Strong interseismic coupling, fault afterslip, and viscoelastic flow before and after the Oct. 9, 1995 Colima-Jalisco earthquake: Continuous GPS measurements from Colima, Mexico

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1. Introduction

The $M_w = 8.0$ October 9, 1995 Colima-Jalisco earthquake ruptured a $\sim 150$-km-long segment of the northernmost Middle America trench that had been seismically quiescent since 1932. Modeling of teleseismic waveforms [Mendoza and Hartzell, 1999] and GPS displacements at 11 sites near the rupture zone [Melbourne et al., 1997; Hutton et al., 2001] suggest that up to five meters of coseismic slip occurred in two principal patches, both above depths of 20 km (Figure 1). Geodetically-measured displacements during a 3.5-year period following the earthquake suggest that the earthquake triggered afterslip below depths of 20 km and along parts of the subduction interface southeast of the main rupture zone [Hutton et al., 2001], presumably representing localized shear along velocity-strengthening regions of the subduction interface.

2. Data and Analysis

The GPS data we use were collected by the Mexican government agency INEGI between April, 1993 and mid-2001 with an L1/L2, C/A code receiver. Due to logistical limitations, we procured only one 24-hour session per week, except for a 10-month period bracketing the Colima-Jalisco earthquake, for which we procured daily GPS sessions. Prior to 1/22/96, the GPS antenna was located on the roof of an older building in the city of Colima, after which it was moved to the roof of a modern nearby building. No geodetic site tie was made between the two sites. We thus link the two coordinate time series in a manner that optimizes their continuity.

We analyzed the GPS data using GIPSY, satellite orbits and clocks from the Jet Propulsion Laboratory, and a standard point-positioning strategy [Zumberge et al., 1997]. Daily free-network station coordinates were transformed to ITRF97 [Boucher et al., 1999], yielding 3D coordinate time series. Motion of the North American plate relative to ITRF97, derived from the velocities of 140 GPS stations in the plate interior for which daily data are analysed at the University of Wisconsin, was removed from the time series so as to recast site motion relative to North America (Figure 2). Daily white noise in the north, east, and vertical components (4 mm, 7 mm, and 11 mm) is comparable to noise reported for many other continuous GPS sites [Zumberge et al., 1997], as is the long-period noise. All velocity and displacement uncertainties quoted herein are 1σ and are estimated using a model
give a velocity of $10 \pm 2.5$ mm yr$^{-1}$ toward N46°E $\pm 12^\circ$, representing margin-normal shortening with respect to the plate interior. Assuming this represents elastic strain accumulating in response to unrelated plate slip along the Rivera-North America and Cocos-North America subduction interfaces, we predicted the velocity at COLI from a uniform elastic half-space model within which geometrically realistic Rivera and Cocos plate subduction interfaces are embedded and fully locked to depths of 25 km. Location-dependent convergence rates and directions along the subduction interfaces are specified by the 0.78 Ma-average Rivera-North America and Cocos-North America angular velocities of DeMets and Wilson [1997].

The elastic half-space model predicts motion at COLI of $9$ mm yr$^{-1}$ toward N38°E. Motion predicted by the heterogeneous FEM is nearly identical.

[9] Provided that the GPS monument at COLI was stable during the years before the earthquake and the plate kinematic model we employ is accurate, the good agreement between the observed and predicted interseismic velocities (Figure 2) implies that the shallow regions of the Rivera-North America subduction interface were fully locked prior to the Colima-Jalisco earthquake.

4.2. Coseismic Motion: October 9 1995

[10] The coseismic displacement, 132 mm toward S66°W (Figure 2), is determined from the difference in the site coordinates predicted for October 9, 1995 by the linear-change model for motion before the earthquake (see previous section) and a linear-logarithmic decay model for transient postseismic motion (see below). Relative to the coseismic displacement at COLI, coseismic motion at nearby GPS site AVAL (Figure 1) is $\sim15\%$ greater and points toward the southwest (Figure 2). As described in the next section, the difference in the coseismic displacements of the two sites agrees well with results reported by Hutton et al. [2001].

4.3. Transient Postseismic Motion

[11] Motion during the twenty month period following the earthquake (Figure 2) decayed rapidly and was dominated by SSW-directed displacement 44° counterclockwise (CCW) from the coseismic direction. The predicted poroelastic response at COLI is too small [Masterlark et al., 2001] to explain the decaying displacement. Similarly, the viscoelastic response predicted by a FEM in which the upper mantle and lower crust are loaded by the well-constrained distribution of coseismic displacement is too small and in the wrong direction (Figure 2). The poor fit persists for a wide range of viscosities for the lower crust and upper mantle ($1 \times 10^{17}$ Pa sec to $1 \times 10^{20}$ Pa sec), as well as for models with non-linear and hence time-dependent upper mantle viscosity [Pollitz et al., 2001]. We conclude that poroelastic and Maxwell viscoelastic responses to the earthquake cannot explain the majority of the transient deformation during this period.

[12] An alternative explanation is that friction along the subduction interface is governed by rate- and state-variable constitutive laws, in which case any afterslip along the subduction interface and hence displacements at the surface will decay logarithmically [e.g. Scholz, 1990]. To test this, we fit $u(t) = d + A * \ln(bt + 1)$ to the north and east displacement components $u(t)$ for times before mid-1997 while requiring a common decay constant $b$ and separate values of $d$ and $A$ for the north and east components. Optimizing the least-squares fit yields good fits to both components for times before mid-1997, but mis-predicts displacements after mid-1997 (Figure 2). We also tested a model that includes logarithmic decay and a linear displacement component, the later presumably resulting from a combination of linearly accumulating strain due to a relocked subduction interface and slowly decaying and thus nearly linear viscoelastic rebound. The least-squares fit of the combined model improves on the fit of the simpler logarithmic decay model at much more than the 99% confidence level. The observations prior to mid-1997 thus contain useful information about the
logarithmic decay parameters and a linear term. The latter has not been previously detected for this earthquake.

These results are consistent with observations and predictions described in Hutton et al. [2001]. The CCW rotation of the direction of postseismic motion at COLI (Figure 2) mirrors similar CCW rotations observed at AVAL and other GPS sites. Modeling by Hutton et al. [2001] suggests this rotation is best interpreted as evidence that most afterslip was focused southeast and downdip from the principal earthquake rupture zone. In addition, the cumulative postseismic displacement at COLI ten days after the earthquake is observed to be \( \sim 15\% \) of the coseismic displacement (Figure 2), consistent with a \( \sim 15\% - 25\% \) estimate predicted by Hutton et al. [2001] from extrapolation of their measured postseismic GPS displacements back to the time of the earthquake.


Since mid-1997, COLI has moved 2.8 ± 1.2 mm yr\(^{-1}\) toward N29°E ± 5° (Figure 2), nearly parallel to but \( \sim 75\% \) slower than the rate of margin-normal shortening before the earthquake. Although the decreased rate of surface displacement could repre-
sent a temporary or possibly permanent decrease in the degree of locking along the subduction interface, we argue below that motion since mid-1997 more likely represents linearly accumulating elastic strain due to a fully relocked subduction interface, offset by the effects of continued fault afterslip and viscous flow in the lower crust and upper mantle.

[15] If we assume that the velocity since mid-1997 will eventually return to the pre-earthquake velocity upon sufficient decay of afterslip and any viscous flow, the present velocity difference between the two (lower right panel, Figure 2) presumably represents the net surface response to continued fault afterslip and viscous flow. Afterslip predicted by the logarithmic-decay and linear-motion model (Figure 2) is in the wrong direction and too slow to account for the observed difference between the pre-seismic velocity and the velocity since mid-1997. Encouragingly, the viscoelastic response since mid-1997 predicted by the FEM points in the correct direction needed to resolve the remaining velocity difference (Figure 2). Although the predicted viscoelastic response is too large by roughly a factor of two, its magnitude can be reduced by a factor of two by increasing the assumed viscosity of the lower crust to a slightly higher value of $1 \times 10^{19}$ Pa sec. We did not fine tune the viscosity structure to optimize the fit because our simplified modeling approach ignores coupling between afterslip and viscous flow, and because the problem is generally underdetermined.

[16] Better estimates of the relative contributions of strain accumulation, afterslip, and viscous flow will require more data, simultaneous modeling of the coupled viscoelastic and afterslip responses, improved estimates of the viscosity structure at depth, and consideration of alternative models for viscous flow (e.g. Maxwell, power-law creep, anisotropic). We are presently modeling the displacements of ~20 additional GPS sites in this region toward this goal.

[17] Acknowledgments. This work is funded by NSF grant EAR-9804905 and is additionally supported by the University of Guadalajara, University of Wisconsin, and INEGI. We thank Geoff Blewitt and Ralph Glaus for developing procedures to utilize C/A code and L1/L2 phase measurements for precise point positioning, Kristine Larson and two anonymous reviewers for constructive comments, Wallis Hutton for assistance at the beginning of this project, and the JPL/IGS analysis center for precise satellite orbits.

References


