Lithospheric rheology constrained from twenty-five years of postseismic deformation following the 1989 M_w 6.9 Loma Prieta earthquake

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The October 17, 1989 M_w 6.9 Loma Prieta earthquake provides the first opportunity of probing the crustal and upper mantle rheology in the San Francisco Bay Area since the 1906 M_w 7.9 San Francisco earthquake. Here we use geodetic observations including GPS and InSAR to characterize the Loma Prieta earthquake postseismic displacements from 1989 to 2013. Pre-earthquake deformation rates are constrained by nearly 20 years of USGS trilateration measurements and removed from the postseismic measurements prior to the analysis. We observe GPS horizontal displacements at mean rates of 1-4 mm/yr toward Loma Prieta Mountain until 2000, and ~2 mm/yr surface subsidence of the northern Santa Cruz Mountains between 1992 and 2002 shown by InSAR, which is not associated with the seasonal and longer-term hydrological deformation in the adjoining Santa Clara Valley. Previous work indicates afterslip dominated in the early (1989-1994) postseismic period, so we focus on modeling the postseismic viscoelastic relaxation constrained by the geodetic observations after 1994. The best fitting model shows an

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elastic 19-km-thick upper crust above an 11-km-thick viscoelastic lower crust with viscosity of $\sim 6 \times 10^{18}$ Pa s, underlain by a viscous upper mantle with viscosity between 3×10^{18} and 2×10^{19} Pa s. The millimeter-scale postseismic deformation does not resolve the viscosity in the different layers very well, and the lower-crustal relaxation may be localized in a narrow shear zone. However, the inferred lithospheric rheology is consistent with previous estimates based on post-1906 San Francisco earthquake measurements along the San Andreas fault system. The viscoelastic relaxation may also contribute to the enduring increase of aseismic slip and repeating earthquake activity on the San Andreas fault near San Juan Bautista, which continued for at least a decade after the Loma Prieta event.

Keywords: Loma Prieta earthquake, postseismic displacement, lithospheric rheology, viscoelastic relaxation, San Francisco Bay Area

1. Introduction

Periods of accelerated postseismic deformation following large earthquakes reflect the response of the Earth's lithosphere to sudden coseismic stress changes. Thus, detailed geodetic measurements of postseismic relaxation effectively probe the rheology of rocks and faults at depth [Bürgmann and Dresen, 2008]. Transient post-earthquake relaxation includes contributions from (1) fault afterslip above [Bürgmann, *et al.*, 1997; Johnson *et al.*, 2006; Freed and Bürgmann, 2004] and below [Tse and Rice, 1986] the base of the seismogenic zone, (2) viscous flow in the lower crust and upper mantle [Pollitz *et al.*, 2001; Burgmann and Dresen, 2008; Thatcher and Pollitz, 2008], (3) poroelastic rebound in the upper crust due to fluid flow in response to coseismic pressure changes [Jónsson *et*

48 *al.*, 2003; *Peltzer et al.*, 1996], and (4) recovery of coseismic dilatancy by fault-zone
49 compaction [*Savage et al.*, 1994; *Savage and Svarc*, 2010; *Fielding et al.*, 2009]. For
50 relatively small earthquakes, shallow and rapidly decaying afterslip and poroelastic
51 relaxation dominate the observed postseismic transients and contributions from below the
52 seismogenic zone are difficult to resolve [*Jónsson et al.*, 2003; *Pollitz et al.*, 1998].
53 Depending on the magnitude of the earthquake and the viscosity structure of the
54 lithosphere, viscous relaxation at depth dominates transient deformation, especially at
55 larger distances from the coseismic rupture [*Freed et al.*, 2012].

56 Much of our knowledge of the earthquake cycle and the rheology of the deep San
57 Andreas fault (SAF) system in central California is based on interpretation of geodetic
58 measurements collected in the decades following the 1906 San Francisco earthquake that
59 ruptured a ~400-km-long section of the SAF [*Kenner and Segall*, 2003]. *Kenner and*
60 *Segall* [2003] consider data collected between 1906 and 1995 in a systematic evaluation
61 of various first-order models of lower-crustal and upper-mantle structure and rheology.
62 They find that models incorporating vertical viscous shear zones in the lower crust within
63 an otherwise elastic or viscous layer provide a good fit to the geodetic data, and are
64 consistent with seismic studies that suggest that narrow fault zones extend through the
65 entire crust [*Henstock et al.*, 1997]. The occurrence of the M_w 6.9 Loma Prieta earthquake
66 provides the first opportunity since 1906 to study postseismic relaxation. Our work here
67 presents the first attempt of measuring over 20 years of Loma Prieta postseismic
68 deformation with modern space geodetic tools to estimate rheological parameters in the
69 region.

70 The deformation measured with GPS immediately following the Loma Prieta
71 earthquake revealed significant postseismic contraction and right-lateral shear across the
72 southern Santa Cruz Mountains northeast of the SAF [*Savage et al.*, 1994; *Bürgmann et*
73 *al.*, 1997]. The localized nature of the transient displacement field indicates relatively
74 shallow deformation sources. The measurements of the first five years can be interpreted
75 to be due to aseismic right-oblique fault slip on or near the coseismic rupture, as well as
76 thrusting up-dip of the buried rupture within the Foothills thrust belt [*Bürgmann et al.*,
77 1997]. Analysis of the time-varying nature of the deformation signal suggests that the
78 shallow transient thrusting ceased in 1992 while resolvable oblique shear at seismogenic
79 depths may have persisted through 1994 [*Segall et al.*, 2000]. Alternatively, afterslip on
80 the downdip extension of the coseismic rupture plus a fault-normal collapse of the
81 rupture zone can explain the observed surface motions [*Savage and Svarc*, 2010].
82 Analysis of the GPS measurements did not resolve a significant contribution of lower
83 crustal or upper mantle relaxation processes, during the first five years following the
84 event [*Pollitz et al.*, 1998].

85 Since the Loma Prieta earthquake, several studies have focused on the interseismic
86 deformation in the Bay Area that accommodates the secular motion between the North
87 American plate and the Pacific plate. *d'Alessio et al.* [2005], *Johanson and Bürgmann*
88 [2005], *Bürgmann et al.* [2006], and *Johnson and Fukuda* [2010] estimate Bay Area
89 interseismic deformation models based on campaign and continuous GPS measurements
90 after 1994 (Fig. 1a). Although the model-predicted displacements generally agree with
91 most of the GPS measurements, a systematic model misfit exists near the Loma Prieta
92 earthquake area in these studies (Fig. S1). This result indicates a mechanism that cannot

be predicted by steady interseismic strain accumulation across the regional fault system. In addition, *Bürgmann et al.* [2006] found a model residual indicating subsidence along the restraining bend of the SAF in which the Loma Prieta earthquake occurred in their joint analysis of the horizontal GPS velocity field and Persistent Scatterer Interferometric Synthetic Aperture Radar (PSInSAR) data.

In this study, we argue that this systematic residual is due to viscoelastic relaxation (VER) of the lower crust and upper mantle following the Loma Prieta earthquake, a process that is still acting in this area after 1994. The pre-Loma Prieta interseismic deformation is estimated using precise trilateration measurements collected since the early 1970s [*Lisowski et al.*, 1991]. To obtain a ~25-year-long post-Loma Prieta observation, we combine campaign and continuous GPS measurements collected during 1989.8 – 2014 [*Segall et al.*, 2000; *USGS*, 2014]. We also generate an 18-year-long InSAR time series between 1992 and 2010 with data from both the ERS-1/2 and Envisat satellites. We use these geodetic data to determine contributions from both afterslip and VER to the transient deformation and estimate viscosities in the lower crust and uppermost mantle of the Bay Area.

2. Geodetic data

The U. S. Geological Survey (USGS) surveyed trilateration networks in the San Francisco Bay area, from 1973 [*Lisowski et al.*, 1991] until as late as three years after the Loma Prieta event. Electronic Distance Measurement (EDM) observations detect changes in distance between station pairs that have been used to determine the secular velocity field between the North American Plate and the Pacific Plate [*Lisowski et al.*, 1991].

116 *Bürgmann* [1997] used the trilateration data to solve for the horizontal interseismic
117 velocity field in the southern San Francisco Bay Area and developed a dislocation model
118 inverted from the trilateration line-length change rates. This model is composed of 78
119 individual fault segments, and each fault segment has a uniform-slip dislocation in an
120 elastic, homogenous, and isotropic half-space. We adopt the same fault dislocations and
121 slip rates to forward model the interseismic velocity before the Loma Prieta earthquake
122 for each GPS station (white arrows in Fig. 1a).

123 *Segall et al.* [2000] analyzed daily GPS solutions at 62 stations collected from
124 1989.8 to 1998.3 (red dots in Fig. S2), and used a Network Inversion Filter to model
125 time-dependent afterslip of the Loma Prieta earthquake. They modeled relative baseline
126 vectors by subtracting the position of a frequently observed reference site (LP1 on Loma
127 Prieta Mountain) from the other simultaneously observed positions, in order to minimize
128 the errors in the absolute position determinations due to translational biases in the
129 reference frame. Later on, *d'Alessio et al.* [2005] published the Bay Area velocity
130 unification (BAVU) solutions based on more than 200 campaign and continuous GPS
131 measurements collected between 1993 and 2003 (Fig. 1a;
132 <http://seismo.berkeley.edu/~burgmann/RESEARCH/BAVU/>). Most of the GPS stations
133 used in *Segall et al.* [2000] continued to be surveyed for BAVU (Fig. 1a). In addition, the
134 USGS [*Savage and Svarc*, 2010] also resurveyed many of the GPS stations through early
135 2013 (http://earthquake.usgs.gov/monitoring/gps/SFBayArea_SGPS/). Data from the Bay
136 Area Regional Deformation (BARD) and Plate Boundary Observatory (PBO) continuous
137 GPS networks complement the campaign measurements
138 (<http://earthquake.usgs.gov/monitoring/gps/SFBayArea/>). To generate a continuous

postseismic displacement time series, we combine GPS measurements from *Segall et al.* [2000], BAVU [*d'Alessio et al.*, 2005], the USGS campaign survey data, and the BARD and PBO observations (see Section 3.2 for more detail). The main challenge of combining the different GPS data sets lies in their heterogeneous observation periods and variable uncertainties. In addition, it is increasingly difficult to separate the interseismic secular motion from later-stage post earthquake measurements.

We use 53 European Space Agency (ESA) ERS-1/2 SAR descending acquisitions (Track: 70) between 1992 and 2006 and 46 Envisat ASAR descending acquisitions (Track: 70) between 2005 and 2010 (see Tables S1 and S2 for all acquisitions). All interferograms are generated using ROI_PAC 3.0 [*Rosen et al.*, 2004], and we use the 90m Shuttle Radar Topography Mission (SRTM) Digital Elevation Model (DEM) to correct the phase due to topography. Snapu 1.4.2 [*Chen and Zebker*, 2002] is used for the phase unwrapping. We use a small baseline subset (SBAS, *Berardino et al.*, 2002) method to generate time series of stable surface point scatterers. We generate interferograms that have less than 250 m perpendicular orbit separation and three years temporal baselines, respectively. We consider a point scatterer to be stable if the phase measurement at that point maintains its spatial correlation higher than 0.4 in at least 50% of all interferograms. We processed 392 interferograms (Fig. S3) from the 99 acquisitions, so solving the time series turns out to be an over-determined inverse problem [*Huang et al.*, 2014].

3. Postseismic deformation

3.1 Pre- and postseismic baseline-length measurements

We combine the 1970-1990 EDM data [Lisowski *et al.*, 1991], 1989.8-1998
 campaign GPS measurements [Segall *et al.*, 2000], and 1993-2014 USGS campaign and
 PBO continuous GPS (CGPS) data to estimate the pre-, co-, and postseismic distance
 changes between station pairs EAUN-LOMA, ALLI-LOMA, and HAML-LOMA (see
 Fig. 1 for station locations). The distance change is a projection of horizontal
 displacement between two stations onto the azimuth between stations. Figures 2a, c & e
 show the line-length changes of different station pairs where black, blue, red, and pink
 points represent the EDM, 1989.8–1998 GPS, 1993–2014 campaign GPS, and post-2000
 CGPS data, respectively. We use the interseismic model proposed by Bürgmann [1997]
 (white arrows in Fig. 1a) to remove secular motion contribution from the post-earthquake
 distance change measurements (Figs 2b, d & f). For station pair EAUN – LOMA, the
 preseismic lengthening rate is 6.1 mm/yr (positive indicates lengthening), and the
 postseismic rate falls back to the preseismic rate after 2000 (< 1 mm/yr; see Fig. 2b). For
 station pair ALLI – LOMA, the preseismic shortening rate is -10.1 mm/yr and stable
 postseismic shortening rate until 2012 (Fig. 2d). Note that we extend the time series by
 adopting nearby CGPS stations MONB and SODB to replace ALLI and LOMA after
 2000. A similar approach is used for pair HAML (MHCB) – LOMA (LP1) (Fig. 2e). In
 this pair, there is an additional distance change in 1984 due to the M 6.2 Morgan Hill
 earthquake on the Calaveras fault. The postseismic transient of this event also affected
 the shortening rate prior to the 1989 Loma Prieta event. In fact, the Loma Prieta
 earthquake contributed less coseismic shortening of this baseline than the Morgan Hill
 earthquake. After removing secular motion, there is still ~ 2 mm/yr shortening rate
 residual before the Morgan Hill event (Fig. 2f). Due to fewer measurements on LP1X, we

choose nearby CGPS stations SODB and P215, both about 9 km far away from LOMA for this comparison (Figs S3a-b). We further examine the distance changes between HAML, LOMA, and AMER in Supplementary Information S1 to explore Morgan Hill earthquake related displacement between these stations. The station AMER is west of the Calaveras fault, and the BARD GPS station LUTZ was built near AMER and surveyed continuously since 1996 (Fig. 1a).

3.2 Postseismic GPS displacement time series

To estimate the GPS time series since the Loma Prieta earthquake, we focus on benchmarks surveyed by *Segall et al.* [2000] and *USGS* [2015]. The uncertainty of GPS positions relative to LP1 is 3-6 mm for the *Segall et al.* [2000] dataset, and ~3 mm for the *USGS* dataset based on uncertainty estimations in their solutions. Due to higher GPS positioning uncertainty prior to 1995, we calculate site displacements at each epoch relative to station LP1 (also known as LP1X or LOMA) [*Segall et al.*, 2000]. We choose LP1 because it is densely surveyed during the 1989 and 2010 time period (Fig. S2a). We do not reprocess the GPS data, and we assume 1994 as the reference time because all GPS data sets include common measurements between 1994 and 1995. As a result, we are able to combine the solutions in a consistent local framework. We use the *Bürgmann* [1997] model to estimate and remove the interseismic contribution to the station displacements. We use logarithmic fits to the east and north components of the postseismic-only time series at each GPS station with a logarithmic function:

$$D(t) = D_{x,y} \ln(1+t / \tau), \quad [1]$$

where $D_{x,y}$ contains the estimated postseismic amplitudes of the east and north displacement components. The logarithmic relaxation time (τ) describes the decay of postseismic displacement, and t is the observation time of each GPS record since the initiation of the postseismic period. For each station, we vary both $D_{x,y}$ and τ until obtaining a minimum model misfit of each component in the time series. Fig. S2a shows the combined time series of the east and north components and the corresponding best-fitting logarithmic functions. However, this LP1 referenced framework cannot well characterize postseismic displacement as LP1 is close to the epicenter and has significant postseismic displacement. As a result, we choose a stable far-field station MOCH east of the Calaveras fault as the reference station for postseismic displacement.

We refer to 1989-1994 as “early” and 1994 onwards as “late” periods of the Loma Prieta postseismic deformation in part because of the apparent change of dominating mechanism [Segall *et al.*, 2000]. We plot the estimated early-to-late period postseismic displacements of each GPS station in Fig. 1b based on their best-fitting $D_{x,y}$ and τ values. The red, orange, and blue arrows are total postseismic displacements in 1989-1994, 1994-2000, and 2000-2013, respectively. We separate three time periods in order to highlight the afterslip dominated period (1989-1994; Segall *et al.*, 2000) and the decay of postseismic relaxation since 1994. Similar to the early postseismic period, the more recent displacements also show NE-SW convergence with a strike-slip component, but the amplitude is about three times lower than in the early period. Across the southern Santa Cruz Mountains, southwest of the Foothills thrust belt, all GPS measurements show a convergent motion with a right-lateral strike-slip component. The velocities of stations around the Foothills thrust belt are generally less than 3 mm/yr since 1994. Since 2003,

GPS and CGPS measurements (red arrows in Fig. 1a) show displacements very similar to the pre-Loma Prieta period (white arrows in Fig. 1a).

3.3 1992 – 2010 InSAR measurements

InSAR observations complement the GPS measurements and are particularly sensitive to vertical motions. Figure 3 shows the InSAR Line-of-sight (LOS) velocity in the Loma Prieta area relative to a reference pixel at the continuous GPS station LUTZ during 1992 – 2000 (Fig. 3a) and 2000 – 2010 (Fig. 3b), respectively. Note the InSAR reference point is different from GPS because point scatterers east of the Calaveras fault are less reliable due to higher vegetation density (Fig. S4). LUTZ is located on bedrock and is less sensitive to hydrologic related seasonal surface deformation [Chaussard *et al.*, 2014]. We remove the secular motion contribution using the same Bürgmann [1997] interseismic model in order to highlight postseismic deformation. We also include the vertical secular motion contribution because the LOS direction is a projection of 3D motions into the 1D LOS direction. The LOS velocity map spanning a larger region can be found in Fig. S4, and positive LOS-change values represent range decrease. The full-size interferogram (Fig. S4) shows substantial uplift and subsidence in the Santa Clara Valley that is associated with seasonal and longer-term groundwater-level changes in the underlying aquifer [Schmidt and Bürgmann, 2003; Chaussard *et al.*, 2014]. Supplementary Information S2 provides more detail about the surface deformation in this area, and Supplementary Information S3 describes how seasonal variation in the InSAR time series is estimated and removed from the time series.

South of the Santa Clara Valley near Almaden and Morgan Hill (Fig. 3), there is a change of LOS velocity between the two periods. In 1992 – 2000 (Fig. 3a), a uniform

InSAR range increase (blue) is obtained in the hilly area, consistent with surface subsidence of up to ~ 2 mm/yr. In the 2000 – 2010 period, this deformation pattern is not found (Fig. 3b). Time series of three points taken in the Almaden – Morgan Hill area (Fig. 4) further illustrate the change of LOS displacement with time. Generally there is a decay of LOS rates to insignificant motions from 1992 to 2000 (Fig. 4a-b), but in the south (Fig. 4c) no significant changes in rate can be resolved. Even though there are no well-correlated InSAR points closer to Loma Prieta, we infer the apparent subsidence south of the Santa Clara Valley in 1992 – 2000 to be associated with the Loma Prieta postseismic deformation. The maximum subsidence rate in this region is ~ 2 mm/yr in 1992 – 2000, and the subsiding area is consistent with the pattern seen in independently processed data in *Bürgmann et al.* [2006] (Fig. S1c). In 2000 – 2010, the mean LOS velocity in the foothills of the southern Santa Cruz Mountains is generally between ± 1.5 mm/yr. Since the seasonal variation in deformation rates in the Santa Clara Valley is similar to measured groundwater aquifer levels, we disregard this area for postseismic modeling in order to exclude hydrological deformation (outside of Fig. 3).

4. Postseismic deformation modeling

We first test the viscoelastic relaxation due to the Loma Prieta coseismic stress change in the lower crust and upper mantle during 1994 – 2014. Subsequently, we use dislocation models in a layered elastic half-space to calculate afterslip in 1989-1994, 1994-2000, and 2000-present time periods, by inverting the residuals from the viscoelastic model predictions. We invert for distributed afterslip on two fault planes

associated with the Loma Prieta fault rupture zone and the Foothills thrust belt, respectively.

4.1 Viscoelastic relaxation (VER)

We use simple elastic dislocation models to compute coseismic stress changes that drive postseismic relaxation in a layered viscoelastic representation of the Earth's lithosphere. By specifying the coseismic fault geometry and slip, and the depth dependent elastic and viscous parameters, we can predict the surface displacement in time due to the stress relaxation. We rely on the coseismic fault models of *Marshall et al.* [1991] and *Arnadottir and Segall* [1994] to develop a two-dislocation model (Table 1). We use this two sub-fault system to represent a rake transition in slip from nearly right-lateral (163°) in the southeast to oblique right-reverse in the northwest (116°). We follow the coseismic slip model setup in *Pollitz et al.* [1998] and set the coseismic slip to extend from 9.5 to 17.4 km in depth.

The layered rheologic model is composed of an elastic upper crust underlain by a viscoelastic lower crust and a viscoelastic upper mantle below 30 km depth (Fig. 5a), both with a Maxwell fluid rheology. The thickness of the upper and lower crust is 19 and 11 km, respectively, following *Pollitz et al.* [1998, 2004]. The VISCO1D code [*Pollitz, 1997*] is used to calculate deformation of the spherically stratified elastic-viscoelastic medium, relying on spheroidal and toroidal motion solutions separable in spherical harmonic degree. We allow for different viscosities of the lower crust and upper mantle, and consider variable viscosity of both layers (η_{lc} for lower crust and η_{um} for upper mantle) between 10^{16} and 10^{20} Pa s. We do not attempt to use the bi-viscous Burgers model (i.e., time dependent viscosity, see *Pollitz, 2003*) to describe the rapid early

postseismic deformation, which was previously found to be dominated by afterslip [Pollitz *et al.*, 1998]. It is difficult to separate the contributions of early afterslip and transient VER, so instead we compare linear viscoelastic models with postseismic measurements after 1994.

4.2 Afterslip

As described in Section 4.1, Segall *et al.* [2000] concluded that afterslip dominates the postseismic displacement until 1994 based on 1989.8-1998 GPS measurements. Here we perform afterslip dislocation inversions to the residuals of the VER models during 1989.8-1994, 1994-2000, and 2000-present, so we can evaluate the contribution of afterslip in the different time periods. We use a two-fault geometry based on the fault parameters found by non-linear inversion of postseismic GPS data in Bürgmann *et al.* [1997], which was also used in afterslip models of Pollitz *et al.*, [1998] and Segall *et al.*, [2000]. One fault roughly coincides with the Loma Prieta earthquake rupture (dipping 70° to the SW), and the other, shallower dipping (30°) fault lies in the Foothills thrust belt (Table 2). For this two-fault system each fault plane is composed of 20×10 subfaults, and the size of each subfault is roughly $3 \times 2 \text{ km}^2$. Each subfault is able to slip with a variable rake along the fault surface but no opening component is allowed, and the smoothing parameter determines the slip variation between sub-faults. We extend the main fault down dip to 25 km depth to allow for deep afterslip [Savage and Svarc, 2010] and evaluate the trade-off between deep afterslip and lower-crustal viscosity. We use the programs EDGRN/EDCMP [Wang *et al.*, 2003] for the calculation of the Green's functions relating unit slip on each sub-fault dislocation to surface displacement in a layered elastic model over a half-space, using elastic parameters from Pollitz *et al.*

[1998]. The smoothing parameters are determined by the method described in *Huang et al.* [2013].

5. Modeling results

5.1 Viscoelastic relaxation

To explore the rheologic structure, we perform a grid search for different lower-crustal and upper-mantle viscosities, allowed to vary between 10^{16} and 10^{20} Pa s, and compare the model surface deformation with GPS time series during 1994 – 2013 and InSAR time series during 1992 – 2010.

We formulate this problem in a Bayesian framework, and use a likelihood function $p(\mathbf{d}|\mathbf{m})$ to describe how well a model prediction given by specific parameters can explain the observed data [*Bodin et al.*, 2012]. The likelihood function for the GPS time series is,

$$p(\mathbf{d}_{\text{GPS}}|\mathbf{m}) = \frac{1}{(2\pi\sigma_{k,i,j}^2)^{\frac{2MN}{2}}} \times \exp \left[-\frac{1}{2}\Phi_{\text{GPS}}(\mathbf{m}) \right], \quad \text{and} \quad [2a]$$

$$\Phi_{\text{GPS}}(\mathbf{m}) = \sum_{k=1}^M \sum_{i=1}^N \sum_{j=1}^2 \left[\frac{O_{k,i,j} - m_{k,i,j}}{\sigma_{k,i,j}} \right]^2. \quad [2b]$$

Here $O_{k,i,j}$ is the j^{th} component of the i^{th} time step for the k^{th} GPS observation, and $m_{k,i,j}$ is the j^{th} component of the i^{th} time step for the k^{th} model prediction. $\sigma_{k,i,j}$ is the uncertainty of the j^{th} component of the i^{th} time step of the k^{th} GPS observation. Here we only consider horizontal measurements so $j = 1, 2$. There are totally 13 GPS stations regularly surveyed since the earthquake, so $k = 13$. The uncertainty of GPS is 3-6 mm for the *Segall et al.* [2000] dataset, and ~3 mm for the *USGS* dataset based on uncertainty estimations in their

studies. All of the GPS stations locate near the Loma Prieta earthquake region, so even stations east of the Calaveras Fault (e.g. HAMI, MOCH, and OSO1) experience some far-field VER. In addition, there may be displacement residual east of the Calaveras fault due to change of slip rate on the Calaveras fault after the Morgan Hill event (Fig. 2f). As a result, we allow for a systematic, common-mode shift to adjust the E-W and N-S components of all GPS stations to minimize the model residual, and we consider this result as a reference-point-free realization of the postseismic deformation.

The highest probability for GPS data favors a lower-crustal viscosity of $\sim 3 \times 10^{18}$ Pa s and upper-mantle viscosity of $\sim 10^{18}$ Pa s (Fig. 5b), but there is a wide high-probability region (yellow to red color in Fig. 5b) between 10^{17} and 10^{19} Pa s for the upper mantle and between 6×10^{17} and 10^{19} Pa s for the lower crust.

Likewise, the likelihood function of the InSAR time series is,

$$p(\mathbf{d}_{\text{InSAR}}|\mathbf{m}) = \frac{1}{(2\pi)^{MN/2}} |\mathbf{Q}_i|^{-1/2} \times \exp \left[-\frac{1}{2} \Phi_{\text{InSAR}}(\mathbf{m}) \right], \text{ and} \quad [3a]$$

$$\Phi_{\text{InSAR}}(\mathbf{m}) = \sum_{k=1}^M \sum_{i=1}^N [(\mathbf{o}_{k_i} - \mathbf{m}_{k_i})^T \mathbf{Q}_i^{-1} (\mathbf{o}_{k_i} - \mathbf{m}_{k_i})] \quad [3b]$$

Here $\mathbf{o}_{k_i}$ and $\mathbf{m}_{k_i}$ is the same as Eq. 2, and there is only one velocity component (LOS) so $j = 1$. $\mathbf{Q}_i$ is the variance-covariance matrix that contains the InSAR uncertainty structure. Calculating the InSAR uncertainty structure is non-trivial (e.g. *Bekaert et al.*, 2015) as the noise could be correlated in space due to atmospheric phase noise. In Supplementary Information S4 we provide the method suggested by *Sudhaus and Jónsson* [2009] to construct $\mathbf{Q}_i$ for each acquisition in Eq. 3. We down-sample the InSAR data into about one point per km^2 and only consider points near the Loma Prieta region (area shown in

Fig. 3). We also exclude areas that experience substantial seasonal variation associated with hydrology (Fig. S5a), in particular in the Santa Clara Valley (Fig. S1), and there are 638 points left after this process. As shown in Fig. 5c, VER with a lower crustal viscosity (η_{lc}) higher than $\sim 5 \times 10^{17}$ Pa s and upper mantle viscosity (η_{um}) $\sim 3 \times 10^{19}$ Pa s can explain the InSAR time series well. The best fitting relaxation model matches the longer-term trend in the time series (Fig. 4).

The joint likelihood function [Bodin *et al.*, 2012] of the GPS and InSAR datasets is,

$$p(\mathbf{d}_{\text{joint}}|\mathbf{m}) = p(\mathbf{d}_{\text{GPS}}|\mathbf{m}) \times p(\mathbf{d}_{\text{InSAR}}|\mathbf{m}) \propto \exp \left\{ -\frac{1}{2} [\Phi_{\text{GPS}}(\mathbf{m}) + \Phi_{\text{InSAR}}(\mathbf{m})] \right\}, \quad [4]$$

where the joint probability is the product of the likelihood functions for GPS and InSAR datasets. In this way, the relative contribution of the two datasets to the joint likelihood distribution is directly determined by data errors, so no user-defined weighting factor is needed in the inversion. Figure 5d shows the joint probability distribution. The result shows a similar distribution as for the InSAR data, but the region with high probability is more confined. Figure S7 shows the joint probability distribution with different InSAR weighting. The highest probability of upper mantle viscosity decreases from $\sim 2 \times 10^{19}$ to $\sim 3 \times 10^{18}$ Pa s when decreasing InSAR weighting, while the preferred lower-crustal viscosity remains $\sim 6 \times 10^{18}$ Pa s. Figure. 6a-c compares the best-fitting VER model with the average GPS velocities in 1989-1994, 1994-2000, and 2000-present and Fig. 6d and Fig. S2 show fits to the time series of total horizontal displacements for a range of lower-crustal and upper-mantle viscosities considered.

Alternatively, we consider GPS data from *Bürgmann et al.* [1997], *d'Alessio et al.* [2005], and continuous and campaign USGS data (<http://earthquake.usgs.gov/monitoring/gps/>) to represent the 1989-1994, 1994-2003, and

2000-present time periods, respectively. We perform the same approach and the results are shown in Fig. S6. As we only consider post-1994 GPS data when determining the model misfit, the best fitting model strongly underestimates the observed 1989-1994 displacement, which we attribute to early afterslip. After 2000, the modeled postseismic velocities generally drop to below 2 mm/yr. The higher VER misfit in 1989-1994 implies additional processes to explain the early stage of postseismic displacement. In addition, we take the best-fitting VER model to predict station-pair length change (green lines in Fig. 2). We do not include baseline-length change data in model fitting because the post-Loma Prieta measurements are GPS based so already included in the viscoelastic modeling. The viscoelastic relaxation prediction agrees well with length change measurements except for ALLI – LOMA (Fig. 2d). The opposite length change is because both ALLI and LOMA move southeastward and the lengthening is due to further southwestward motion at LOMA. In the predicted relaxation, LOMA has slightly northward movement whereas southward at ALLI, and hence predicting shortening.

5.2 Afterslip inversion

For afterslip inversions, we extend the Loma Prieta fault to 25 km depth to allow for the contribution of afterslip in the lower-crust down-dip of the coseismic rupture [Savage and Svarc, 2010] and examine the tradeoff between lower-crustal viscosity and deep-seated afterslip. In order to estimate the afterslip component, we use the VER model residuals of GPS and InSAR datasets during the 1989-1994, 1994-2000, and 2000-present periods as dislocation inversion inputs. The upper-mantle viscosity is fixed at 2×10^{19} Pa s, whereas the lower-crustal viscosity varies from 5×10^{16} to 10^{20} Pa s. We

calculate the reduced χ^2 misfit of each afterslip-viscoelastic relaxation model to evaluate model fitting. The χ^2 misfit is defined:

$$\chi^2 = \frac{1}{MNP} \sum_{k=1}^M \sum_{i=1}^N \sum_{j=1}^P \frac{(o_{k,i,j} - m_{k,i,j})^2}{\sigma_{i,j,k}^2}, \quad [5]$$

where notations the same as Eq. 2b.

Figure S8 shows afterslip distributions and the (reduced) χ^2 misfit for a range of lower-crustal viscosities. The result suggests more deep afterslip appeared when lower-crustal viscosity is higher. Figure 7 shows afterslip distribution and fitting to GPS with the best fitting lower-crustal viscosity (6×10^{18} Pa s). In the 1989-1994 time period, the GPS χ^2 misfit is 16.4 when solely using the VER model (black arrows in Fig. 7a), and the χ^2 misfit is 3.0 and 1.4 in the 1994-2000 and 2000-present period, respectively. Overall the improvement in fit from the addition of afterslip is better near the Loma Prieta region than the far-field regions (PAWT, EAUN, and stations east of Calaveras Fault).

The χ^2 misfit of the afterslip inversion is 3.2 for 1989-1994, 1.04 for 1994-2000, and 0.5 for 2000-present. They are all smaller than the viscoelastic model chi-square misfits because it is taking VER residual as data input, so the $(o_k - m_k)^2$ term of afterslip models in Eq. 5 is generally lower. The dislocation inversions in Fig. 7d-f show shallow oblique right-lateral strike-slip on the Loma Prieta fault and dip-slip dominated afterslip in the northwestern portion of the Foothills thrust in all time periods. This strike-slip dominated Loma Prieta fault afterslip and dip-slip dominated Foothills thrust afterslip is similar to the pattern obtained by *Segall et al.* [2000]. Alternatively, we perform the same afterslip inversion to the VER model residuals of GPS data (Fig. S6) from *Bürgmann et al.* [1997], *d'Alessio et al.* [2005], and recent USGS solutions of campaign and

continuous GPS data to represent the 1989-1994, 1994-2003, and 2000-present time periods. The results for these larger datasets are shown in Fig. S8 and the χ^2 misfit are 3.9, 1.2, and 1.7 for the 1989-1994, 1994-2003, and 2000-present time period, respectively.

6. Discussion

6.1 Early to late period Loma Prieta postseismic deformation

Fitting logarithmic functions (Eq. 1) to the GPS time series, there is an azimuthal change in the postseismic displacement close to the Loma Prieta epicenter between the early and the late periods that might imply a change of source mechanism (Fig. 1b). Combining both VER and afterslip mechanisms we can explain the GPS data well in all time periods near the Loma Prieta region (Fig. 8). The VER model also predicts a range increase in InSAR LOS (Fig. 8b) in the 1992-2000 time period that agrees with observed LOS increase (Figs 3 and 4). These results are consistent with a change of dominant relaxation mechanism from afterslip [Bürgmann *et al.*, 1997; Pollitz *et al.*, 1998; Segall *et al.*, 2000] and/or fault zone collapse [Savage *et al.*, 1994; Savage and Svarc., 2010] in the early postseismic period to VER in the later period. The total moment of the 1989-1994 afterslip is equivalent of M_w 6.3, about 11.5% of the coseismic moment. We cannot rule out the contribution of afterslip during 1994-2000, or even after 2000 (Figs 7 and S8).

Postseismic deformation fitting in the later stage (2000 to present) remains challenging because the postseismic transient is increasingly difficult to separate from the interseismic secular motion. In the study area there were no long-term GPS measurements prior to the Loma Prieta earthquake, so we rely on the interseismic secular

motion model by *Bürgmann* [1997], which is inverted from pre-Loma Prieta earthquake trilateration surveys [*Lisowski et al.*, 1991]. As shown in Fig. 2f, the 1984 Morgan Hill earthquake influenced part of the interseismic model, and additionally we cannot rule out the possibility of un-modeled interseismic motion, which may contribute to uncertainties of our postseismic deformation time series. This may explain the poorer later-stage postseismic GPS fitting (Fig. 6). Likewise, since there is no pre-Loma Prieta InSAR data, we also rely on the same interseismic model to remove the interseismic LOS contribution throughout the entire InSAR time series, so the interseismic model error could again result in a bias for the later postseismic period.

6.2 Afterslip in different time periods

Savage and Savrc [2010] postulated that most of the postseismic displacement in the first 3 years can be attributed to a 1.56 m right-lateral and 0.6 m reverse afterslip on the downdip extension (depth range 16 to 21 km) of the Loma Prieta earthquake fault. In addition, they propose a 0.11 m postseismic fault zone collapse (fault-normal displacement) of the rupture zone (depth ranges 5 to 16 km). *Bürgmann et al.* [1997] combined 5 years of GPS and leveling measurements and suggested about 2.9 cm/yr uniform oblique-reverse afterslip on the Loma Prieta fault plane and 2.4 cm/yr uniform reverse afterslip on a buried fault within the Foothills thrust belt.

In this study, we do not allow opening (fault zone collapse) along the fault planes, so the dislocation prediction (white arrows in Fig. 7a) is solely due to afterslip on the fault planes. The amount of deep afterslip in the 1989-1994 period strongly depends on the choice of lower-crustal viscosity, and there is a clear tradeoff between lower-crustal viscosity and the amount of deep afterslip (Fig. S8). The inverted afterslip model can fit

the VER model residual when lower-crustal viscosity is higher than 5×10^{17} Pa s, so we cannot separate the afterslip or the lower-crustal relaxation contribution in the early postseismic period. Nevertheless, based on the best fitting model, afterslip dominates at the depth between 8 and 15 km during 1989-1994 (Fig. 7d) and falls to below 3 cm/yr afterwards (Fig. 7e & f). In the Foothills thrust, the peak afterslip based on the 13 GPS stations time series is about 80 mm/yr, but is inconsistent with the result using GPS data from *Bürgmann et al* [1997] (Fig. S9a). This is possibly due to different number of geodetic constraints. Besides, it is difficult to directly compare afterslip with previous studies (e.g. *Bürgmann et al.*, 1997; *Segall et al.*, 2000) because we invert the VER model residuals instead of postseismic deformation.

Figure 8 compares the predicted postseismic deformation from both afterslip and VER. At later time periods (1994-2000 and 2000-present) the relatively uncertain GPS velocities of PAWT and EAUN (2-5 mm/yr) NW of the Loma Prieta fault may map into afterslip in the NW section of the Foothills thrust, leading to small-scale model LOS change not observed in InSAR (Fig. 4).

6.3 Constraining lithospheric rheology

Recent studies (e.g. *Hearn et al.*, 2002; *Johnson et al.*, 2009; *Barbot et al.*, 2010; *Bruhat et al.*, 2011; *Rousset et al.*, 2012; *Pollitz*, 2015; *Rollins et al.*, 2015) have explored the afterslip-and-viscoelastic-relaxation coupling model that could potentially separate the contributions of the two mechanisms by jointly inverting both mechanisms. In this study, since afterslip dominated the first 5 years of postseismic deformation [*Segall et al.*, 2000], we focus on using post-1994 measurements for exploring lower-crustal and upper mantle viscosities.

The best-fitting VER model suggests that the viscosities of the lower crust and upper mantle are 6×10^{18} Pa s and $3 \times 10^{18} - 2 \times 10^{19}$ Pa s, respectively (Fig. 5d). The viscosity of the upper mantle estimated in this study is consistent with *Pollitz et al.* [1998] ($\sim 10^{19}$ Pa s), *Kenner and Segall* [2003] ($3 \times 10^{19} - 4 \times 10^{19}$ Pa s), and *Pollitz et al.* [2004] (10^{19} Pa s), but higher than *Johnson and Fukuda* [2010] (4×10^{18} Pa s) for the San Francisco Bay Area. Other studies for upper mantle viscosity in southern California also infer a range of $6 \times 10^{18} - 3 \times 10^{19}$ Pa s [*Freed and Bürgmann*, 2004; *Behr and Hirth*, 2014]. A recent study by *Smith-Konter et al.* [2014] using tide gauge data to constrain interseismic vertical deformation also predicts asthenosphere viscosity of $\sim 10^{19}$ Pa s. For the lower crust, the viscosity estimated in this study is ~ 1.6 times lower than in *Pollitz et al.* [1998] ($\sim 10^{19}$ Pa s), and also lower than values obtained from a range of observations in the western United States ($10^{19} - 10^{21}$ Pa s) [*Thatcher and Pollitz*, 2008]. As described in Section 6.2, we may have underestimated the lower-crustal viscosity because we cannot separate VER and afterslip well when the lower-crustal viscosity is higher than 5×10^{17} Pa s. In addition, part of the early inferred deep afterslip might be due to a transient lower-crustal viscosity [*Pollitz*, 2003; *Freed et al.*, 2012; *Huang et al.*, 2014].

Our modeling approach could potentially bias the estimate of the early VER contribution. However, we only consider post-1994 geodetic data for VER fitting, in which the contribution of afterslip is likely to be negligible [*Segall et al.*, 2000]. The low quality of data collected in the early 90s does not allow for incorporating higher-order viscous flow or rate-state fault rheologies. The main challenge of constraining lithospheric rheology here stems from the limited constraints on the pre-earthquake measurements, lower precision of geodetic measurements in the early 1990s, and

relatively modest postseismic VER generated by a M_w 6.9 earthquake. We consolidate GPS measurements from *Segall et al.* [2000], *d'Alessio et al.* [2005], and the USGS campaign survey (1994 – 2013) and continuous BARD (1990s – present) and PBO (~2006 – present) stations, but some of the stations were established after the earthquake, whereas others were discontinued afterwards. On the other hand, the afterslip-dominated early postseismic deformation does not allow us to easily discern a likely early VER signal in the GPS time series, and therefore we are unable to investigate a time-dependent viscosity such as a bi-viscous Burgers rheology [*Pollitz*, 2003].

6.4 Accelerated aseismic slip along the San Andreas fault

Turner et al. [2013] document repeating earthquake activity along aseismically creeping sections of the San Andreas and the Sargent faults near San Juan Bautista, southeast of the Loma Prieta rupture zone. They estimated the aseismic slip rate of both faults based on the repeater activity, and found that the San Andreas creep rate fell back to the pre-Loma Prieta slip rate only about 10 years after the event, consistent with results from creepmeters in the area [*Bokelmann and Kovach*, 2003].

Near San Juan Bautista, our VER model shows a decrease of differential stressing rate from ~3 kPa/yr in 1990 to ~1.5 kPa/yr at 10 km depth in 1999 (Fig. S11). This VER decay is similar to the observed decay of slip rate along the San Andreas fault [*Turner et al.*, 2013], and hence provides a source that drives the enduring decay of the fault creep and the repeating earthquakes on the creeping segment of the San Andreas fault. Accelerated aseismic slip on the creeping San Andreas fault in the aftermath of large earthquakes on the adjoining locked sections may account for the apparent interseismic slip deficit along this section of the fault [*Ryder and Bürgmann*, 2008] and may play a

role in stress transfer across the central San Andreas fault [*Ben-Zion et al.*, 1993; *Lynch et al.*, 2003].

7. Conclusion

More than two decades of deformation measurements following the 1989 Loma Prieta earthquake document that the postseismic displacement rates fell from > 4 cm/yr to millimeter-per-year levels since 2000. Based on dislocation modeling, afterslips on the fault that ruptured in the earthquake and a buried fault within the Foothill thrust belt dominate the rapid early postseismic deformation. Our modeling using a 1D viscoelastic structure composed of an elastic upper crust and Maxwell viscoelastic lower crust and upper mantle, implies that the viscosities of the lower crust and upper mantle are about 6×10^{18} Pa s and $3 \times 10^{18} - 2 \times 10^{19}$ Pa s, respectively. A combined afterslip-viscoelastic model can better explain early and late Loma Prieta postseismic deformation, where the afterslip is reduced in the later period. Our geodetic measurements cannot tightly resolve lithospheric viscosity due to the low amount of the postseismic displacement from this M_w 6.9 earthquake and higher data uncertainty in the early 1990s. The enduringly accelerated activity of repeating earthquakes and fault creep on the San Andreas fault and the Sargent fault after the Loma Prieta earthquake is coherent in time with the modeled postseismic deformation. While early afterslip on the creeping SAF relieved static coseismic stress increases, the added loading due to postseismic shear below the creeping section appears to have lead to more enduring creep-rate increases. This suggests that viscoelastic relaxation could be the source to drive accelerated shallow slip on these creeping faults following the Loma Prieta earthquake.

567

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Figure 1. Geodetic horizontal measurements in the southern San Francisco Bay Area, shown in the blue box in the map. **(a)** Surface deformation before and after the Loma Prieta earthquake relative to station LUTZ (yellow triangle). The white arrows are the preseismic secular motion prediction inverted from EDM measurements [Bürgmann, 1997]; the blue arrows are the BAVU GPS measurements during 1993 - 2003 [d'Alessio *et al.*, 2005]; the red arrows are the USGS velocities from campaign and continuous GPS measurements since 2003. The white star indicates the Loma Prieta earthquake epicenter. **(b)** The Loma Prieta postseismic displacements during 1989 -1994, 1994 - 2000, and 2000 - present. The circles are the standard deviation of the logarithmic fitting [Eq. 1] misfit to each GPS station in each time period. The GPS time series is shown in Fig. S2a. Secular motions have been removed based on Bürgmann [1997], and the postseismic displacement is relative to MOCH. The black lines are the major fault lines, and the red lines are the coseismic fault geometry by Marshall *et al.* [1990].

Figure 2. Distance changes between selected stations before and after the Loma Prieta earthquake (see Fig. 1b for the location of stations). **(a)** Lengthening between stations EAUN and LOMA (or LP1) and after removing interseismic secular motion **(b)**. The secular motion in **(b)** is removed based on *Bürgmann et al.* [1997]. **(c-d)** ALLI-LOMA, and **(e-f)** HAML-LOMA. The green line in each sub-figure shows the predicted shortening due to viscoelastic relaxation based on the best-fitting model (Fig. 5a). The magenta points are distance change between BARD sites MONB-SODB (**c** and **d**) or MHCB – SODB (**e-f**), where MONB is close to ALLI, MHCB is close to HAML, and SODB is 10 km northwest of LOMA. Locations of all substitute stations are shown in Fig. 1a.

Figure 3. InSAR line of sight (LOS) velocity during (a) 1992-2000 and (b) 2000-2010. Positive LOS value is range decrease, which represents eastward and/or uplift motion and vice versa. The LOS secular motion is removed based on interseismic dislocation models [Bürgmann, 1997], and the velocity is relative to station LUTZ (black triangle). The black lines are the major fault lines, and the red lines are the coseismic fault geometry by Marshall *et al.* [1990]. The black squares are the nearby towns and the black squares are the locations of time series plots in Fig. 4.

Figure 4. Selected InSAR time series at points near Loma Prieta region (see Fig. 3 for locations). The circles are the LOS displacements. The colored lines represent different viscoelastic relaxation predictions, and the color index is also shown. The thick black lines show the best-fitting model.

Figure 5. Viscoelastic relaxation (VER) model. **(a)** Rheology structure based on the best-fitting model corresponds to the joint probability in **(c)**. **(b)** Probability density function (PDF) of model fitting to GPS time series and **(c)** InSAR time series. **(d)** Joint GPS and InSAR probability density function.

Figure 6. Viscoelastic relaxation (VER) model in (a) 1989 – 1994, (b) 1994 – 2000, and (c) 2000 - present time periods. The black arrows show GPS postseismic displacement after correcting for a common shift in E-W and N-S components to minimize model misfit (see text) in different time periods based on Eq. 1, and the white arrows are VER predictions due to stress change from coseismic slip on the fault planes (red rectangles). The background colors are vertical viscoelastic predictions. (d) Selected GPS stations (see Fig. S2b for fitting to all stations) in horizontal time series. The colored lines represent different viscoelastic model (the same color-code as Fig. 4).

Figure 7. Inverted afterslip distribution on each fault planes in (a) 1989 – 1994, (b) 1994 – 2000, and (c) 2000-present time periods. The fault geometries are based on *Bürgmann et al.* [1997]. The orientation of the black arrows indicate rake. (d-f) Afterslip predicted surface displacement from the two faults. The black arrows are viscoelastic model misfits (the residuals in Fig. 6) and white arrows are afterslip prediction.

Figure 8. Viscoelastic relaxation and afterslip combined model fitting to GPS data in (a) 1989 – 1994, (b) 1994 – 2000, and (c) 2000 - present time periods. The black arrows are GPS postseismic displacements in different time periods, and the white arrows are the multiple-mechanism models in different time periods. The background colors show the predicted surface relaxation prediction in LOS, and the black outlined circles are InSAR velocity in each time periods.

873 Table 1: Coseismic fault parameters (after *Marshall et al.*, 1991)

Fault	Length (km)	Width (km)	Strike (°)	Dip (°)	Rake (°)	Slip (m)	Depth (km)	Moment (N m)
Plane 1								
(NW)	17	9.1	128	60	116	2.1	9.5-17.4	1.62×10^{19}
Plane 2								
(SE)	17	9.1	128	60	163	1.0	9.5-17.4	7.7×10^{18}

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877 Table 2: Fault geometry for afterslip model (after *Bürgmann et al.*, 1997)

Fault	Length (km)	Bottom depth (km)	Top depth (km)	Strike (°)	Dip (°)	Longitude	Latitude*
Plane 1 (Loma							
Prieta)	53.82	15.57	1.48	130	70	-121.55	36.80
Plane 2 (shallower							
fault)	61.40	6.11	1.62	132	30	-121.59	36.86

878 * Coordinates are of the center of lower edge of the model fault plane.

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