Transient fault slip in Guerrero, southern Mexico

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Abstract. The Guerrero region of southern Mexico has accumulated more than 5 m of relative plate motion since the last major earthquake. In early 1998, a continuous GPS site in Guerrero recorded a transient displacement. Modeling indicates that anomalous fault slip propagated from east to west along-strike of the subduction megathrust. Campaign GPS and leveling data corroborate the model. The moment release was equivalent to an $M_w \geq 6.5$ earthquake. No $M > 5$ earthquakes accompanied the event, indicating the frictional regime is velocity-strengthening at the location of slip.

Introduction

Subduction of the Cocos plate under southern Mexico has generated more than twenty $M > 7$ earthquakes this past century (Figure 1). Rapid 5–7 cm/yr convergence generates major earthquakes on the subduction megathrust at 30-100 year intervals [*Kostoglodov and Ponce*, 1994]. Earthquake rupture zones support the seismic rebound hypothesis that elastic strain energy accumulates on frictionally locked portions of the plate interface and releases in earthquakes. Currently, the segment with the largest deficit in seismic energy release, the "Guerrero gap", is also the nearest to Mexico City (population ~20 million). Since the most recent large ($M_s = 7.6$) earthquake in 1911, >5 m of plate convergence has generated only a few $M_s \sim 6$ events near the edges of this segment.

The next major interplate earthquake in Guerrero could have moment magnitude $M_w = 8.1 - 8.4$ [*Suárez et al.*, 1990], but estimates of seismogenic potential simplify fault behavior by assuming an area of frictional locking that would rupture in a single event. Aseismic frictional slip varies with temperature, rock composition, gouge, fluids, stress and the history of fault slip [*Marone*, 1998]. Hence, aseismic fault slip velocity may change as the fault system (particularly stress) evolves during the seismic cycle [*Lapusta et al.*, 2000].

Transient fault slip is recorded infrequently because of the dearth of near-field geodetic instrument arrays. Nevertheless, four different types of aseismic fault slip have been observed. These include i) afterslip following large earthquakes [e.g., *Hutton et al.*, 2001]; ii) changes in creep rate on creeping sections of the San Andreas [*Linde et al.*, 1996]; iii)

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preslip before great earthquakes [*Linde and Silver*, 1989]; and iv) "slip events" that occur without clear space-time relation to large earthquakes [*Dragert et al.*, 2001]. Afterslip is commonly observed because we know when and where to look for it, but preslip is infrequent in the geodetic record. Elastodynamic friction models predict both of these behaviors [*Lapusta et al.*, 2000]. Slow earthquakes (i.e., transient slip events) have received less scrutiny.

GPS Measurements at CAYA

In January of 1997, a continuous GPS receiver was installed in Cayaco, Guerrero (CAYA, Figure 2). No other GPS receivers operated continuously nearby, so CAYA data were analyzed in a regional GIPSY network solution [*Lichten and Border*, 1987] which included Monument Peak (MONP), Ensenada (CICE), and Pie Town (PIE1). CICE and MONP are relatively near each other (115 km), which aided in network ambiguity resolution. Orbits were defined in the ITRF97 reference frame [*Boucher et al.*, 1999]. Each solution in the time-series of CAYA coordinates (Figure 3) is a 24 hour average of the baseline relative to PIE1, with the North American no-net-rotation prediction removed.

Coordinate time series at CAYA were initially fit by straight lines (i.e., constant velocity) and lines superimposed by hyperbolic tangent functions. The L_2 norm of misfit (weighted by the inverse of coordinate variance) is 7.49 mm for a line and 6.35 mm with a hyperbolic tangent function superimposed. The corresponding improvement of the χ^2 parameter of misfit is significant at $\gg 99\%$ confidence and suggests a static displacement of 2 mm east, 26 mm south and 16 mm up occurred over a period of several months in early 1998.

We considered several possible mechanisms for the signal. These include i) monument instability or localized motion, ii) groundwater variations, iii) changes in reflection or absorption of the GPS microwave signal, iv) changes in antenna or cabling, v) changes in the GPS reference frame, and vi) earthquakes. None of these can plausibly generate a static displacement with time-dependence, direction and magnitude consistent with the observed signal. The antenna is fixed atop a 5-m-long, 15-cm-diameter steel pipe bolted to a seismic vault. The vault is anchored and cemented to an unweathered gneiss which has high strength and low permeability. Tilt of the pipe assembly relative to a bedrock fiducial point is verified periodically with a permanent plumb-bob attached to the base of the antenna and has not deviated more than 3 mm from its original position. Groundwater near Cayaco varies primarily with rainfall, but CAYA motions are uncorrelated with precipitation records. The antenna ground plane is well above the nearby vegetation and structures, and there have been no changes of equipment at CAYA. March 1, 1998 coincides with a reference frame change, but these data were analyzed with orbits unconstrained and all solutions were then transformed into

the ITRF97 reference frame. Campaign GPS data exhibit similar anomalous motions between April 1996 and October 1998 (Figure 2), indicating the displacement is tectonic. However, seismic displacements at CAYA (dashed line in Figure 3) total less than a few mm, and early 1998 is seismically quiet. Consequently, the likely signal source is an aseismic slip event on the subduction megathrust near CAYA.

Deformation Modeling

Displacements at a single site cannot uniquely define a dislocation model, but CAYA data may be combined with ancillary information to reduce the number of unknowns. In Guerrero, the location and geometry of the subduction megathrust is known from local seismicity and gravity modeling [*Kostoglodov et al.*, 1996] (Figure 4). We model the deformation source as anomalous slip on the subduction interface, using

$$\mathbf{x}(t) = \mathbf{x}(t_0) + \mathbf{V}t + \oint \mathbf{S}(\zeta, t)G(\mathbf{x}, \zeta)d\zeta, \qquad (1)$$

in which $\mathbf{x}(t)$ are site coordinates at time t, $\mathbf{V}$ is constant velocity at CAYA in the absence of anomalous slip, ζ denotes location on the fault surface, G is the deformation Green's function [*Okada*, 1992], and $\mathbf{S}$ is a functional describing the transient slip. In choosing a form for $\mathbf{S}$, we assumed that the source was rectangular with constant slip rate $\mathbf{U}$, length L, and width W=10 km (Figure 4). The center of the source ζ_0 was permitted to propagate along strike of the fault as $\zeta_0=((t - t_0)\mathcal{V}, y)$. The propagation velocity $\mathcal{V}$, strike-perpendicular distance to the center of slip y, L, magnitude of the slip rate vector $\|\mathbf{U}\|$ and azimuth of slip θ were all assumed to be time-invariant.

The model has twelve parameters, five of which (t_0, y, $\mathcal{V}$, L and θ) relate nonlinearly to displacement. We parameterize these with a directed grid search. We first define a 200×200 grid of a pair of nonlinear parameters (e.g., t_0 and y). At each grid point, we estimate $\mathbf{x}(t_0)$, $\mathbf{V}$, and $\|\mathbf{U}\|$ from linear, weighted least-squares inversion. After completing search of a parameter pair, the two parameters are fixed to values yielding the minimum error norm and another pair of parameters is searched. After all possible permutations of parameter pairs have been searched, the search sequence is repeated until convergence is achieved (usually less than eight iterations). The search region is adaptively scaled to twice the range of the 95% confidence region, as determined from F-test statistics of the χ^2 parameter. By this method we assess the best-fit and confidence regions of all five nonlinear parameters.

The solution space has several error minima, but only two fall within 95% confidence of the global minimum. Model parameters for both are given in Table 1, along with extremal values at 95% confidence, and displacements for the global minimum are shown in Figure 3. Slip azimuth at the global minimum is more consistent with expected interplate moment release: the NUVEL1A [*DeMets et al.*, 1994] relative plate motion vector is 213° azimuth, and pure dip-slip motion is ∼200°. Most model parameters are tightly constrained, but L, $\mathcal{V}$, and $\|\mathbf{U}\|$ are cross-correlated so relatively poorly defined. However the total anomalous slip $\|\mathbf{U}\|L/\mathcal{V}$=1.42 m is well determined. The range of viable models is narrow despite using data from a single site, partly because of limiting assumptions about the deformation source and partly because each of the parameters yields a different time-dependence for displacement.
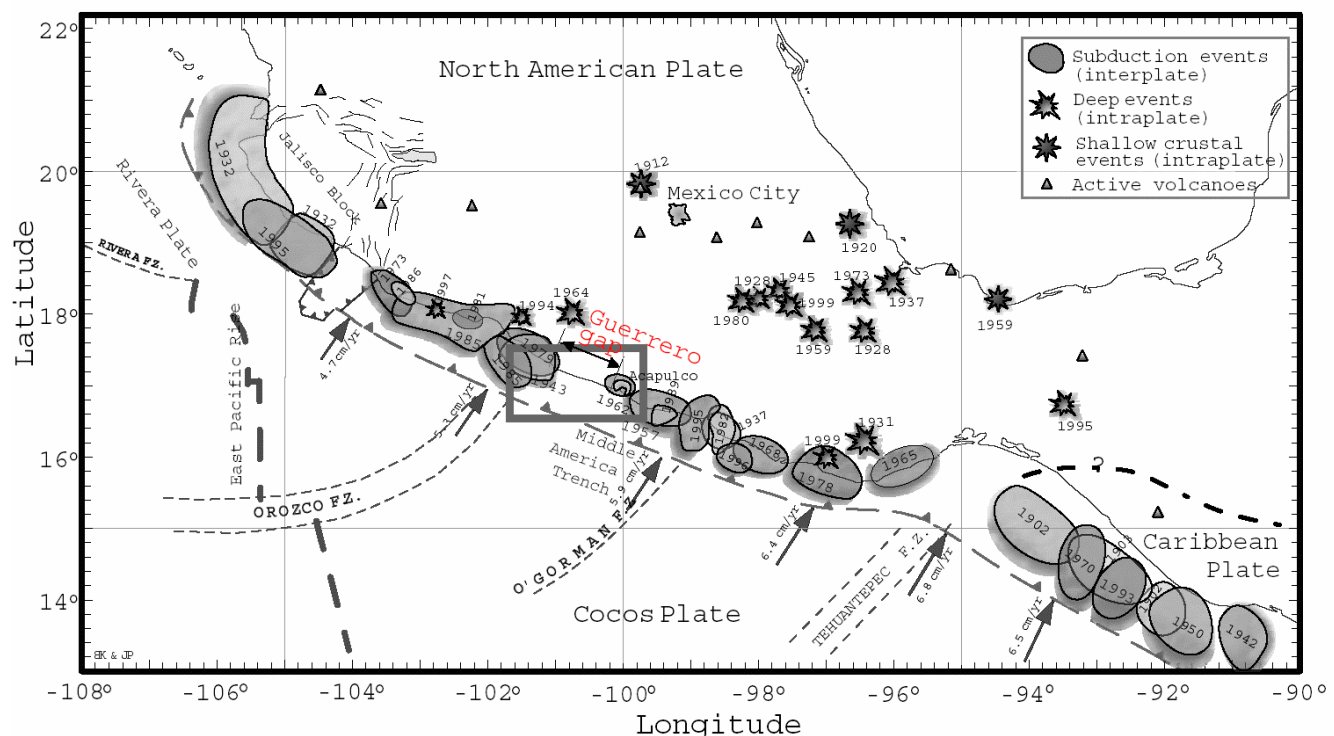

Figure 1. Seismotectonics of the Cocos-North America plate boundary. Gray box delimits the area in Figure 2.

Table 1. Best-fit model of displacement at station CAYA. Parameters of the global minimum are given first; parameters from the secondary minimum are in parentheses.

Parameter	Best-Fit Value	Units	95% Minimum	95% Maximum
t_0	0.223 (0.253)	Decimal 1998	0.158 (0.194)	0.290 (0.317)
$\mathcal{V}$	233 (299)	km/yr west	114 (178)	1641 (735)
y	-3.2 (27.0)	km seaward	-2.8 (37.9)	-3.7 (11.8)
L	117 (144)	km	38 (52)	699 (346)
$\|\mathbf{U}\|$	2.8 (1.6)	m/yr	2.0 (0.5)	6.7 (6.3)
θ	198.0 (160)	degrees azm	197.4 (150)	198.5 (176)
V_E	4.1 (4.1)	mm/yr east	3.9 (3.8)	4.3 (4.2)
V_N	7.2 (7.2)	mm/yr north	6.4 (6.6)	7.7 (7.8)
V_U	-14.7 (-15.0)	mm/yr up	-19.8 (-20.1)	-9.5 (-13.7)
$\|\mathbf{U}\|L/\mathcal{V}$	1.42 (0.78)	m	1.23 (0.23)	1.52 (2.36)
$\min\|\mathbf{x}_{\mathrm{obs}} - \mathbf{x}_{\mathrm{mod}}\|$	6.31 (6.32)	mm		

Discussion and Conclusions

Several lines of evidence support the modeling assumption that slip in the 1998 Guerrero event propagated along strike. Deflection of the site eastward by about 5 mm at the beginning of the transient, then westward at the end (Figure 3) cannot be duplicated on a static slip patch. This feature is significant: Models producing a purely static displacement in the east coordinate can be rejected at >95% confidence. The most plausible explanation for the east-west signal is tilt toward a propagating transient, first as it approaches and again as it leaves. The campaign GPS (e.g., Figure 2) and leveling data support this interpretation despite infrequent sampling. Anomalous slip had already begun near the GPS site in Acapulco by April 1996, was about half complete when leveling data were collected at Atoyac in March 1998, and the slip pulse stopped propagating near Papanoa shortly after the November 1998 GPS campaign. Other silent slip events on subduction megathrusts exhibit similar along-strike propagation of slip [*Ozawa et al.*, 2001; *Dragert et al.*, 2001].

The model also assumed a fixed fault width W=10 km. This was necessary because single-station data cannot con-

strain W. We inverted using larger W and found that the only parameter significantly affected was the slip rate $\|\mathbf{U}\|$. $\|\mathbf{U}\|$ does not trade linearly against W, as might otherwise be expected from the expression for moment release, $M_0=\mu\|\mathbf{U}\|L\mathcal{L}W/\mathcal{V}$, where $\mathcal{L}\sim$150 km is the total length of the slip anomaly. Equivalent moment magnitude estimates for the slip event range from M_w=6.5–6.8 for W=40–10 km, respectively.

At this time, we discern no clear relationship between the aseismic slip event and Guerrero earthquakes. Seismic-
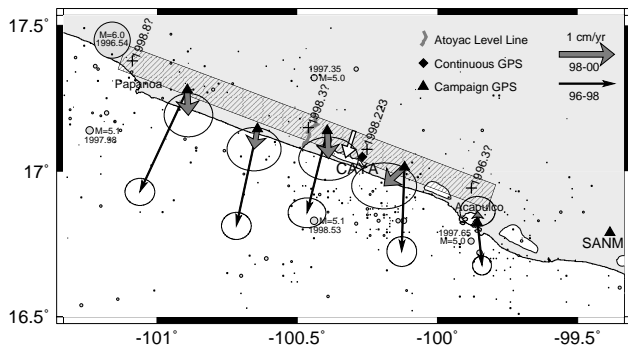

Figure 2. Location of GPS station CAYA. Surface projection of best-fit slip model is hachured; crosses are inferred timing, and white arrow shows slip vector azimuth. Earthquakes (circles) from 1996–2000 derive radius R from the model estimate of total slip (1.42 m) via $R^2=M_0/1.42\pi\mu$, where shear rigidity $\mu=2\times10^{10}$ Pa. 98–00 campaign GPS velocities relative to station SANM (thick gray arrows) are similar to 92–96 velocities (not shown), but differ significantly from 96–98 (thin black arrows). Error ellipses are 95% confidence.
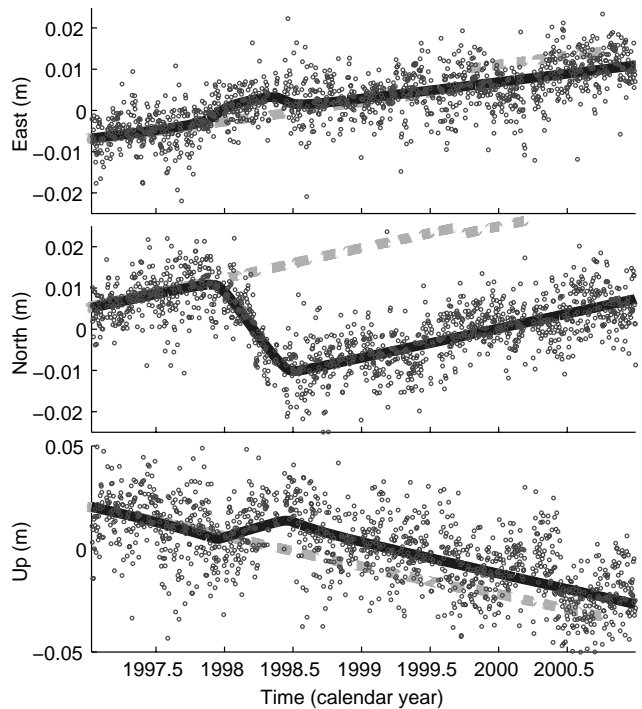

Figure 3. Time series of CAYA coordinates. Circles are daily estimates of position (weighted average standard errors are 1.2, 1.5 and 6.4 mm in the east, north, and up). Solid line is the best-fit model of deformation by slip on the subduction interface. Dashed gray line is best-fit background velocity $\mathbf{V}$ superimposed by displacement estimated for earthquakes in the Servicio Sismológico Nacional (SSN) catalog.
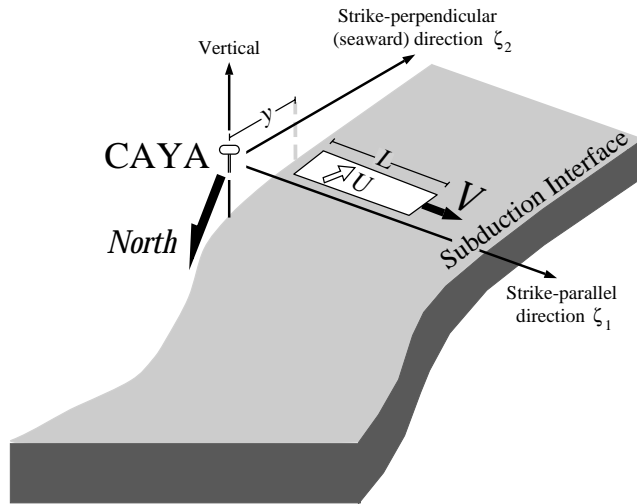
Figure 4. Parameters and geometry of the slip model.

ity is sparse in the slip region between 1996 and 2000 (Figure 2), and the enhanced seismicity up-dip (particularly between Acapulco and Atoyac) occurred months or more after the slip event. However we cannot rule out the possibility that the slip event was actually triggered by afterslip of the September 14, 1995 Copala earthquake (M_w=7.3; Figure 1), ~100 km east of Acapulco. Afterslip of this earthquake generated several centimeters of anomalous displacement at SANM and sites further east between October 1995 and April 1996. The data are not adequately sampled to determine whether slip in Guerrero is spatially and temporally connected to afterslip at Copala.

One of the goals of future research into fault slip transients will be to distinguish aseismic slip events from true earthquake precursors like those which apparently preceded the 1960 M_w=9.5 Chile and 1997 M_w=7.9 Kronotskoe earthquakes [*Linde and Silver*, 1989; *Gordeev et al.*, 2000]. The Guerrero event did not generate significant seismicity, suggesting that the portion of the megathrust which slipped has velocity-strengthening frictional properties. The slip event released ~2–5% of the estimated moment accumulation on the Guerrero segment, and one could argue that similar slip events occurring once every 2–5 years could accommodate all of the relative plate motion in Guerrero aseismically. However, the longer-term trend of velocities at CAYA and other GPS sites suggest strong coupling of the megathrust in a velocity-weakening zone up-dip. Hence it is more likely that the 1998 slip event acted to increase shear stress in the velocity-weakening region and thus increase the seismic hazard in Guerrero, as has been suggested also for the 1999 Cascadia event [*Dragert et al.*, 2001].

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