geodetically-observed postseismic displacement following several earthquakes have been found to be linearly related (e.g., Hsu *et al.*, 2006; Savage and Yu, 2007; Savage and Langbein, 2008), suggesting that the two phenomena are

driven by the same underlying process (e.g., Perfettini and Avouac, 2004; Savage, 2007; Savage *et al.*, 2007).

The Calaveras fault, part of the San Andreas fault system in the San Francisco Bay Area, exhibits creep, abundant microseismicity (including repeating earthquakes), and moderate earthquakes such as the 1979 M 5.9 Coyote Lake and 1984 M 6.2 Morgan Hill events (Oppenheimer *et al.*, 1990; Schaff *et al.*, 2002; Manaker *et al.*, 2003). Most recently the 31 October 2007 M 5.4 Alum Rock earthquake nucleated on the Calaveras fault just south of its inferred juncture with the Hayward fault (Fig. 1) and ruptured to the southeast (Oppenheimer *et al.*, 2008; Seekins and Boatwright, 2008; see also the Data and Resources section).

The 2002 Working Group on California Earthquake Probabilities (hereafter referred to as WG02) divided the Calaveras fault into three segments with the transition from the northern to central segment occurring at the Calaveras reservoir (Fig. 2) near the junction between the Hayward

Figure 1. (a) Mapped surface traces of the Hayward and Calaveras shown as dark gray lines and other Quaternary active faults from Lienkaemper (1992) and Graymer *et al.* (2006) shown in light gray. The seismicity is from Waldhauser and Schaff (2008). The MST connects the Hayward and Calaveras faults at depth. The observed southward extent of the 1868 earthquake ground surface rupture in southern Fremont is also indicated (Lawson, 1908). Star, epicenter of 2007 Alum Rock earthquake; southern Calaveras fault (SCF). (b) The estimated locations of the major fault surfaces at 6 km depth inferred from seismicity and geophysical data are shown as dark gray lines. The seismicity plotted is for events at 6 km and deeper. Star, hypocenter of 2007 Alum Rock earthquake.

fault at depth and the Calaveras fault (Working Group on California Earthquake Probabilities [WGCEP], 2003). South of the junction the geologically estimated long-term slip rate for the central segment of the Calaveras fault, which includes the region of the 2007 Alum Rock event, is 14 ± 5 mm/yr (Kelson *et al.*, 1998). Surface creep measurements made between 1997 and 2001 on the central Calaveras at Coyote Ranch (Fig. 2) indicate a peak creep rate of ~ 14 mm/yr (Galehouse and Lienkaemper, 2003). In contrast, on the northern Calaveras fault the WG02 assigned a long-term slip rate of 6 ± 2 mm/yr based on geologic and geodetic data. At the Calaveras reservoir the surface creep rate is just 2.2 ± 0.5 mm/yr, although farther north on the Calaveras the observed rate is slightly higher (Galehouse and Lienkaemper, 2003). The decrease in slip rate from the central to northern segments suggests that part of the slip on the central Calaveras is transferred to the Hayward fault, which the WG02 estimates to have a long-term slip rate of 9 ± 2 mm/yr.

Although no surface creep rate measurements have been made on the Calaveras fault at the 2007 Alum Rock epicenter, Prescott *et al.* (1984) estimated a creep rate of 9.4 ± 0.4 mm/yr about 15 km southeast of the 2007 Alum Rock epicenter using the Grant Ranch trilateration network (Fig. 2). Manaker *et al.* (2003) used geodetic data to infer the spatial distribution of interseismic creep on the Calaveras fault. In their model the fault crept at 5.3 ± 2.7 mm/yr at depths above 2.5 km in the vicinity of the 2007 event with essentially no creep below that depth.

As can be seen in Figure 1a, at the Earth's surface there is no simple connection between the Hayward and Calaveras faults, but rather a ~ 5 km wide deformation zone with multiple oblique faults extending southeast from the south end of the creeping Hayward trace (Graymer *et al.*, 2007). However, recent studies (e.g., Ponce *et al.*, 2004; Manaker *et al.*, 2005; Williams *et al.*, 2005) have concluded, using a combination of seismicity, geological, and geophysical data, that the Hayward fault at depths below ~ 6 km connects in a simple way to the Calaveras fault as a structure following the Mission Seismic Trend (MST) (Fig. 1b). In this model at shallow depths the southern Hayward fault and subsidiary strands dip to the northeast to accommodate oblique compression in the contractional stepover from the Calaveras fault to the Hayward fault as mapped at the surface (Manaker *et al.*, 2005; Evans *et al.*, 2007; Hardebeck *et al.*, 2007; Watt *et al.*, 2008). The apparent continuity of the Hayward fault surface at depth through the stepover (Fig. 1b) is further supported by the observation of repeating microearthquakes at depth on the MST (Evans *et al.*, 2007). The presence of repeaters here suggests that fault creep extends along the Hayward fault through the junction and implies that coseismic (and potentially postseismic) slip could extend from the Calaveras to the Hayward fault (Graymer *et al.*, 2008). In an alternate model, the MST does not define the main active Hayward fault surface at depth that last ruptured in 1868. Rather, the active Hayward fault is a vertical structure underlying the Hayward creeping surface trace and extending southward from Fremont (Andrews *et al.*, 1993).

Figure 2. Map of the study area showing locations of GPS sites. Sites with bold italicized labels are those used in principal component analysis to extract the postseismic signal common to sites in the network. Star: epicenter of the 2007 Alum Rock event; Grant Ranch tri-lateration network (GR); former site of Coyote Ranch alignment array (CR); faults from [Graymer et al. \(2006\)](#); black: historic and Holocene activity; gray: Quaternary.

Spatial patterns of microseismicity on the Calaveras fault have been observed to be stable through multiple earthquake cycles, and aftershocks tend to occur in the same locations as interseismic microearthquakes, suggesting long-lived locked and creeping patches. These observations have been interpreted as evidence that the Calaveras fault slips through a combination of moderate earthquakes that nucleate near the base of the seismogenic zone to fill holes that do not exhibit microseismicity and aseismic slip above 5 km and in the deeper areas of interseismic microseismicity ([Oppenheimer et al., 1990](#); [Schaff et al., 2002](#)). The Calaveras fault has experienced a northwest-progressing series of moderate earthquakes since 1979 comprised of the 1979 M 5.9 Coyote Lake event, the 1984 M 6.2 Morgan Hill earthquake, and the 1988 M 5.0 Alum Rock event. This trend continued with the 2007 M 5.4 Alum Rock earthquake, which appears to have filled one of three aseismic holes on the Alum Rock section of the Calaveras fault (zone V of [Oppenheimer et al., 1990](#)) and to have been a repeat of a 1955 M_L 5.5 earthquake ([Oppenheimer et al., 1990](#); [Oppenheimer et al., 2008](#)).

Significant postseismic deformation was observed in electronic distance measurement (EDM) observations spanning the Coyote Lake and Morgan Hill earthquakes ([Prescott et al., 1984](#)). Although the amount of afterslip following the Coyote Lake earthquake was estimated to be only one-third the coseismic slip, the postseismic slip extended up to the Earth's surface ([Prescott et al., 1984](#)). Cumulative afterslip inferred from EDM data following the Morgan Hill earthquake was estimated to be 80% of the coseismic slip four months after the mainshock ([Prescott et al., 1984](#)) and had reached 180% after two years ([Oppenheimer et al., 1990](#)). In addition to the mainshock slip, the sizable afterslip observed following Calaveras fault earthquakes could further stress neighboring parts of the fault system, and [Oppenheimer et al. \(1990\)](#) suggested this as a factor contributing to spatial and temporal earthquake clustering in this region.

Despite its modest size, both coseismic offsets and postseismic transient deformation are visible in time series from continuous Global Positioning System (GPS) sites operating in the vicinity of the 2007 Alum Rock earthquake. In this study we assess the relative magnitudes of coseismic and

postseismic moment release in this event, the temporal relationship between aftershock decay and postseismic displacement, the spatial distribution of coseismic and postseismic slip, and the implications of remaining slip deficit on the Alum Rock rupture zone. We also calculate the change in Coulomb stress on the Hayward fault and northern Calaveras fault due to the coseismic and postseismic slip of the Alum Rock event considering two possible model fault geometries, one in which the Hayward and Calaveras Faults are connected at depth via a surface coincident with the MST and one in which the Hayward fault extends south as an essentially vertical plane under the southeastward extrapolation of its creeping mapped surface trace (Fig. 1).

Data and Modeling

We analyzed GPS observations collected from 1 January 2000 to 25 February 2008 by continuously recording stations of the Bay Area Regional Deformation (BARD) and Plate Boundary Observatory (PBO) networks (see [Data and Resources](#) section; Fig. 2). Stations with less than 1.5 yr of data were omitted. The data were processed using GIPSY/OASIS II software developed by NASA's Jet Propulsion Laboratory (JPL) to obtain daily position estimates for each station. We used satellite orbit and clock parameters provided by JPL and a precise point positioning approach ([Zumberge et al., 1997](#)) followed by ambiguity resolution ([Blewitt, 1989](#)). The solutions were transformed into the global ITRF2000 reference frame using a network of global tracking stations and subsequently into a North America fixed reference frame using the Euler pole defined by [Altamimi et al. \(2002\)](#).

Geodetic time series often exhibit common mode error ([Wdowinski et al., 1997](#)) that manifests itself as network-wide movement in the same direction on a given day. Following the approach outlined by [Hammond and Thatcher \(2007\)](#) we apply regional filtering to the time series. However, rather than applying a daily rotation we use a simple translation given the limited spatial extent of the network subset used. The sites used to define the regional filter were CMBB, DIAB, FARB, HOPB, MHCB, MONB, MUSB, SAOB, SUTB, CHO1. Two of these sites, MHCB and MONB, were omitted after the earthquake as they experienced coseismic and postseismic displacement. The result of regional filtering is a noticeable reduction of day-to-day scatter in the time series. However, even after regional filtering we noted a systematic northwest trend in the estimated displacements for stations relatively far from the epicenter (e.g., near San Juan Bautista and westward to the coast). For sites west of the Calaveras fault, northwestward displacement would be consistent with right-lateral slip, and if actually a common mode signal, could bias coseismic and postseismic displacement estimates. Therefore, we constructed a second filter using only sites greater than 40 km from the epicenter and on the east side of the Calaveras fault to avoid removing any fault-related signal. The filter was constructed following the approach of [Wdowinski et al.](#)

(1997) by finding the median scatter about the detrended data for each day. Even the subset of sites east of the Calaveras showed a small component of common mode displacement to the northwest, which would not be expected to be caused by right-lateral slip. This common mode filter was applied to the time series to reduce the possibility of common mode mapping into coseismic and postseismic displacement estimates.

We analyzed each station-component time series in order to estimate the coseismic and postseismic displacements and their standard errors. [Langbein \(2004\)](#) presents a maximum likelihood method for simultaneously fitting a time series with a predetermined temporal model and optimizing the noise model used. The time series were assumed to follow the form

$$\begin{aligned} d &= D_o + Rt + S \sin(2\pi ft) + C \cos(2\pi ft) & \text{for } t < T_{eq} \\ &= D_o + Rt + S \sin(2\pi ft) + C \cos(2\pi ft) + D_{eq} \\ &\quad + A \log(1 + (t - T_{eq})/\tau) & \text{for } t \geq T_{eq}, \end{aligned} \quad (1)$$

where D_o is a nominal position, R is the secular rate, t is time, S and C are the amplitudes of sine and cosine terms with frequencies f (where $f = 1$ and 2 yr^{-1}) that approximate seasonal effects, D_{eq} is an offset at the time of the earthquake (T_{eq}), and A is the amplitude of postseismic displacement following Omori's law (e.g., [Langbein et al., 2006](#)). The parameter τ represents the time delay of onset of the postseismic signal. The data noise were assumed to be a combination of white, flicker, and random walk; and the amplitudes of these components were optimized simultaneously with fitting the time series data ([Langbein, 2004, 2008](#)).

The Alum Rock earthquake occurred at 03:04:54 Coordinated Universal Time on 31 October 2007, and we estimated the GPS position solution for the day of the earthquake using only data collected on that date after the time of the event. Because we are using daily position solutions, it is possible that the coseismic data contain a signal from any immediate postseismic deformation that occurred in the hours following the earthquake.

Even the largest coseismic and postseismic displacements are relatively small (~ 4 mm), making it difficult to robustly estimate both A and τ for the Omori's law decay. Therefore, we use principal component analysis ([Savage, 1988](#); [Savage and Langbein, 2008](#)) to extract the temporal behavior common to sites near the epicenter (those with italicized labels in Fig. 2). Assuming an Omori's law decay, following the earthquake the principal components of the first mode of deformation (Fig. 3a), which represent the time dependence of this mode, are best fit with a τ of 2 days. We hold this value fixed in the time series analysis described previously. Because the optimal τ is close to the sampling rate of the data (one sample per day), this is likely an upper limit on the value of τ ; smaller values may also fit the data.

Figure 3. (a) Cumulative displacement as represented by the principal components of the first mode obtained through principal component analysis of the GPS time series data. (b) Principal components of the first mode as a function of the cumulative number of aftershocks, $N(t)$.

For each time series we compared the maximum likelihood estimator (MLE) for a model that included both an offset at the time of the earthquake and postseismic decay following Omori's law to a model that included only the offset (Langbein, 2004; ⑤ Table S1 is available in the electronic edition of *BSSA*). A significant improvement in MLE when using the first model was observed on at least one component (east or north) for only a small number of stations near the epicenter. For these stations we define the coseismic displacement as D_{eq} and the postseismic displacement as the cumulative displacement over the time period 31 October 2007–25 February 2008 as calculated from the estimated value of A and τ . For the remaining stations the inclusion of the postseismic decay did not produce noticeable improvement in fit as measured by the MLE, and in some cases the estimated postseismic displacement was of opposite sign to the coseismic because the optimization was fitting noise in the data. We conclude that the signal at these stations was too small to estimate both coseismic and postseismic displacement. For these stations we define the total offset as the D_{eq} estimated from time series that included postseismic data up to 25 February 2008; this quantity includes both coseismic and postseismic displacement. An example of time series after removal of the secular rate and periodic terms are shown for four GPS sites in Figure 4. Time series for all GPS sites used in this study are given in Table S1, and the observations are tabulated in Table S2,

⑤ both of which are available in the electronic edition of *BSSA*.

We use these displacement estimates in an inversion to infer coseismic and postseismic fault slip on rectangular dislocations in a homogeneous elastic half-space (Okada, 1985). The coseismic and postseismic model fault geometries (Fig. S2 and Table S3, ⑤ available in the electronic edition of *BSSA*) consist of several large rectangular subfaults representing the Calaveras and Hayward faults in the vicinity of the earthquake and are based on relocated microseismicity (Hardebeck *et al.*, 2007; Waldhauser and Schaff, 2008) and aftershocks of the 2007 event. The spatial relationship between the aftershock distribution and the portion of the model geometry in the vicinity of the aftershocks is shown in Figure 5. The postseismic fault geometry includes greater extents of both the Calaveras and Hayward faults than does the coseismic to accommodate possible afterslip over a larger area. Geologic mapping (e.g., Graymer *et al.*, 2006) has led to the identification of several subparallel dipping faults with oblique and reverse motion to the southwest of the Calaveras fault (Fig. 1), one or more of which may comprise the shallow manifestation of the Hayward fault zone extending southeast of the mapped Hayward creeping trace in the stepover region (Manaker *et al.*, 2005; Graymer *et al.*, 2007; Watt *et al.*, 2008). Triggered slip on these structures may have been responsible for aftershocks with reverse mechanisms that were observed between the

Figure 4. Cumulative GPS displacements over time for selected sites. For each station the east and north time series are shown with the estimated secular rate and seasonal terms removed. The coseismic displacement is shown by the heavy black vertical line at the time of the earthquake, and the estimated postseismic displacement as a function of time by the heavy gray line following the earthquake.

Figure 5. (a) Fault-face view of the central Calaveras and southern Hayward faults looking toward N55E showing 20 hr of Alum Rock aftershocks and six of the dislocation rectangles shown in Figure 7a. The hypocenters of three moderate earthquakes on the central Calaveras with years and magnitudes indicated at the top (stars), aftershocks following the 2007 Alum Rock event located by J. Hardebeck (personal comm., 2007) using a 3D velocity model (open circles), double-difference relocated seismicity from 1984–2003 (Waldhauser and Schaff, 2008) within 2 km of a model Hayward–Calaveras fault surface (small dots), rupture patch inferred for the 2007 Alum Rock event (ellipse) by Seekins and Boatwright (2008; see also Data and Resources section), the approximate extent of Calaveras locked zones (horizontal lines) described by Oppenheimer *et al.* (1990), location of cross section in (b) (vertical dashed line). Horizontal distances are measured in a S35E direction from the northernmost onshore point of the Hayward fault at Point Pinole (Lienkaemper, 1992). (b) Fault perpendicular cross section passing through 2007 Alum Rock hypocenter (star). Earthquakes within 2 km along strike are projected onto this cross section; other symbols as in (a).

Calaveras fault and the reverse slip Silver Creek fault to the southwest following the 1984 Morgan Hill earthquake (Cockerham and Eaton, 1987; Oppenheimer *et al.*, 1988). To accommodate the possibility of triggered postseismic slip on such a structure, the dipping shallow dislocations (outlined in red in Fig. S2, [E](#)) available in the electronic edition of *BSSA* are permitted right-lateral/reverse-oblique slip.

We simultaneously invert the coseismic offsets, cumulative postseismic displacements over the time period 31 October 2007–25 February 2008, and total horizontal displacements to find a self-consistent distribution of coseismic slip and afterslip. For the subset of stations for which the MLE of at least one component (east or north) showed a significant improvement when modeling the component time series with both coseismic and postseismic displacement, we use the coseismic and postseismic displacement estimates in the inversion. For the rest of the stations we use only the total offsets. In the case of P222, P226, and SODB, the estimated coseismic displacements in the north, east, and east components, respectively, were of opposite sign to the estimated postseismic, indicating that the signal-to-noise ratio was insufficient to discriminate between coseismic and postseismic displacement for those components. However, for the other component at each of these stations inclusion of postseismic decay in the time series model resulted in a significant improvement in MLE. Therefore, for these three sites we used the coseismic and postseismic displacement estimates for the component for which these quantities were well-resolved and the total displacement for the other component. The observation equation was set up as follows:

$$\begin{bmatrix} d_c \\ d_p \\ d_t \end{bmatrix} = \begin{bmatrix} G_{c1} & 0 \\ 0 & G_{p1} \\ G_{c2} & G_{p2} \end{bmatrix} \begin{bmatrix} s_c \\ s_p \end{bmatrix}, \quad (2)$$

where d_c are the coseismic, and d_p are the postseismic displacements for the station-components for which the coseismic and postseismic could be estimated separately, and G_{c1} and G_{p1} are the corresponding Green's functions relating displacement to fault slip. For the remaining station components, the total displacements (offsets at the time of the earthquake estimated from fitting time series data up to 25 February 2008) are given by d_t , and G_{c2} and G_{p2} are the corresponding Green's functions. The coseismic and postseismic slip estimates are given by s_c and s_p . Although the number of data exceeds the number of model parameters, the problem is mixed determined. To regularize the inversion we assume that the faults slip right laterally (with reverse slip possible in some cases as noted previously). This is enforced in the inversions using positivity constraints, and the solution to (2) found through constrained linear least squares.

Results

Oppenheimer *et al.* (1990, 2008) suggest that microseismicity on the Calaveras fault outlines a locked patch near the

2007 Alum Rock hypocenter (Fig. 5a). We will refer to this region as zone V following Oppenheimer *et al.* (1990). We first sought to test whether the geodetic data are consistent with coseismic slip being confined within this region, which has an area of $\sim 10 \text{ km}^2$. To do so, we restricted the coseismic model geometry to a single vertical rectangle 5 km long and 2 km in down-dip dimension that approximated the locked patch of zone V. Although the area outlined by seismicity is irregularly-shaped given the depth of the dislocation, the GPS data are insensitive to the details of its shape; a rectangular dislocation of the correct area should adequately represent this slip surface. To produce the moment release estimated from seismic data, $1.85 \times 10^{17} \text{ Nm}$, the slip on this area must be 62 cm. The fit to the coseismic GPS offsets using this model is shown in Figure 6. For the three sites for which it was only possible to estimate coseismic and postseismic displacement for one component (P222, P226, and SODB) the fit to the data is provided in Table 1. While the majority of coseismic data can be fit at the 95% confidence level, coseismic displacement at GPS site MHCB is significantly underfit, and coseismic displacement at GPS site MONB is significantly overpredicted.

Based on these initial results we expanded the coseismic fault model, using the relocated background microseismicity (Hardebeck *et al.*, 2007; Waldhauser and Schaff, 2008) as a guide to the strikes, dips, and locations of fault segments (Fig. 7a). We included subfaults representing the central

Figure 6. Observed (black) coseismic displacements and those predicted (white) from a model in which coseismic slip was only permitted on a dislocation of area 10 km^2 coinciding with zone V as defined by Oppenheimer *et al.* (1990). Star marks epicenter. Ellipses show 95% confidence region.

Table 1
Displacements at Stations Where Estimates of Coseismic and
Postseismic Displacements for Only One Component
Were Possible*

Station	Component	Coseismic		
		Observed (mm)	Predicted (mm)	2-Sigma Error (mm)
P222	East	−0.8	−1.1	0.5
P222	North			
P226	East			
P226	North	3.1	3.1	0.6
SODB	East			
SODB	North	0.4	1.0	1.8

*Assuming that coseismic slip is confined to zone V as defined by Oppenheimer *et al.* (1990), this table gives the observed and predicted displacements. Quoted uncertainties were obtained from time series analysis. All displacements in the table are fit at the 95% confidence level.

Calaveras fault from the surface to 12 km depth and extending ~15 km southeast of the hypocenter. This allows for the possibility of shallow slip and reflects the southeastward extent of aftershocks and the inference from seismic data that rupture directivity was to the southeast (Oppenheimer *et al.*, 2008). In addition, we included subfaults to allow for slip on the Hayward and northern Calaveras faults in the ~5 km northwest of the hypocenter. Finally, we found that including a dipping structure connecting the central Calaveras fault below 6 km to one of several faults mapped at the Earth's surface southwest of the Calaveras improved the fit to the coseismic displacement at MONB, so we included such a structure in our coseismic model. We term this structure the Hayward extension (Fig. 1a) because it bounds the collection of mapped strands in the deformation zone that extends southeast of the creeping Hayward trace in Fremont. The postseismic fault geometry included additional subfaults to the northwest and southeast (Fig. 7b).

The resulting coseismic and postseismic slip estimates are shown in Figure 7 (and listed in Table S3 (E) available in the electronic edition of *BSSA*); the fit to the data is shown in Figure 8, Table 2 (for P222, P226, and SODB), and in Table S2, (E) available in the electronic edition of *BSSA*, for the full dataset. The inferred coseismic slip extends southeast of zone V, and small amounts of slip are imaged above 6 km depth. This slip distribution fits all coseismic observations at the 95% confidence level. The moment release of the coseismic slip distribution is 1.84×10^{17} N m, in keeping with the estimate from seismic data. The postseismic slip distribution obtained in the joint inversion (Fig. 7b) is less successful than the coseismic in fitting the data (Fig. 8b; Table 2; Table S2, (E) available in the electronic edition of *BSSA*). Several sites on the southwest side of the fault are underfit, while the predicted displacement at MHCB well exceeds the modest postseismic displacement actually observed there. The predicted total offsets adequately fit the stations with the largest displacements, although the small signals observed at some stations more distant from the epicenter are not fit within errors (Fig. 8c).

The nonlinear nature of the inversion (due to the positivity constraints needed to regularize the problem) make it difficult to apply standard statistical tests such as the F test for linear inverse problems to assess the improvement in fit when the additional model parameters are included. However, the variance reduction, defined as $100 \times \{1 - [(\mathbf{r}^T \boldsymbol{\Sigma}^{-1} \mathbf{r}) / (\mathbf{d}^T \boldsymbol{\Sigma}^{-1} \mathbf{d})]\}$ where $\mathbf{r}$ is the residual between observed and predicted data, $\mathbf{d}$ is the data vector, and $\boldsymbol{\Sigma}$ is the data covariance matrix, can provide a measure of comparison. The variance reduction for the coseismic data using the model in which coseismic slip was confined to zone V was 80.4% while that for the model in Figure 7a was 96.6%.

Inverting the coseismic and postseismic displacements independently and omitting data from the station-components for which only the total slip could be estimated results in a coseismic slip distribution that is not substantially different from that in Figure 7a. However, in comparison to the distribution in Figure 7b the inferred postseismic slip is concentrated farther southeast, and ~8 cm of slip is imaged below 6 km depth on the Hayward fault at the northwest end of the model geometry. This slip distribution fits the observed postseismic displacements better than that presented in Figure 7b. However, summing the independently estimated coseismic and postseismic slip results in a model that significantly misfits the total displacements at some stations left out of the inversion. In particular, this total slip model predicts westward motion of P227 and eastward motion of SODB, both of which are clearly not seen in the time series (Figs. S11 and S1s, respectively, (E) available in the electronic edition of *BSSA*).

The fault geometry used in the inversions discussed previously is a highly simplified representation of the structurally complex junction (Graymer *et al.*, 2007) between the Calaveras and Hayward faults (Fig. 1). Some of the inability of the inferred postseismic slip to fit the data may be due to details of the complex fault geometry not captured by the first-order model presented previously. To explore this possibility, we repeated the inversions using a nonplanar fault surface subdivided into triangles approximately 2.5 km on a side. This geometry was delineated by relocated seismicity and mapped surface traces of the Hayward and Calaveras faults (Ellsworth *et al.*, 2000; Graymer *et al.*, 2006). (E) Details of the inversion are provided in the electronic edition of *BSSA*. The resulting coseismic slip distribution (Fig. S3, (E) available in the electronic edition of *BSSA*) fits the data better than does the coarse fault model, although the improvement is not significant (Fig. S4a and Table S4, (E) available in the electronic edition of *BSSA*). The postseismic model fits the data for MHCB, MONB, P218, P222, and P240 significantly better; however, displacements at stations southwest of the fault continue to be underpredicted (Fig. S4B and Table S4, (E) available in the electronic edition of *BSSA*). Given the low signal-to-noise ratio of this dataset and the inherent nonuniqueness in source inversions, particularly using a large number of small subfaults, we focus

Figure 7. Inferred slip distributions using preferred fault geometry (see text): (a) coseismic, (b) postseismic, (c) total slip. Origin of local coordinate system is at epicenter. White arrows indicate sense of slip of the footwall with respect to the hanging wall on subfaults that were allowed to have both right-lateral strike slip and reverse slip. For subfaults without arrows estimated slip was pure right-lateral. Alum Rock hypocenter (star), Central Calaveras (CC), Hayward Extension (HE), Northern Calaveras (NC), Mission Seismic Trend (MST).

Figure 8. Observed (black vectors) displacements and those predicted (white vectors) from the model shown in Figure 7: (a) coseismic, (b) postseismic, (c) total. Heavy gray lines mark surface trace of model subfaults. Star marks epicenter. Ellipses show 95% confidence region. In (a) and (b), stations without observed displacements and error ellipses are those for which total displacement was used in the inversion or for which coseismic and postseismic displacements were only used for one component; predicted displacements are shown for reference. In (c), stations with no observations (nor error ellipses) are those for which estimates of coseismic and postseismic (rather than total) displacement were used in the inversion or for which total displacement was only used for one component; predicted total displacements are shown for reference.

Table 2
Observed Displacements and Those Predicted Using the Fault Model in Figure 7*

Station	Component	Coseismic			Postseismic			Total		
		Observed (mm)	Predicted (mm)	2-Sigma Error (mm)	Observed (mm)	Predicted (mm)	2-Sigma Error (mm)	Observed (mm)	Predicted (mm)	2-Sigma Error (mm)
P222	East	-0.8	-0.7	0.5	-2.0	-1.1	0.6			
P222	North							0.6	0.5	0.4
P226	East							0.4	0.1	0.7
P226	North	3.1	3.0	0.6	2.4	2.3	0.7			
SODB	East							0.0	0.6	0.8
SODB	North	0.4	0.8	1.8	3.5	1.2	2.7			

*At stations for which estimated coseismic and postseismic displacements for only one component were used in the inversion. For the other component, the estimated total offset was used. Predicted total offset is sum of predicted coseismic and postseismic. Quoted uncertainties were obtained from time series analysis. All displacements in Table 2 are fit at the 95% confidence level except the east postseismic displacement at P222.

on the results using the coarse fault geometry in the following discussion.

Discussion

Oppenheimer *et al.* (1990) suggest, based on the persistence of spatial patterns of microseismicity before and after the Morgan Hill earthquake, that aseismic zones of the Calaveras fault between 5 and 10 km depth are locked patches with different constitutive properties than the zones of microseismicity separating them (that also experience aftershocks) and the shallow aseismic portions of the fault above 5 km, both of which creep and experience afterslip. Our findings are compatible with this interpretation. The majority of the coseismic observations can be fit with slip confined to zone V. However, the inclusion of minor amounts of slip on the central Calaveras fault above 6 km and on the shallow dipping Hayward extension significantly improves the fit to the data at MONB and MHC. This may reflect the occurrence of triggered slip immediately following the event on shallow dipping faults southwest of the Calaveras. The GPS data, while insufficient to strongly constrain the depth extent of the coseismic slip, are for the most part consistent with slip having been at the depths suggested by the aftershock distribution. Hellweg *et al.* (2008) used broadband data from the Berkeley Digital Seismic Network in a finite fault inversion for the Alum Rock event. Slip in their model is concentrated at 11–13 km depth below the hypocenter, deeper than most seismicity on the central Calaveras fault. They note, however, that the depth is poorly constrained and that they had difficulty fitting the data. Seekins and Boatwright (2008; see also the Data and Resources section) estimated rupture directivity of the 2007 event from the recorded peak ground acceleration. Using this and the magnitude of the event they inferred a rupture area (Fig. 5a) that agreed well with the aftershock distribution.

If we assume that the model of Oppenheimer *et al.* (1990) is correct and that the coseismic slip did occur in zone V (with minor triggered slip or immediate afterslip at shallow depths), we can use the area of this zone to calculate an upper bound on the amount of slip required to produce the recorded seismic moment of 1.85×10^{17} N m. Assuming an area of 10 km^2 for zone V implies that 62 cm of slip occurred there. The estimated long-term slip rate of the central Calaveras is $15 \pm 3 \text{ mm/yr}$ (WGCEP, 2003), and Oppenheimer *et al.* (2008) provide evidence that the 2007 Alum Rock event is a repeat of a 1955 earthquake. Given this range of slip rates and a 52 yr recurrence interval, the accumulated slip deficit at the time of the 2007 earthquake would have been 62–94 cm at the 1 – σ confidence level, implying that a slip deficit may have remained on zone V after the event.

One way that this slip deficit may be made up is that a given locked patch can rupture either by itself or in conjunction with neighboring patches during a larger event like the 1984 Morgan Hill earthquake. Konca *et al.* (2008) argue that this type of behavior can explain the pattern of large earth-

quakes on the Sumatra megathrust. There, two large earthquakes in 2007 are inferred to have ruptured only part of the estimated slip areas of major earthquakes that occurred in 1797 and 1833 and to have only released about 25% of the moment deficit that had accumulated since the 1833 event. They attribute this to spatially variable interseismic coupling and local stress variations caused by previous events that result in temporary slip barriers. Similarly, Murray and Langbein (2006) found that slip deficit remained on portions of the Parkfield segment of the San Andreas fault that had slipped in the two most recent $M \sim 6$ earthquakes at that location.

Alternatively, the potentially locked zones outlined by seismicity on the Calaveras fault may actually experience both seismic and aseismic slip. Chen and Lapusta (2009) present a model for small repeating earthquakes on asperities surrounded by a creeping fault that requires as much as 99% of slip on a source asperity to occur aseismically in order to reproduce the scaling of seismic moment with recurrence time observed for repeating earthquakes on the San Andreas fault at Parkfield. In this model the percentage of aseismic slip (that included afterslip) was inversely proportional to seismic moment of the repeater, and the maximum magnitude simulated was 3.7. Uchida *et al.* (2007) describe a similar phenomenon for larger events in northeast Japan. There a rupture patch producing repeating $\sim M$ 4.9 earthquakes appears to also experience repeating microearthquakes during the time between the moderate events. The rupture patch of the microearthquakes must be reloaded repeatedly during the larger patch's earthquake cycle, which Uchida *et al.* interpret as evidence that the rupture area of the larger events creeps between $\sim M$ 4.9 earthquakes. The peak postseismic slip in our results is seen over the 10 km southeast of the hypocenter (Fig. 7). To be compatible with the hypothesis of Oppenheimer *et al.* (1990) this would imply that postseismic slip occurred in the aseismically slipping areas where microseismicity clusters at depths below 6 km. It is also possible, though, that the inferred slip is, at least in part, a result of afterslip on the rupture patch itself and/or neighboring zone IV. The GPS data cannot resolve spatial variations in slip over areas as small as zone V. However, if a slip deficit typically remains on the rupture zones of moderate Calaveras earthquakes (as identified by spatial patterns of seismicity), whether this deficit is made up by afterslip, interseismic creep, or in larger earthquakes has important implications for the predicted ground shaking in the vicinity of the fault.

Prescott *et al.* (1984) inferred that afterslip of the Morgan Hill earthquake occurred at shallower depths than the coseismic slip; these authors also noted that afterslip following the Coyote Lake earthquake broke the surface. However, Templeton *et al.* (2009) used repeating microearthquakes to image the spatial distribution of cumulative afterslip 6 and 18 months following the Morgan Hill event. They found that geodetic line-length changes were underpredicted by the inferred slip but that the fit was improved with the inclusion of additional hypothetical afterslip below 10 km.

In the inversion results presented here for the 2007 Alum Rock event, the postseismic slip was concentrated at depths greater than 6 km. We repeated the inversion using a postseismic fault geometry in which slip was only allowed above 6 km depth. To avoid the coseismic slip compensating for reduced postseismic slip area, we constrained the coseismic moment to be the moment estimated from seismic data. The resulting postseismic slip distribution is shown in Figure 9a; the fit to the postseismic data is given in Figure 10a and for P222, P226, and SODB in Table 3. Using this model provides a significantly worse fit to the postseismic data than does the model of Figure 7b, degrading the fit in terms of both the

magnitude and direction of displacement, particularly for P217, P218, and P228, as well as the east component at P222. The variance reduction for the postseismic data using the model in Figure 9a was 80.6% whereas that for the model in Figure 7b was 91.9%. The interseismic creep rates that Manaker *et al.* (2003) inferred above 6 km depth on the Calaveras fault near Alum Rock were, although not well-constrained, relatively low compared to the estimated long-term fault slip rate. This implies substantial moment deficit existed on shallow portions of the fault prior to the 2007 Alum Rock event, and our results suggest this deficit was not reduced by afterslip due to that earthquake. It is possible

Figure 9. (a) Postseismic slip distribution inferred using a fault geometry in which no postseismic slip was permitted below 6 km depth. (b) Postseismic slip distribution inferred using the same fault geometry as in Figure 7b but with the postseismic moment constrained to be no more than the coseismic moment, 1.85×10^{17} N m, determined from seismic data. In (a) and (b) reverse slip was permitted on some subfaults (see text) in addition to right-lateral strike slip, however, reverse slip estimates were zero. Labels are the same as in Figure 7.

Figure 10. Observed (black vectors) displacements and those predicted (white vectors) from (a) the model shown in Figure 9a and (b) the model shown in Figure 9b. Heavy gray lines mark surface trace of model subfaults. Other symbols are the same as in Figure 8.

that this deficit will be made up via the coseismic or postseismic slip of larger future events.

The improvement in fit to the postseismic data obtained when using the nonplanar fault geometry consisting of small subfaults (Fig. S4b, [E](#) available in the electronic edition of *BSSA*) may result largely from modeled slip on isolated subfaults. However, geologic mapping ([Graymer et al., 2006](#)) has identified numerous dipping faults subparallel to, and in some cases possibly intersecting, the Calaveras fault to the southwest. Our model geometry includes a single

Table 3

Observed Postseismic Displacements and Those Predicted Using the Fault Model in Figure 9a*

Station	Component	Postseismic		
		Observed (mm)	Predicted (mm)	2-Sigma Error (mm)
P222	East	-2.0	-0.6	0.6
P222	North			
P226	East			
P226	North	2.4	2.7	0.7
SODB	East			
SODB	North	3.5	0.8	2.7

*At stations for which estimated coseismic and postseismic displacements for only one component were used in the inversion. For the other component, the estimated total offset was used. See Table 2 footnote for details. All displacements in Table 3 are fit at the 95% confidence level except the east postseismic displacement at P222.

dipping shallow fault surface southwest of the Calaveras fault. It is plausible that the postseismic displacements observed at sites west of the fault, which are large relative to that of GPS site MHCB (Fig. 8b), are the combined response to triggered shallow slip on multiple fault strands. Seismic tomography has revealed low velocity material in the Evergreen Basin southwest of the Calaveras fault (roughly between the southern end of the San Francisco Bay and GPS site P218) as well as a high velocity body northeast of the Calaveras near MHCB ([Thurber et al., 2007](#)). Allowing for across-fault or vertical variations in elastic properties when calculating the Green's functions for our model might also have improved the fit to the postseismic data.

The GPS observations and modeling suggest that, as seen for other events on creeping faults (e.g., [Sharp et al., 1982](#); [Prescott et al., 1984](#); [Langbein et al., 2006](#)), the postseismic deformation following the Alum Rock earthquake was relatively large. At most stations the cumulative postseismic displacement over the four months following the event was comparable in magnitude to the coseismic displacement (Fig. 8 and Fig. S1, [E](#) available in the electronic edition of *BSSA*), although at MHCB the postseismic signal was relatively modest. The estimated postseismic moment release for the model shown in Figure 7 (3.1×10^{17} N m) exceeds the seismic moment (1.85×10^{17} N m). To investigate the relative amounts of coseismic and postseismic moment release, we repeated the inversions with the added constraint that the postseismic moment could not exceed the seismic moment. The resulting postseismic slip distribution (Fig. 9b) significantly underpredicts the observed postseismic displacement at P217, P218, and the east component of P222 as well as degrades the fit at LUTZ and P228 (Fig. 10b and for P222, P226, and SODB, Table 4). In contrast, MONB and MHCB are fit better with the inclusion of the moment constraint, suggesting that the displacement magnitudes at these two stations relative to those of the rest of the network cannot be fit with the simple model fault geometry and material properties used here. The variance reduction for postseismic data using the model in Figure 9b

Table 4
Observed Postseismic Displacements and Those Predicted
Using the Fault Model in Figure 9b*

Station	Component	Postseismic		
		Observed (mm)	Predicted (mm)	2-Sigma Error (mm)
P222	East	-2.0	-0.9	0.6
P222	North			
P226	East			
P226	North	2.4	2.1	0.7
SODB	East			
SODB	North	3.5	0.8	2.7

*At stations for which estimated coseismic and postseismic displacements for only one component were used in the inversion. For the other component, the estimated total offset was used. See Table 2 footnote for details. All displacements in Table 4 are fit at the 95% confidence level except the east postseismic displacement at P222.

was 86.8% (compared to the 91.9% using the model of Fig. 7b). When the constraint on maximum postseismic moment is further reduced to 75% of the coseismic moment, LUTZ, P228, and SODB are no longer fit at 95% confidence, and the variance reduction for the postseismic data is 82.3%. Based on these observations we conclude that the postseismic moment release was comparable to the seismic moment.

Oppenheimer *et al.* (1990) and Templeton *et al.* (2009) suggest that static stress changes due to afterslip that fills in the area surrounding mainshocks may be a significant factor in the occurrence of subsequent mainshocks and perhaps is responsible for the series of moderate earthquakes that have been observed on the Calaveras fault. The interpretation that at depth the Hayward fault is a continuation of the Calaveras fault, rather than the junction between the two faults being a left stepover as seen at the Earth's surface, has highlighted the possibility that an earthquake could rupture from one fault to the other, thus, potentially increasing the seismic hazard in this region (Manaker *et al.*, 2005; Evans *et al.*, 2007; Graymer *et al.*, 2008). We calculated the change in Coulomb failure function (ΔCFF) on the Hayward and Calaveras faults due to the total slip distribution shown in Figure 7c. Two model geometries were used for the Hayward fault at depth, one in which it forms a simple connection with the Calaveras fault as a continuous dipping surface that follows the MST, and one in which the Hayward–Calaveras transition is modeled as a discrete left stepover with the vertical Hayward fault under the mapped surface trace extending parallel to the Calaveras fault for 10 km southeast of Fremont (Fig. 1). The effective coefficient of friction used in the calculations was 0.4, which is in the range of plausible

Figure 11. Coulomb stress changes due to slip distribution shown in Figure 7c on two possible models of Hayward fault geometry. Symbols are the same as in Figure 7. (a) Model in which the Hayward and Calaveras faults connect at depth via the MST. (b) Model in which the Hayward fault is assumed to extend southeastward as a vertical strike-slip fault.

values inferred in other studies (Harris, 1998). If the Hayward fault at depth connects to the Calaveras fault via the MST, stress changes due to the combined coseismic and postseismic slip are estimated to be approximately 0.6 bars on the MST and on the northern Calaveras fault ~5 km northwest of the Alum Rock hypocenter (Fig. 11a). A stress increase of ~0.15 bar is predicted on the MST near the southeast end of the 1868 Hayward earthquake surface rupture projected down-dip to the MST. Alternatively, if the active Hayward fault at depth is an essentially vertical surface beneath the 1868 surface rupture, a ~0.2 bar stress decrease is predicted at ~10 km depth at its southeast end (Fig. 11b). In either case, Coulomb stress increases exceeding 2 bars are calculated for parts of the central and northern Calaveras fault adjacent to the slipped area, including zone VI of Oppenheimer *et al.* (1990). It has been suggested that static stress changes as low as 0.1 bar can influence aftershock locations (Harris, 1998), although stress changes associated with triggered mainshocks have been inferred to be on the order of a few bars (e.g., Stein, 1999). The implications of the predicted stress changes due to the 2007 Alum Rock earthquake depend strongly on the nature of the junction between the Hayward and Calaveras faults, whether they form a continuous fault surface at depth, and the extent to which deformation at this fault junction is accommodated by secondary faults and nonrecoverable deformation (e.g. King and Nábělek, 1985). It should be noted that these calculations do not account for time variable stress or the temporal evolution of postseismic slip.

Several studies have noted a linear relationship between the cumulative number of aftershocks and cumulative postseismic displacement recorded geodetically (e.g., Perfettini and Avouac, 2004; Savage *et al.*, 2007). Perfettini and Avouac (2004) hypothesize that both aftershocks and postseismic displacement result from slip on the brittle creeping fault zone down dip of the coseismic rupture. Savage *et al.* (2007) showed that the linear relationship was observed for five earthquakes including the 2004 Parkfield event for which Langbein *et al.* (2006) inferred that a considerable portion of the afterslip occurred at seismogenic depths. Following the principal component analysis approach of Savage and Langbein (2008) we compared the cumulative number of aftershocks (see Data and Resources section) to the temporal function of the first mode of postseismic displacement (that captures the dominant temporal dependence common to the network of stations as a whole) for the 2007 Alum Rock earthquake and found them to be linearly related (Fig. 3b). As discussed by Savage *et al.* (2007), several distinctly different physical mechanisms could produce the observed similarity in time dependence of aftershock decay and postseismic relaxation, and thus, the underlying processes are still unclear.

Conclusions

Continuous GPS data record coseismic and postseismic deformation associated with the 2007 *M* 5.4 Alum Rock

event. As with earlier events on the Calaveras fault and other creeping faults, the postseismic displacement (and inferred postseismic moment release) is relatively large, likely comparable to the coseismic. The geodetically inferred spatial distribution of coseismic slip is compatible with the model of Oppenheimer *et al.* (1990) in which asperities below 5 km depth are the source of moderate earthquakes on the Calaveras. The majority of afterslip is inferred to have occurred at similar depths to the coseismic slip which, to be compatible with the model of Oppenheimer *et al.* (1990), would imply that afterslip took place in regions of microseismicity and aftershocks. Details of the postseismic displacement field suggest that triggered slip may have occurred on one or more east-dipping shallow fault surfaces southwest of the Calaveras fault. The temporal relationship between postseismic deformation and aftershocks supports the possibility that both arise from the same underlying process. Under certain assumptions regarding the geometry of the Hayward–Calaveras junction, the combined coseismic and postseismic slip inferred from the GPS data predict Coulomb stress increases in excess of 0.5 bar on portions of the northern Calaveras fault and a fault surface coincident with the MST within ~5 km of the hypocenter. If the model of Oppenheimer *et al.* (1990) and estimates of the interseismic slip rate on the Calaveras fault are correct, a slip deficit remains on the 2007 Alum Rock rupture patch, which could be made up if this zone participates in larger earthquakes or if it experiences aseismic as well as coseismic slip.

Data and Resources

The GPS data used in this article came from sites of the BARD and PBO networks. Data from these networks can be obtained from <http://www.ncedc.org/bard/> and <http://pboweb.unavco.org/>, respectively (last accessed 15 July 2009). The aftershock data plotted in Figure 11 were obtained from the Northern California Earthquake Data Center (<http://www.ncedc.org/ncedc/catalog-search.html>, last accessed 15 July 2009). Maps were plotted using the Generic Mapping Tools version 4.2.1 (www.soest.hawaii.edu/gmt/; Wessel and Smith, 1998; last accessed 15 July 2009). Information regarding the rupture patch inferred by Seekins and Boatwright (2008) for the 2007 Alum Rock event and referred to in the Discussion section and in the Figure 5 caption is available at <http://earthquake.usgs.gov/regional/nca/rupture/aftershocks/?id=20071030-2004> (last accessed 15 July 2009).

Acknowledgments

We wish to thank Roland Bürgmann, Russell Graymer, John Langbein, David Oppenheimer, David Ponce, David Schmidt, Janet Watt, and an anonymous reviewer for useful discussions.

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Manuscript received 17 January 2009