Postseismic Strain Following the 1989 Loma Prieta Earthquake From GPS and Leveling Measurements

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Abstract

Postseismic deformation in the five years following the 1989 Loma Prieta earthquake has been measured with GPS and precise leveling. Post-earthquake velocities at distances greater than ~20 km from the coseismic rupture are not significantly different from those observed in the 20 years prior to the earthquake. However, velocities at stations within approximately 20 km of the rupture exceed pre-earthquake rates and exhibit unanticipated contraction normal to the strike of the San Andreas fault system. A combination of forward modeling and non-linear optimization suggests that the observed postseismic deformations were caused by aseismic oblique reverse slip averaging 2.9 cm/yr on the San Andreas fault and/or the Loma Prieta rupture zone, and 2.4 cm/yr reverse slip along a buried fault within the Foothills thrust belt. The best-fitting sources of postseismic deformation are all located at depths of less than 15 km . We find no evidence for accelerated flow or shear below the Loma Prieta rupture in the first five years following the earthquake. The inferred postseismic slip is likely to have been caused by the coseismic stress change updip of the 1989 rupture.

Introduction

Geodetic measurements subsequent to some large earthquakes show accelerated strain rates in the years to decades following the rupture [e.g., *Thatcher*, 1986; *Pollitz and Sacks*, 1992]. For example, following the 1906 San Francisco earthquake, strain rates in the Point Reyes and Point Arena triangulation networks north of San Francisco were about 2-3 times the rates measured in the 1970's [*Thatcher*, 1974]. Postseismic transients have also followed subduction-zone earthquakes in Japan [*Kasahara*, 1981; *Thatcher*, 1986].

Using simple models of postseismic relaxation, *Thatcher* [1974; 1983] concluded that existing deformation data can be explained by either the viscous relaxation of a ductile (asthenospheric) layer underlying an elastic (lithospheric) plate, or by the downward propagation of aseismic slip along a lower crustal extension of the fault zone. Many mechanical models of deep seated deformation transients following large dip-slip and strike-slip earthquakes have been proposed [*Bott and Dean*, 1973; *Nur and Mavko*, 1974; *Anderson*, 1975; *Budianski and Amazigo*, 1976;

Rundle and Jackson, 1977; *Savage and Prescott*, 1978; *Cohen*, 1979; *Lehner et al.*, 1981; *Li and Rice*, 1987; *Pollitz*, 1992], however, there is a lack of adequate data to test and differentiate between them.

Fault creep at shallow depths (0 to 3 km) has been observed after several strike-slip earthquakes along the San Andreas fault (SAF) system. Shallow creep typically decays exponentially within a few years of the earthquake [*Smith and Wyss*, 1968; *Bilham*, 1989]. Postseismic accelerated creep at seismogenic depths has been inferred from trilateration measurements adjacent to seismic ruptures on the Calaveras fault in 1979 and 1984 [*Oppenheimer et al.*, 1990]. *Shen et al.* [1994] observed postseismic deformation with a relaxation time of about 34 days following the 1992 Landers earthquake. They suggest that slip on and below the rupture surface, as well as on secondary faults, best explain the observed motions.

Observations of historic seismicity patterns indicate that earthquakes tend to propagate along fault zones, rupturing fault segments immediately adjacent to, and within several years of a previously active break. Examples involving strike-slip faults have been reported along the North Anatolian fault system [*Toksöz et al.*, 1979], in NE China preceding the 1975 Haicheng earthquake [*Scholz*, 1977], the Calaveras fault zone [*Oppenheimer et al.*, 1990], and the SAF zone [*Savage*, 1971; *Wood and Allen*, 1973]. In the 19th century, earthquakes along the SAF and the Hayward fault in the San Francisco Bay area occurred in clusters spread over several years [*Ellsworth*, 1990]. This suggests the existence of transient processes that occur with characteristic time scales of several years. Lacking geodetic data on the time-dependent deformation around previously ruptured fault segments, we can only speculate on the mechanics of postseismic deformation and the relation between strain transients and the clustering of earthquakes.

The M 7.1 Loma Prieta earthquake of October 17, 1989, occurred on a previously unrecognized fault plane along a restraining bend of the SAF zone (Figure 1). The earthquake ruptured a buried 35-km-long fault dipping ~70° to the SW from about 8km to 18km depth [*Dietz and Ellsworth*, 1990]. Strike-slip and reverse dip-slip of about 1.5 m are inferred from geodetic data [*Lisowski et al.*, 1990a; *Marshall et al.*, 1991; *Snay et al.*, 1991; *Williams et al.*, 1993]. The slip distribution is more complex in detail, with slip of up to 5 m occurring on two patches to the NW and the SE of the hypocenter [*Beroza*, 1991; *Hartzell et al.*, 1991; *Steidl et al.*, 1991; *Wald et al.*, 1991; *Arnadottir and Segall*, 1994; *Horton*, 1996].

Southwest-dipping reverse faults northeast of the SAF, including the Monte Vista, Berrocal, and Shannon faults, show evidence of Quaternary deformation [Figure 1; *Aydin and Page*, 1984; *McLaughlin*, 1990; *Bürgmann et al.*, 1994; *Hitchcock et al.*, 1994]. We refer to these faults collectively as the Foothills thrust belt. Broad zones of contractional surface deformation during the Loma Prieta earthquake follow the trend of these fault zones intermittently over a total distance of ~20 km suggesting triggered slip [*Haugerud and Ellen*, 1990]. Triggered coseismic shallow slip was also recognized along the Sargent and San Andreas faults [*Aydin et al.*, 1992]. This zone has experienced strongly increased earthquake activity since the Loma Prieta event [*Reasenberg and Simpson*, 1992]. Focal mechanisms along the Foothills thrust belt show slip on reverse faults sub-parallel to the SAF [*Bürgmann et al.*, 1996].

Measurements of surficial postseismic slip subsequent to the Loma Prieta earthquake find shallow displacements along the SAF to be generally less than ~1 cm in the first year after the earthquake [*Behr et al.*, 1990; *Langbein*, 1990; *Rymer*, 1990]. The cumulative slip of aftershocks in the years subsequent to the earthquake [~5 mm; *King et al.*, 1990] is too small to be detected geodetically. Behr et al. [1996] find that slip rates on the creeping section of the SAF near San Juan Bautista (Figure 1) accelerated from a pre-earthquake rate of 7-8 mm/yr to a rate of ~13 mm/yr since the earthquake. About 2-3 cm of excess slip occurred over the northernmost 15 km of the creeping section in the 3 years following the earthquake, probably down to a depth of 1-3 km [*Gwyther et al.*, 1992; *Behr et al.*, 1996]. Surficial creep rates on the southern Hayward and Calaveras faults decreased for about three years following the Loma Prieta earthquake [*Galehouse*, 1992, Lienkaemper, 1993, pers. commun.; *Lienkaemper et al.*, 1992].

In this study we present Global Positioning System (GPS) data collected between October 1989 and May 1994 from a number of geodetic networks (Figure 2). In this paper we refer to (1) the Black Mountain profile across the SAF 30 km NW of the Loma Prieta rupture [*Bürgmann et* *al.*, 1996]; (2) the Loma Prieta profile through the epicentral region [*Savage et al.*, 1994]; (3) the 5-station Loma Prieta monitor network [*Davis et al.*, 1989; *Savage et al.*, 1994]; and (4) the 22 station Santa Cruz Mountains network that was observed in March and April 1990 [*Snay et al.*, 1991; *Williams et al.*, 1993] and reoccupied in April and May 1994 (Figure 2, Table 1). The Loma Prieta monitor network was surveyed with GPS and geodolite for several years prior to the earthquake [*Davis et al.*, 1989; *Lisowski et al.*, 1990b; *Savage and Lisowski*, 1995]. We also make use of velocities of two Bay area VLBI stations, determined relative to stable North America, for the time periods 1984 through 1989 and 1989 through 1991. VLBI observations at these sites were terminated in 1991 [*Argus and Lyzenga*, 1994]. Finally, high precision vertical displacements are derived from a comparison of two occupations of a leveling line across the epicentral region of the Loma Prieta earthquake in March of 1990 [*Marshall et al.*, 1991] and November of 1992 [*Bürgmann et al.*, 1993]. Our objectives are to determine the amount, nature, and source of deformation following the Loma Prieta earthquake and to examine the consequences for the potential of future earthquakes on the peninsular SAF and other faults in the San Francisco Bay region.

GPS Data Collection, Processing, And Analysis

Data Collection

A total of 12 GPS surveys of the Black Mountain profile, 12 surveys of the Loma Prieta profile, and 19 surveys of the Bay area monitor network were carried out between October 1989 and May 1994 (Figure 3 and Tables 1 and 2). Nine of these surveys included stations in both the Black Mountain and Loma Prieta profiles (usually LP1_ and BEND), allowing the relative coordinates of stations in both subnetworks to be accurately determined. In March and April of 1990 the U.S. Geological Survey, the National Geodetic Survey, and Stanford University coordinated two weeks of GPS observations of both profiles, the monitor network, and 26 additional sites, referred to here as the Santa Cruz Mountains network, in order to constrain coseismic deformation during the 1989 earthquake [*Lisowski et al.*, 1990a; *Snay et al.*, 1991; *Williams et al.*, 1993]. In April and May of 1994 we reoccupied 22 stations within the Santa Cruz

Mountains network together with the profile and monitor stations to better resolve the spatial distribution of postseismic deformation.

Stations in the Loma Prieta profile and Bay area monitor network were commonly observed once during each survey, whereas stations in the Black Mountain profile and Santa Cruz Mountains network were usually occupied at least twice in each survey (Table 2). Each station occupation commonly consisted of 6-8 hours of data collection, however, sites that could be left without an operator collected up to 24 hours of data per day. Tables 1 and 2 summarize the GPS data analyzed here. Table 1 presents the station codes, benchmark inscriptions, and station coordinates. Table 2 gives the occupation dates, GPS receiver types used, the agency that collected the field data, the fiducial stations used in the analysis (or source of precise orbit), and the networks occupied. For each survey with multiple occupations, Table 2 lists the day-to-day repeatability, expressed as the weighted root-mean-square in the North, East, and Up directions.

Data Processing

Our GPS data analysis follows the methods described by *Davis et al.* [1989], and utilizes the Bernese GPS analysis software (Version 3.5) [*Beutler et al.*, 1987; *Rothacher et al.*, 1990]. All data are processed with a satellite elevation cut-off angle of 15°, because higher or lower cut-off angles appear to reduce the short-term repeatability of the station coordinates. All double-difference carrier-phase residuals are visually inspected after processing the data with an automatic cycle-slip fixing program to remove remaining cycle slips and outliers. We attempt to resolve phase cycle ambiguities to integers by first resolving 'wide-lane' ambiguities and then fixing the remaining linearly independent integer-cycle ambiguities using the ionosphere-free linear combination. This combination of the two L-band frequencies (L1 and L2) eliminates most first-order effects of dispersive ionospheric delays that affect longer baselines. Baselines shorter than 3 km that are not significantly affected by ionospheric variations are processed in the less noisy L1 frequency. Practically all integer-cycle ambiguities in the doubly differenced phase measurements are reduced to integer values in the local data. Atmospheric propagation delays are modeled with a zenith delay parameter at each station which is piecewise constant over ~3-hour periods.

Errors in the broadcast satellite ephemerides can be significantly reduced by monitoring the GPS satellites from a global network of tracking stations with well known coordinates [e.g., *Larson et al.*, 1991]. For data collected prior to December 1991 we compute improved GPS satellite orbits utilizing data from North American fiducial stations (usually Mojave, California, Westford, Massachusetts, and Richmond, Florida) that are constrained to ITRF-92 coordinates and velocities. In a few surveys (see Table 2) we add data from tracking stations at Penticton and Yellowknife in Canada, and from Kokee on Hawaii to compensate for the lack of data from the three principal fiducial stations. The tracking data are combined to solve for orbital parameters, solar radiation pressure, and `Y-bias' over three- to four-day arcs. We do not attempt to resolve integer phase ambiguities for baselines to or between tracking sites that are more than 600 km apart. For data collected from December 1991 through 1992 we utilize precise daily orbits computed by the Scripps GPS processing facility. Beginning in January 1993 we use precise IGS (International GPS Service) orbits [*Kouba*, 1995], and include data from the nearest IGS tracking sites (Quincy and DSSC10) in California. Final estimates of the local station coordinates, atmospheric zenith delay parameters, satellite orbit parameters (until December 1991), and synchronization errors, are determined by processing the data from the fiducial stations together with the ambiguity-fixed local data.

Data Precision and Accuracy

The precision of GPS measurements depends the satellite and receiver clocks, propagation delays, satellite orbit and constellation characteristics, length and time of observation, the geographic location of a network, receiver types and post-processing software, set-up procedures, and stability of the monument.

Figure 3a shows changes in the N and E components of the station coordinates in the Black Mountain profile relative to BEND. In Figure 3b the time series of the sites in the Loma Prieta profile and the monitor network are shown relative to station LP1_ on Loma Prieta. Four receiver station days are left out in the final computations as operators noted significant shifts in antenna positions during the surveys caused by bovine impact or tripod instability (occupations of HLMR in May 1992 and January 1993, ANTE in May 1992, and BAYS in January 1993). Between January and September 1993 our measurements indicate a horizontal offset of > 5 cm of ANTE (Figure 3a). We assume that this motion was due to a disturbance of this site located in the curb of a residential road and do not use that data in our final velocity solution. Station BAYS in the Black Mountain profile was disturbed by vehicle impact in the fall of 1990. Following this, a tie between BAYS and a new reference mark BAYR was measured in December 1990. In May and June 1991 the old benchmark was surveyed, and the position of the reference mark at the time of these campaigns was calculated from the established tie. The reference mark was also disturbed in late 1992 and a new site Z137 was established nearby. Figure 3a shows the time series before and after the first benchmark disturbance determined by subtracting the BAYS-BAYR tie from the later measurements. Because the tie was measured after the disturbance we can not measure the true offset. Therefore the total displacement rate for BAYS is determined by solving for a common velocity of BAYS before the first incident, BAYS and BAYR from February 1991 to May 1992, and Z137 since 1992.

Table 2 summarizes the day-to-day repeatabilities achieved in campaigns including two or more site occupations. Between October 19 and November 1, three occupations of the Loma Prieta profile were conducted. In the June 1991, May 1992, and September 1993 Black Mountain profile campaigns most baselines were measured 2-to-4 times to estimate the precision of the survey results. Furthermore the Black Mountain profile was measured within the same week with TI4100 and Trimble receivers in the January, April, and September campaigns of 1990. The Santa Cruz Mountains network sites were observed 2-14 times in 1990 and in 1994. We note apparent improvements in day-to-day repeatabilities from >10-hour observation times and in more recent surveys that benefitted from better satellite coverage, receiver hardware, and the use of precise IGS orbits.

Postseismic Displacement Field From GPS

Least-squares estimates of the relative station velocities are shown in Figure 3 as solid lines. A visual analysis of the time series along the Loma Prieta profile suggests a possible decay in the postseismic displacement rates for some stations. *Savage et al.* [1994] found that the decay was particularly evident in the SAF-normal component. Data collected in 1995 shows that postseismic deformation rates had decayed to near pre-1989 values by 1993 [*Bürgmann et al.*, 1995]. While there is evidence for time-dependent deformation in some baselines, to first order the data can be fit with constant rates of motion. Because the data is irregularly sampled in time and the geometry of the active structures is not known *a priori*, it would be difficult to simultaneously solve for both fault geometry and time-dependent deformation history. In this paper we focus on determining the geometry of the source of postseismic deformation. In a subsequent paper we will address the time dependence of strain.

We compute site velocities assuming linear motions over the 4.4-year time period. Coordinate estimates and covariance matrices for each observation file are combined to determine least-squares estimates of station velocities. As appears to be the case in many GPS studies, we find that the formal errors computed during the GPS processing underestimate the observed scatter. The observed short-term repeatability and scatter about the best fitting linear displacement model suggest that the Bernese formal standard deviations must be scaled by a factor of 6 with computed orbits, or a factor of 10 when using IGS orbits.

The station velocities relative to PAWT and their 95% confidence ellipses, using all data for all sites between October 1989 and May 1994, are shown in Figure 4. Note that the velocities of stations in the Santa Cruz Mountains network are based on only two surveys in March/April of 1990 and April/May of 1994. Figure 4 also shows the VLBI derived velocities for the Pacific plate and the Sierra Nevada - Great Valley (SNGV) block [*D. F. Argus and R. G. Gordon*, written commun. 1996]. A total of 41.3 ± 3 mm/yr relative motion between the SAF-bounding blocks occurs at this latitude (oriented N34.5°W ±2.5°)[*D. F. Argus and R. G. Gordon*, written commun. 1996]. The measured displacement-rates of sites along the Black Mountain profile are consistent with right-lateral shear across the SAF system. About 32 mm/yr of right-lateral motion is accommodated across the 100 km width of this profile. Velocities of sites along the Loma Prieta profile and in the Santa Cruz Mountains network are also consistent with broadly distributed right-lateral shear, except for the region between Loma Prieta and the Calaveras fault zone where sites move almost due south relative to PAWT. The sense of shear is left-lateral between MAZZ and METC. Between OSO1 and BRUS the network accommodates 34 mm/yr of right-lateral shear directed N32°W.

The scatter in the vertical component is several times greater than in the horizontal. Fig. 3c shows the time series for the relative elevation changes of the stations along the two profiles and the monitor network. We find that only few vertical displacement rates are formally significant at the 2-sigma level, and even these should be viewed with caution. We include the GPS-derived vertical motions in our analysis, however, because of their large uncertainties, they do not influence our results strongly.

Leveling Data And Analysis

The low vertical precision (1-3 cm) in the GPS data did not allow us to resolve significant vertical motions (Table 3). Vertical motions can help to resolve between competing models of postseismic deformation, as discussed below. For this reason, a leveling line across the epicentral area of the 1989 Loma Prieta earthquake (Line 4 of *Marshall et al.*, [1991], Figure 5) was resurveyed to measure postseismic elevation changes [*Bürgmann*, 1993; *Bürgmann et al.*, 1993].

The leveling line is part of a larger network that was last occupied in the spring of 1990 to determine coseismic elevation changes [*Marshall et al.*, 1991]. First order single-run leveling surveys were performed by the National Geodetic Survey (NGS) in February and March of 1990 and in November of 1992. The line extends from Capitola at the Pacific coast via Loma Prieta to Coyote in the central Santa Clara valley, roughly parallel to the Loma Prieta GPS profile. The leveling line spans much of the width of the zone of anomalous fault-normal displacements determined from GPS. The only pre-earthquake occupation of this line occurred in 1953 [*Marshall et al.*, 1991].

Corrections and Error Analysis

A detailed analysis of the data and associated errors shows that the measured displacements are significant and apparently of tectonic origin [*Bürgmann*, 1993]. The data are corrected (by up to 4 mm) for systematic errors due to level collimation, rod calibration, thermal expansion of the rod tapes, earth tides and associated gravitational effects, and atmospheric refraction by the NGS. The elevation changes between 1990 and 1992 are listed as the differences of the normal orthometric heights relative to the southwestern-most benchmark in Table 4 and are shown in Figure 6. Data from three apparently unstable benchmarks (Stations HS5204, HS5213, and HS 5235 in Table 4) were discarded in the following analysis.

Random error is expected to accumulate with the square root of distance and was estimated as $0.8\ \mathrm{mm}/\sqrt{\mathrm{km}}$ for the 1990 survey based on double-run sections [*Marshall et al.*, 1991]. The corresponding error of the 1992 survey is $0.65\ \mathrm{mm}/\sqrt{\mathrm{km}}$ based on 10 double-run sections. Random error estimates based on the observed misclosure of circuits that included Line 4 in 1990 are $2.5\ \mathrm{mm}/\sqrt{\mathrm{km}}$ [*Marshall et al.*, 1991], however, it is not possible to know how much Line 4 itself contributes to the misclosure.

Slope-dependent systematic errors may be caused by errors in the rod calibrations. We test for the accumulation of elevation-dependent errors by performing an unweighted least squares fit of geodetic tilt to topographic slope along individual sections [*Stein*, 1981]. The correlation is not significant at the 95% level [*Bürgmann*, 1993]. A rough correspondence between elevation change and topography is apparent in Figure 6 on the eastern flank of the southern Santa Cruz Mountains. Because this correlation is not observed on the remainder of the line, we suggest that this is due to thrust faulting along the Foothills thrust system (see below) rather than measurement error.

The five northeastern-most stations along Line 4 are located in unconsolidated Santa Clara Valley sediments. They may be affected by well-water withdrawal and seasonal changes in water level. A detailed analysis of historic leveling data in the area and well-water records shows that water induced subsidence is not a significant problem for these benchmarks [*Bürgmann*, 1993]. Assuming that the errors of the two campaigns are not correlated, the relative uncertainty in the elevation change is $\delta_i^2 = \alpha_1^2 + \alpha_2^2$, where α_1 and α_2 are the random errors in $\mathrm{mm}/\sqrt{\mathrm{km}}$ determined for the two surveys. The random errors thus accumulate with distance along the line at a rate of $\delta = 1\ \mathrm{mm}/\sqrt{\mathrm{km}}$.

The signal to noise ratio of the leveling data is $\left[\frac{d^T \Sigma^{-1} d}{N}\right]^{1/2} = 3.2$.

Here, d is the $N \times 1$ vector of the elevation changes, Σ is the data covariance matrix, and N = 46 are the total number of sections used. This value compares with the signal to noise ratio of 9.3 of the coseismic uplift along Line 4 that had very large elevation changes, but also much larger associated errors [*Marshall et al.*, 1991].

Pre-Loma Prieta Earthquake Displacement Field

The velocity field shown in Figure 4 includes interseismic as well as postseismic motions. To isolate the deformation associated purely with post-1989 processes, we first construct a model of the interseismic (pre-earthquake) velocity field and subtract it from the measured velocities in Figure 4. The interseismic model is based on pre-1989 geodetic observations. Trilateration measurements in the San Francisco Bay area have been carried out for two decades prior to the Loma Prieta earthquake [*Lisowski et al.*, 1991]. Rigid-body rotations of the network are constrained by GPS observations at trilateration stations (or by ties between nearby GPS and trilateration marks) at LP1_, EAGL, ALLI, HAMI, and BRUS (Figure 2). Except for the effects of the 1979 Coyote Lake and 1984 Morgan Hill earthquakes, the deformation rates appear to have been constant during that time period. Figure 7 shows the horizontal velocity field in the region based on 20 years of trilateration data and up to 5 years of GPS data from the five Loma Prieta monitor stations (solid arrows), relative to Loma Prieta. Unfilled arrows represent the velocities predicted by a dislocation model of Bay Area strike-slip faults discussed below. Grey bold lines indicate the surface projection of the model strike-slip faults.

The interseismic velocity field in the southern San Francisco Bay area accommodates 38 ± 5 mm/yr of right-lateral motion across a zone about 120 km wide, as described by Lisowski et al. [1991]. The velocity vectors are approximately parallel to the regional trend of the SAF system and do not appear to be influenced by the local left bend in the SAF trace through the Santa Cruz Mountains. The displacement field broadens significantly from San Juan Bautista towards the northwest.

Figure 8 shows a close-up view of the measured pre-earthquake (unfilled arrows relative to BMT RF) and postseismic (solid arrows relative to PAWT) motions. Only a few of the postseismic stations were surveyed before the Loma Prieta earthquake. At some of the stations for which both pre- and postseismic data exist, such as HAMI, AMER, and GILR, the postseismic velocities exhibit significant fault-normal convergence that is not observed in the preseismic velocities. In order to compare pre- and post-earthquake velocities at the remainder of the GPS stations we must interpolate the pre-earthquake velocity field between stations. To do so, we developed a dislocation model of the San Francisco Bay area [*Bürgmann et al.*, 1996]. Uniform-slip dislocations in an elastic, homogenous, and isotropic half-space represent the Bay area faults. In this model, interseismic shear about a locked strike-slip fault is approximated by slip on a buried fault plane below the seismic zone at a rate comparable to the average geologic slip-rate [*Savage and Burford*, 1970; *Lisowski et al.*, 1991]. Surface creep on the Hayward, Calaveras, and the central SAF is modeled by shallow fault elements.

We estimate fault slip-rates using a linear inversion method [*Du et al.*, 1992] that uses the trilateration and GPS data to estimate slip on 78 individual segments of the Bay area faults. All faults are assumed to be vertical with pure strike-slip offset. *A priori* estimates of slip-rate are taken from *Lienkamper et al.'s* [1992] compilation of geologically determined slip rates. Results of the inversion indicate: (1) Deep (> 15 km) slip on the peninsular segment of the SAF at about 15-20 mm/yr; (2) deep slip (> 11 km) on the Hayward fault at 15 ± 3 mm/yr; (3) deep slip (> 10 km) on the northern Calaveras fault at about 10-12 mm/yr and at up to 35 mm/yr on the southern Calaveras fault; and (4) shallow creep on the East Bay faults and the SAF south of San Juan Bautista at rates comparable to the *a priori* values. For the San Gregorio fault we compute 2 ± 1 mm/yr slip from 10 to 15 km depth. Below 15 km the model predicts insignificant left-lateral slip (0.7 ± 1 mm/yr). The deep slip rate on the southern Calaveras fault is clearly too high and may be related to unaccounted for effects of the Coyote Lake and Morgan Hill earthquakes. Note that our objective here is to simply develop an approximation of the preseismic velocity field which can be used to compute the expected motions of stations for which preseismic data are not available.

Figure 7 compares the measured and modeled station velocities relative to Loma Prieta. Formal uncertainties in the predicted velocities are computed from the uncertainty in the data. Because the inversion constrains the predicted velocities to be generated from slip in an elastic half-space, the uncertainties in the predicted velocities are actually less than the uncertainties in the observed velocities. We thus scale the uncertainties in the predicted velocity field, by roughly a factor of 3, so that the predicted motions are not better determined than the observations.

In general the model predicts the observed velocity field quite well. The only significant exception is along the southern Calaveras fault in the Hollister network. Stations just east of the Calaveras fault appear to move towards the fault at an oblique angle. It is possible that this is the result of the complex series of creeping and locked fault segments along the Calaveras fault [*Oppenheimer et al.*, 1990]. Alternatively the southern Calaveras or the southern Sargent faults may accommodate convergence by oblique slip [*Matsu'ura et al.*, 1986]. In any event, there is no postseismic data in the Hollister region, so this should not significantly effect our results. The model also underpredicts the velocities southwest of the SAF south of Loma Prieta by up to 3 mm/yr.

Postseismic Displacement Field Anomaly

Horizontal displacements from GPS

The dislocation model of preseismic deformation can now be used to compute the predicted interseismic velocities and their uncertainties at all sites in the postseismic network. Subtracting the pre-earthquake velocities from the observed postseismic velocities yields the residual postseismic velocity field. The residual postseismic velocities shown in Figure 9 represent the component of the measured signal that exceeds the secular deformation field. The velocity field in Figure 9 is shown relative to PAWT, which we argue has not been affected by the Loma Prieta earthquake. VLBI measurements at Ford Ord (near BRUS) show that the velocity of BRUS relative to stable North America did not change significantly following the 1989 earthquake [*Bürgmann*, 1993; *Argus and Lyzenga*, 1994; 1996]. The insignificant change in the velocity of BRUS relative to

PAWT suggests that the residual postseismic velocities are adequately tied to an external reference frame (Figure 9).

Along the Black Mountain profile 44 km NW of the 1989 epicenter, postseismic velocities E of the Hayward fault agree well with their pre-earthquake estimates (Figure 9). ALLI appears to have an anomalous southwesterly motion that is not observed at neighboring sites. The combined velocity determined for BAYS and Z137 is suspect due to the problems associated with this site described above. Sites west of the SAF appear to have undergone a small but consistent acceleration, which is also observed at EAGL. Of the Loma Prieta-profile sites east of the Calaveras fault, HAMI moved with a higher fault normal velocity following the earthquake. The two easternmost sites, MOCH and OSO1, have marginally significant westward-directed motions.

The most significant residual postseismic motions occurred in a region around and northeast of Loma Prieta (Figure 9). Significant velocity increases and reorientations occurred in the near field of the earthquake. Fault-parallel velocities slightly increased by up to 8 mm/yr to the SW of the SAF. This acceleration is apparent at least as far to the NW as EAGL. The most surprising feature of the postseismic velocity field is a zone of SAF-normal contraction at rates exceeding 10 mm/yr, centered NE of Loma Prieta and extending as far east as HAMI across the Calaveras fault. The motion of AMER, located about midway between the two profiles, indicates that the zone of contraction extends beyond the NW end of the coseismic fault rupture. Unfortunately, we have little constraint on the extent of these motions southeast of the rupture zone.

Vertical Displacements From Leveling

The observed elevation changes are shown in Figure 6 together with a schematic cross-section of active fault zones along the line [*McLaughlin*, 1990] and the first two years of aftershocks. The aftershocks shown in Figure 6 outline the dipping Loma Prieta rupture below 10 km, a vertical SAF above 10 km, and distributed seismicity NE of the SAF. Relative to Capitola there is uplift of about 2 cm SW of the SAF, whereas the last 15 km of the line subsided by up to 3.5 cm. The steepest displacement gradient occurs directly in the hanging wall of the Berrocal fault zone (see also Figure 1). For comparison, the coseismic elevation changes on this line [*Marshall et al.*, 1991] are also shown in Figure 6. Notice that stations W of the Zayante fault, which uplifted coseismically due to slip on the Loma Prieta fault, showed no vertical deformation between 1990 and 1992. The fact that the postseismic deformation is shorter wavelength than the coseismic uplift and the absence of deformation on the Santa Cruz coastal plain W of the Zayante fault, both indicate that the postseismic deformation is caused by shallow rather than deep sources. Note also that Loma Prieta, which subsided ~10 cm in the earthquake, uplifted ~1 cm postseismically.

The measured uplift exceeds by about an order of magnitude the average late Cenozoic uplift rate of the range [*Bürgmann et al.*, 1994]. Northeast of the SAF are a series of reverse structures dipping towards the SW that show evidence of Quaternary slip and whose late Cenozoic offsets produced much of the observed geologic uplift [*Bürgmann et al.*, 1994; *McLaughlin and Clark*, 1996]. Rapid geologic uplift of about 1 mm/yr can be deduced from apatite fission track ages, geomorphometric analyses, and offset Quaternary terraces and soil horizons [*Bürgmann et al.*, 1994; *Hitchcock et al.*, 1994]. Uplift rates inferred from our model of preseismic deformation are also ≤1 mm/yr. We conclude that the uplift determined from leveling is likely to have been dominated by transient postseismic processes.

Data Interpretation and Models

In summary, the postseismic displacement field is characterized (1) by a ~40 km wide zone of accelerated right-lateral shear, sub-parallel to the SAF, centered near Loma Prieta, and (2) by a zone of anomalous SAF-normal contraction across the foothills of the Santa Cruz Mountains and the adjacent Santa Clara valley, centered NE of Loma Prieta (Figure 9). The complex fault structure in the region does not allow an unambiguous identification of the source(s) of the postseismic deformation. We therefore evaluate a range of models based on the observed crustal deformation and additional geologic and geophysical evidence. We model the displacements with rectangular dislocations in an elastic half-space [e.g., *Okada*, 1985].

Previously Proposed Models

The postseismic displacement pattern following the Loma Prieta earthquake has also been interpreted by *Savage et al.* [1994] and by *Linker and Rice* [1996]. *Linker and Rice* [1996] model the displacement field resulting from the interaction of the earthquake rupture with a linear viscoelastic relaxation zone representing the deep aseismic portion of the fault. They also developed a non-linear model with rate- and state-dependent friction on the down-dip extension of the rupture. Their models, involving a relatively deep deformation source below the coseismic rupture, predict the fault-parallel velocities in the Loma Prieta profile quite well, but do not predict the observed fault-normal motions and slightly overpredict the rates in the Black Mountain profile.

Savage et al. [1994] propose oblique slip on a 5-km-wide down-dip extension of the Loma Prieta rupture plane (1.5 m right-lateral and 0.9 m reverse), and 0.1 m of fault-normal collapse from 5 to 16 km depth to explain the fault-normal motions observed along the Loma Prieta profile during the 3.3-year period of their observations. *Savage et al.* [1994] also estimate right-lateral slip on vertical dislocations along the SAF (0.5 m) and the Calaveras fault (0.4 m) below 5 km that represent the background displacement field. That is, instead of subtracting the preseismic displacement field from the postseismic measurements, they solve for the postseismic anomaly *and* the secular displacement rates. This model predicts the horizontal displacement measurements quite well, but significantly misfits the leveling data, as shown below.

Previously, we proposed a model involving postseismic slip on two faults that fits the first two years of GPS data [*Bürgmann and Segall*, 1991; *Bürgmann et al.*, 1996]. The best-fit model for the early GPS data included predominantly strike-slip on the Loma Prieta coseismic rupture and reverse slip on a shallow southwest-dipping fault northeast of the SAF [*Bürgmann et al.*, 1996]. Here we evaluate models based on the leveling data and 4.5 years of GPS data.

First-Order Models

We first develop a series of first-order models to illustrate the characteristic displacement fields for a number of possible sources. We then test previously proposed models of postseismic relaxation and develop a model that best satisfies geologic and geophysical constraints (Table 5).

Previously observed or predicted postseismic relaxation processes include accelerated shear below the coseismic rupture and aseismic slip within or adjacent to the rupture zone (Figure 10). Oblique slip on a 5-km-wide SW-dipping fault below the hypocenter (16.8 km) with equal amounts of strike slip and dip slip (as in the mainshock) produces westerly oriented displacements and broadly distributed uplift SW of the fault and southerly motions to the NE of the fault (Figure 10a). The zone of anomalous deformation is approximately 60 km wide. Oblique slip on the coseismic rupture from 8 to 16.8 km depth produces a similar displacement field but with larger displacements spread over a narrower region (Figure 10b). Figure 10c evaluates the effect of fault zone collapse. This model predicts an asymmetric zone of contraction, as well as significant subsidence SW of the fault and a small amount of uplift NE of the fault. A model of reverse slip on a shallow 45° SW-dipping fault from 3.5 to 8 km depth produces a ~40-km-wide zone of contraction and a narrow uplift zone centered above the thrust (Figure 10d). None of these simple models predict the observed pattern of the horizontal and vertical motions.

Kinematic Models From Nonlinear Inversions

We continue our analysis by formally inverting for model fault parameters. We use a constrained non-linear optimization algorithm described by *Arnadottir and Segall* [1994], which allows us to estimate the geometry (parameterized by length, depth, width, dip, strike, location) and the amounts of strike slip, dip slip, and opening of one or more faults that best fits the GPS and leveling data. Specifically, we seek models that minimize the weighted residual sum of squares (WRSS, Table 5). It is possible to apply bounds (such as constraints on the depth of faulting or range of permissible fault strikes) to find best-fitting sources representing, for example, deep shear or slip on the coseismic rupture. Table 5 summarizes the results from a series of models. For comparison, note that a null model with no postseismic deformation has a model misfit $\sqrt{\mathrm{WRSS}/(\mathrm{n}-\mathrm{p})} = 3.26$ (Table 5).

To test the possibility of postseismic slip down-dip of the earthquake, we constrain the strike (132°) and dip (70°) to be the same as the coseismic rupture, and force the depth to be greater than the mainshock hypocentral depth (16.8 km). Note that, if we loosen the depth constraint, the inversion favors shallower model faults. The model misfit $\sqrt{\mathrm{WRSS} / (\mathrm{n}-\mathrm{p})} = 3.12$ indicates that models of postseismic relaxation below the earthquake rupture fit the data very poorly. In particular the deep source fails to reproduce the fault-normal displacements and produces a poorer fit to the leveling data than the null model. The deep-slip model (see Figure 10a) predicts westerly directed motions SW of the SAF, rather than the observed fault-parallel motions. It also predicts a region of broad uplift SW of the fault and a very gentle gradient NE of Loma Prieta; both features not observed in the data.

Next, we evaluate models of continued slip on the coseismic rupture dipping 70° and striking 132° [*Lisowski et al.*, 1990a]. The best-fitting model fault in this class lies close to the Loma Prieta rupture and has the maximum permitted fault width (from 16.8 km to 5 km depth) and maximum length of 45 km. The slip is nearly pure strike slip at a rate of ~5 cm/yr. If we remove the depth constraint on the upper edge of the dislocation, the best-fitting model fault extends to 1 km depth and the misfit is reduced to 2.97 (not shown in Table 5). As seen in Figure 10b, models with slip in the coseismic rupture zone also fail to predict the general characteristics of the observed deformation. The best-fitting unconstrained single fault is implausibly deep and is not judged to be geologically reasonable (Table 5).

Savage et al. [1994] suggested that fault-zone compaction together with accelerated fault creep below the coseismic rupture may explain the observed motions. The displacement field from a collapse model shown in Figure 10c shows asymmetric contraction and subsidence to the SW of the SAF. Adding these displacements to those in Figure 10a results in horizontal displacements that resemble those we observe. We test two models that include fault-zone collapse. A model that involves slip as well as contraction across the coseismic rupture with the same dip, strike, and depth constraints as the previous rupture model produces an improved fit to the GPS data, but does not fit the leveling data well. A model resembling *Savage et al.'s* [1994] favored model of fault collapse and creep below the rupture plane fits the GPS data even better, but does not improve the fit of the leveling data (Figure 11). The favored rate of fault-zone collapse is 3.5 cm/yr, for a total of about 15 cm during the ~4-year observation period, on a 10-km-wide and 28-km-long model fault. About 20 cm/yr of oblique slip on a 5-km-wide down-dip extension of the coseismic rupture is also predicted.

Because of the known thrust faults in the Santa Cruz Mountains, the observed coseismic ground deformation along these faults, and the predominance of reverse mechanisms of aftershocks NE of the SAF, we have previously suggested that the observed contraction is due to reverse faulting [*Bürgmann et al.*, 1996]. Slip on a shallow thrust fault produces localized uplift and contraction across the fault (Figure 10d). Clearly, shallow thrust faulting alone can not predict the observed motions. We find that a two-fault model is required to fit the observed horizontal and vertical velocities.

A model that includes slip on two faults, reverse and strike-slip on a fault near the SAF zone and thrusting on a fault NE of the SAF, provides the best fit to the GPS and leveling data (Table 5 and Figures 11 and 12). One model fault dips at a comparable angle to the "Loma Prieta fault" and has subequal thrust and strike-slip components, but is shallower than the coseismic rupture (Figure 13). The second model fault, located NE of the SAF, is quite long, dips at the lower bound (30°) of permitted fault dips and accommodates 2.4 cm/yr of thrust faulting. The SE ends of the model faults are not well constrained by the network and extend further than the coseismic Loma Prieta rupture.

We find that this model predicts all the first-order patterns observed in the postseismic data including accelerated fault parallel motion SW of the SAF, as well as contraction and associated uplift NE of the fault (Figure 11). This model fits the horizontal data well, but somewhat underpredicts the fault-normal motions in the eastern Santa Clara Valley and at Mt. Hamilton (Figure 12). It also slightly underpredicts the fault-parallel velocities SW of the SAF. Figure 11 shows the observed and predicted elevation changes along the leveling line for the best fitting two-fault model. The model fits the data well from the Pacific coast to the western edge of Santa Clara Valley. However, it does not predict the continued subsidence pattern at the NE end of the leveling line.

Although this is the best fitting, geologically reasonable model we have found, the misfit of 2.5 indicates that the model does not fully explain the data within the reported errors. This may be caused by an underestimate of the errors in the data, or in the computed interseismic correction.

Alternatively, the misfit may indicate that our model insufficiently represents the physical processes and fault geometries responsible for the deformation. Material heterogeneities, multiple fault surfaces, complex slip distributions, and distributed inelastic deformation may contribute to the misfit.

Discussion of Model Results

We find that single source models do not adequately explain the observed postseismic deformations. Accelerated flow or fault slip below the coseismic rupture does not appear to be the main source of the observed deformation. Deep processes produce long-wavelength displacement fields, not the narrowly focused deformation we observe. We find that much of the deformation is best explained by relatively shallow sources above 15 km depth. Our preferred model involves contemporaneous oblique-slip and thrust faulting at high rates on two faults with similar strikes. The dip of the first fault is not tightly constrained and we can not clearly differentiate if slip occurs on the SAF or the Loma Prieta rupture. However, we find that fault slip extends to very shallow depths, suggesting that faulting took place up-dip of the Loma Prieta fault. The inferred slip on the SAF or Loma Prieta fault may represent creep above, or between, the two high-slip asperities that released most of the earthquake moment [*Beroza*, 1991; *Hartzell et al.*, 1991; *Wald et al.*, 1991; *Arnadottir and Segall*, 1994; *Horton*, 1996].

The Foothills thrust belt has been active in the Quaternary and experienced shallow triggered slip during the Loma Prieta earthquake [*Haugerud and Ellen*, 1990; *McLaughlin*, 1990]. Our model thrust fault appears to coincide with the Berrocal fault zone, but it is possible that postseismic slip is distributed between several faults in this complex thrust system. Focal mechanisms of $M \geq 2.5$ Loma Prieta aftershocks, located NE of the SAF near the Berrocal fault zone, indicate that these events occurred predominantly on reverse faults (Figure 13). The seismicity near the Loma Prieta rupture on the other hand is very complex with all types of fault plane solutions [*Oppenheimer*, 1990; *Beroza and Zoback*, 1993]. Note that the fault slip we model exceeds by several orders of magnitude the slip accumulated in aftershocks and must be dominantly aseismic.

Discussion

Stress redistribution accompanying the Loma Prieta mainshock would increase the loading on thrust faults NE and updip of the rupture [*Reasenberg and Simpson*, 1992]. Figure 13 shows a cross-section perpendicular to the SAF with the contoured coseismic stress changes on thrust faults NE of the SAF overlaid on a plot of the aftershocks and the model faults. Shear stress changes and the thrust-normal stress changes both affect the loading of N60°W striking, 45° dipping thrust planes through a Coulomb Failure Function [*Reasenberg and Simpson*, 1992] CFF = $\tau_s + \mu\sigma_n$, where τ_s is the shear traction in the dip direction, σ_n the fault-normal traction, and μ the coefficient of friction. A coefficient of friction of 0.6 has been chosen for the model in Figure 13. As the complex slip distribution during the earthquake is expected to be important in the induced stress field in the close vicinity of the earthquake rupture we used the heterogeneous fault slip model of Beroza [1991] to compute the stress changes. Coseismically induced stress changes of 1 to 2 MPa that enhance thrust loading occurred NE of the SAF. The stress changes on fault planes of somewhat different strikes and dips are comparable to those shown. Apparently the increased micro-seismicity and the aseismic thrusting inferred from geodetic data reflect deformation in response to the load increase.

An enigma associated with the Loma Prieta earthquake was the observation that the highest topography of the Santa Cruz Mountains around Loma Prieta subsided by ~10 cm during the earthquake [*Lisowski et al.*, 1990a]. Long term uplift rates in the Loma Prieta area deduced from apatite fission track dating and geomorphic analyses average about 1 mm/yr over the last 5 Ma [*Bürgmann et al.*, 1994]. While different fault-plane geometries could produce the long-term uplift pattern [*Schwartz et al.*, 1990], slip on the reverse faults at the base of the southern Santa Cruz Mountains at combined rates of 2-3 mm/yr would result in the topographic expression of the southern Santa Cruz Mountains [*Bürgmann et al.*, 1994]. Our measurements indicate that some of this uplift may occur concurrently with aseismic fault slip at depth on the Berrocal fault zone following large regional earthquakes.

Accelerated slip on faults in the San Francisco Bay region will change the loading of nearby segments of the SAF and the Hayward fault. A detailed analysis of coseismic static stress changes on Bay area faults shows that micro-seismicity rates increased, or decreased, in accordance with the computed stresses [*Reasenberg and Simpson*, 1992]. Reduced creep rates, observed along the southern Hayward and Calaveras faults following the earthquake, are evidence of induced left-lateral shear stress across these faults [*Galehouse*, 1992; *Lienkaemper et al.*, 1992]. *Bürgmann et al.* [1996] modeled the change in loading on NW striking fault planes due to the Loma Prieta earthquake and postseismic fault slip with the method outlined above, assuming that fault-parallel shear stress as well as fault normal stresses determine the fault response through a Coulomb failure criterion. Postseismic fault slip further increases the load on the SAF immediately to the NW of the Loma Prieta rupture, while both the coseismic and the postseismic static stress changes have a retarding effect on the Hayward fault zone and most of the Calaveras and San Gregorio-Hosgri fault zones. The yearly postseismic stress changes are about an order of magnitude less than those induced coseismically.

Similar postseismic adjustments occurred following the 1979 Coyote Lake and the 1984 Morgan Hill earthquakes along the Calaveras fault [*Oppenheimer et al.*, 1990]. While the coseismic stress perturbations on neighboring fault segments induced by these moderate earthquakes were negligible [*Du and Aydin*, 1993], postseismic fault creep adjacent to the coseismic ruptures may have caused the observed south-to-north propagation of earthquakes along the Calaveras fault [*Oppenheimer et al.*, 1990].

The stress changes caused by co- and postseismic fault slip only add to those accumulated by background loading of the Bay area faults. The January 15, 1993 Gilroy earthquake that occurred in a region of induced left-lateral shear, east of the southern termination of the Loma Prieta rupture is a timely reminder of this. The peninsular segment of the SAF, the updip extension of the Loma Prieta rupture (from about 8 km to the surface), and several thrust faults that parallel the SAF through most of the Bay region, appear to represent significant seismic hazards since the Loma

Prieta earthquake. Continued reverse faulting may also relieve normal traction along the southern Hayward fault and increase the likelihood of faulting there, despite a decrease in right-lateral stress.

In 1865 a M ~6.5 earthquake occurred in the southern Santa Cruz Mountains [*Ellsworth*, 1990]. *Tuttle and Sykes* [1992] noted that intensities for this event were greater in San Jose in comparison to the Loma Prieta earthquake, but less in Monterey. On this basis, *Tuttle and Sykes* [1992] suggested that the 1865 event was located NE of the SAF. In their reanalysis of nineteenth century triangulation data in the San Francisco Bay area, *Yu and Segall* [1996] found that station Loma Prieta was displaced NE, perpendicular to the strike of the SAF, between 1850 and 1880, presumably as a result of the 1865 earthquake. The observed fault-normal displacement is inconsistent with slip on the Loma Prieta rupture and not easily explained by slip on the SAF. The fault-normal displacement is, however, exactly what would be expected for slip on a southwest dipping thrust fault located NE of the SAF. Slip on the shallow dipping thrust fault shown in Figures 12 and 13, with moment equivalent to a M 6 3/4 quake, explains the 1850-1880 displacements quite well [*Yu and Segall*, 1996]. Taken together these observations suggest, but certainly do not prove, that the 1865 earthquake was caused by slip on a thrust fault in the Foothills thrust belt and that the same fault(s) slipped aseismically following the Loma Prieta earthquake. We note that an earthquake comparable to the 1865 event would constitute a significant hazard to the south San Francisco Bay region.

Conclusions

Leveling and GPS data collected during 5 years following the 1989 Loma Prieta earthquake show a significant acceleration of deformation within about 20 km of the Loma Prieta rupture. The most striking feature in the postseismic motions is a zone of rapid uplift and contraction northeast of the SAF. Model inversions of the postseismic data suggest that aseismic oblique-reverse slip on the SAF and/or the Loma Prieta rupture and thrusting along the Foothills thrust belt are the most likely sources of the observed deformation. The first few years of postseismic deformation are dominated by aseismic fault slip up-dip of the coseismic rupture, probably triggered by coseismic stress changes. We do not find conclusive evidence for accelerated shear below the rupture or fault zone collapse. Continued monitoring of the transient deformation will provide valuable information about mechanisms of time-dependant deformation following a large earthquake and its effects on neighboring faults.

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Figures

Figure 1. Fault map with shaded relief of the San Francisco Bay area showing major fault zones and localities described in text. Fault abbreviations: MVF, Monte Vista fault; ShF, Shannon fault; BeF, Berrocal fault; SgtFZ, Sargent fault zone; the Foothills thrust belt consists of the MVF, the BeF, and the ShF. Fault locations are from *Aydin and Page* [1984].

Figure 2. Post-Loma Prieta earthquake geodetic network in the southern San Francisco Bay area consisting of two fault crossing profiles, the Loma Prieta monitor network (station names underlined), the Santa Cruz Mountains network (open circles), and two VLBI stations which were observed until 1991.Tables 1 and 2 summarize station coordinates and occupation times.

Figure 3. (a) Time series of relative changes in the horizontal (N and E) baseline components along the Black Mountain profile with respect to station BEND. The slope of the line fitted to the data is the relative station velocity. (b) Time series of horizontal coordinates of Loma Prieta profile and Bay area monitor network stations relative to station LP1_. (c) Time series of vertical baseline component for Black Mountain profile stations relative to BEND and Loma Prieta profile and Bay area monitor network sites relative to LP1_.

Figure 4. Postseismic velocity field computed from GPS data collected between October 19, 1989 and May 31, 1994. Velocities and their 95-percent confidence ellipses are shown relative to station PAWT along the Black Mountain profile. Also shown are the plate velocities of the Pacific plate and the Sierra Nevada - Great Valley block determined from VLBI data [*Argus*, written commun. 1996]. Most of the relative right-lateral plate motion occurs within our network. A zone of > 1cm/yr contraction northeast of the San Andreas fault is the most apparent deviation from simple distributed right-lateral shear.

Figure 5. Schematic geologic map of southern San Francisco Bay area showing the location of benchmarks (triangles) of the leveling line. Sites on bedrock (shaded patterns) are not significantly influenced by ground-water level changes. Also shown are the location of the nearby GPS stations of the Loma Prieta profile.

Figure 6. Postseismic elevation changes along the leveling line from SW to NE. The error bars represent the error in elevation change relative to the southwestern end of the baseline near Capitola. Also shown is a structural cross-section across the SAF and the Foothills thrust belt (modified from *McLaughlin*, 1990) and Loma Prieta aftershocks. The plotted earthquakes are from a 15 km wide zone centered on the line of section and include all well-located events from one month after the Loma Prieta earthquake until December, 1992. Note that the steepest gradient in the elevation changes is located in the hanging wall of the Berrocal fault, which also appears to be associated with high aftershock activity. The coseismic elevation changes (from *Marshall et al.,* 1991) on this line are shown for comparison.

Figure 7. Comparison of pre-Loma Prieta earthquake site velocities with 95-percent confidence ellipses (solid vectors) and displacements computed from an elastic model (open vectors) derived by linear inversion of the preseismic data. All displacements are shown relative to Loma Prieta. Shaded bold lines indicate the surface projection of the dislocation elements of the interseismic model.

Figure 8. Close-up of postseismic (solid vectors relative to PAWT) and preseismic (open vectors relative to BMT RF) velocities in the epicentral region of the Loma Prieta earthquake. The contraction northeast of the SAF observed in the postseismic data is not apparent in the pre-earthquake data.

Figure 9. Residual velocities determined by subtracting the pre-earthquake model velocities from our observations. Note a slight acceleration of fault parallel motions SW of the SAF and pronounced contraction NE of the SAF.

Figure 10. Displacement fields computed from first-order models of proposed sources of postseismic deformation. The location of GPS and leveling stations and the shaded surface projections of the model faults are shown. (a) Oblique slip below the Loma Prieta hypocenter (16.8 km to 21.5 km depth on 70° SW-dipping fault). Unit slip produces up to 0.07 units of uplift SW of the fault. (b) Oblique slip on coseismic rupture (8 km to 16.8 km depth on 70° SW-dipping fault).

(c) collapse of coseismic rupture. (d) Reverse slip on shallow thrust fault (3.5 km to 8 km depth on 45° SW dipping fault).

Figure 11. Measured elevation changes (with error bars) and vertical displacements predicted from the two-fault model (black bold line) shown in Figure 12. The model predicts the overall pattern of uplift, but does not predict the continued drop off at the NE end of the leveling line. Also shown (dashed line) are the predicted displacements from a model involving collapse of the Loma Prieta rupture and oblique slip below.

Figure 12. Postseismic residual velocities (black vectors with 95-percent confidence ellipses) and displacements (gray vectors) computed from the best-fitting dislocation model involving slip on two faults. The two rectangles show the projections of the faults used to compute the modeled station displacements (See Table 5 for model parameters).

Figure 13. Contoured stress changes (in MPa) induced by the Loma Prieta earthquake on 45° SW dipping fault planes. The coseismic displacement distribution determined by Beroza (1991) is discretized into 272 uniform slip fault elements in an elastic, isotropic, and homogeneous halfspace. Both fault-normal stresses and dip-parallel shear stresses influence the faults through a Coulomb failure criterion with a coefficient of friction $\mu = 0.6$. Also shown are the two model faults from Figure 11. The first fault lies a few km above the Loma Prieta rupture, the second fault lies in a cluster of seismicity NE of the SAF. Focal mechanisms of the main shock and 1989-to-1994 aftershocks ($M \geq 2.5$; projected onto a vertical N40°E-striking plane) from a ~10-km-wide profile across the SAF show a predominance of thrust mechanisms NE of the fault. The focal mechanisms are from the Northern California Earthquake Data Center selected using standard quality criteria [*Oppenheimer*, 1990].

Tables

Table 1. Benchmarks used in postseismic GPS survey and WGS-84 coordinates

Table 2. List of Experiments

Table 3. Preseismic and postseismic station velocities relative to PAWT

Table 4. Postseismic elevation changes along Line 4 across Loma Prieta

Table 5. Fault parameters for geodetically determined models from non-linear inversion of GPS and leveling data.