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GPS measurements across the northern Caribbean plate boundary zone: Impact of postseismic relaxation following historic earthquakes

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Abstract. GPS measurements in the northern Caribbean suggest that the rate of Caribbean plate motion relative to North America is about 10 mm/yr faster than predicted by global plate motion model NUVEL–1A. Several of the key sites used in the GPS study are located in the Dominican Republic, near the rupture zones of large earthquakes in 1946 and in the previous two centuries. Postseismic relaxation of the crust and upper mantle is a possible explanation for the plate velocity discrepancy. We explore a range of fault mechanisms and crustal and mantle rheology to place an upper bound on postseismic relaxation effects. The upper bound velocity contribution in the southern Dominican Republic is 5-6 mm/yr, and the most plausible contribution is 1-2 mm/yr, suggesting that postseismic effects cannot account for the discrepancy. This implies that the NUVEL–1A model underestimates the rate of motion of the Caribbean plate.

Introduction

Space-based geodetic techniques have found widespread application in recent years to directly measure long term ground motion in a variety of tectonic settings. Comparisons of predicted horizontal ground velocity from global plate model NUVEL–1A [*DeMets et al.*, 1994] and from VLBI, SLR, or GPS measurements have yielded excellent agreement in nearly all areas where comparisons have been made [e.g., *Robaudo and Harrison*, 1993; *Robins et al.*, 1993]. A decade of GPS measurements spanning the North America - Caribbean plate boundary zone, however, indicate that the Caribbean plate has a velocity that is about 10 mm/yr faster than the NUVEL–1A model prediction [*Dixon et al.*, 1998]. For example, the velocity of ROJO, southern Dominican Republic, moves at 21 ± 1 mm/yr to the east relative to TURK in the Turks and Caicos Islands on the stable North American plate (Figure 1). This differs at 95% confidence from the predicted NUVEL–1A value of 11 ± 3 mm/yr [*DeMets et al.*, 1994]. The purpose of this paper is to test one possible explanation for this disagreement.

The relative velocity between geodetic stations can change from one time period to the next. Apart from the obvious example of coseismic offsets, these temporal changes are often associated with major earthquakes preceding the observations, and

they are often explained by postseismic viscoelastic relaxation of the lower crust and upper mantle years to decades after the event. Postseismic relaxation from historic earthquakes in Hispaniola may explain part of the excess velocity because large historic earthquakes in 1751, 1842, and 1946 are documented [*Kelleher et al.*, 1973; *Sykes et al.*, 1982] and near-field relaxation effects in other tectonically active areas are known to persist for decades [*Thatcher et al.*, 1980; *Tabei*, 1989; *Rydelek and Sacks*, 1990; *Pollitz and Sacks*, 1992; *Pollitz and Sacks*, 1994] and suspected to continue up to 150 years or longer [*Rydelek and Pollitz*, 1994]. We shall systematically investigate postseismic relaxation effects following the historic earthquakes, considering a range of crust and upper mantle rheologies. Specifically, we seek to place an upper bound on the possible contribution of postseismic relaxation to the excess velocity field in the northeast Caribbean.

Fault Models

Representative fault planes of the 1751/1770, 1842, and 1946 earthquake sequences are shown in Figure 2. For simplicity we regard the three separate shocks which occurred from 1751 and 1770 as part of a continuous rupture over a 300 km long part of the Enriquillo fault zone (EFZ) in southern Hispaniola. The 1842 fault plane corresponds to a single event which ruptured a ~ 300 km long segment of the western Septentrional fault (SFZ) in northern Hispaniola [*Mann et al.*, 1984]. Based on geological observations we assume that the 1751/1770 and 1842 events involve purely left lateral strike-slip motion on vertical faults.

The 1946 sequence is dominated by the $M_s = 7.8$ August 4, 1946 event, followed by a few smaller events with magnitude up to $M_s = 7.3$ [*Russo and Villaseñor*, 1995; hereafter R&V95]. The distribution of aftershocks obtained by R&V95 suggests that the total area involved in significant rupture has a length of about 175 km and width of about 20 km. The fault geometry obtained by them (strike=N303° E, rake=74° on a northward steeply dipping fault) has a significant component of reverse slip. A different fault geometry for the August 4, 1946 event has been advocated by *Dolan and Wald* [1997; hereafter D&W97] and involves a combination of thrust and minor left lateral strike-slip motion on a similar trending fault with a shallow S/SW dip. D&W97 support their model by interpreting the focal mechanisms of shallow 1946 aftershocks and other nearby historic events as representing slip both updip and adjacent to the 1946 main shock. As discussed by these authors the difference between the two models is in the choice of nodal plane, the focal mechanism by itself yielding two equally plau-
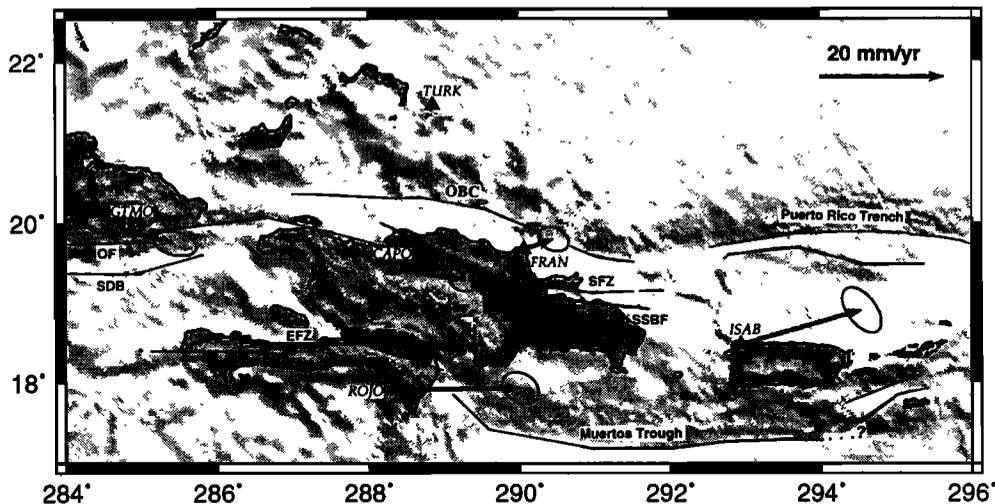
Figure 1. Velocities and 95% confidence ellipses of several GPS sites in the northern Caribbean relative to TURK, located in the Turks and Caicos Islands on the stable North American plate, for the period 1986-1995 from *Dixon et al.* [1998]. OF is Oriente Fault, SDB is Santiago Deformed Belt, SFZ is Septentrional Fault Zone, EFZ is Enriquillo Fault Zone, OBC is Old Bahama Channel, SSBF is South Samana Bay Fault.

sible choices. Assuming identical slip values and constraints on the depth extent of faulting, the choice of nodal plane is important for the predicted postseismic relaxation patterns because the area of the shallowly dipping plane is larger than the area of the steeply dipping plane. The predicted near-field relaxation patterns also depend heavily on the choice of fault geometry. Therefore we consider both fault geometries. For each geometry we parameterize the total 1946 moment release in terms of slip on a single fault plane trending parallel to the major faults in northern Hispaniola (Figure 1). We fix the strike to be N69° W (or S69° E) and either: 1. fixed the dip at that value determined by R&V95 (62° towards the NE) and fixed the rake at 45°, rather than the 74° determined by R&V95 for the August 4, 1946 event, to allow roughly equal components of strike-slip and thrust motion; or 2. fixed the dip at 12° towards the S/SW with rake of 55°, yielding a motion vector of the footwall of S55° W, somewhat more westerly than the motion vector of S34° W determined by D&W97.

Since precise constraints on the total moment of any of the events are not available, we fix the slip associated with each fault plane at 5 m. This is consistent with the fault lengths of the various earthquake ruptures and their large felt magnitudes. *Mann et al.* [1984] suggest a 300 year recurrence interval for great earthquakes in northern Hispaniola, in which case a maximum characteristic slip of 5 m is implied by the maximum possible long term relative plate velocity (20-25 mm/yr) combined with the upper bound of 75% total plate motion that either the SFZ or EFZ are likely to sustain [*Dixon et al.*, 1998]. The possibilities of aseismic interseismic slip and significant strain accumulation offshore north of the island also argue against larger slip values for the great inland earthquakes.

In order to determine the time-dependent deformation following an earthquake, we specify a three-layer elastic-viscoelastic coupling model as follows: 1. Purely elastic upper crust (0 – 16 km depth), bulk modulus $\kappa = 65$ GPa, shear modulus $\mu = 36$ GPa; 2. Viscoelastic lower crust (16 – 33 km), $\kappa = 95$ GPa, $\mu = 53$ GPa, viscosity $= \eta_c$; 3. Viscoelastic upper mantle (> 33 km), $\kappa = 150$ GPa, $\mu = 70$ GPa, viscosity $= \eta_m$. This simple crustal structure appears appropriate for several of the
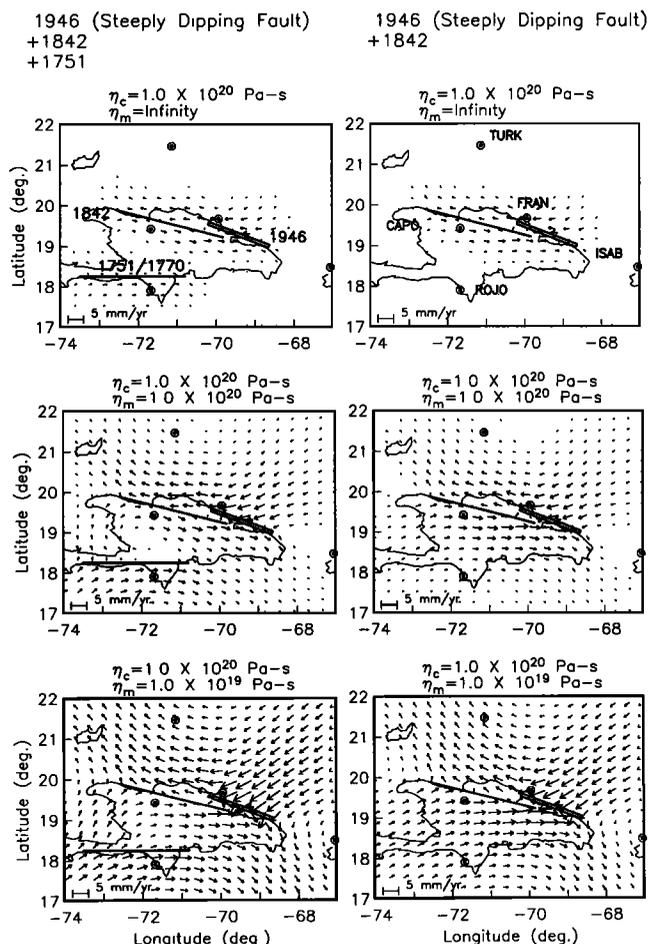
Figure 2. Left panels: postseismic velocity fields over the period 1986-1995 calculated for the summed 1751, 1842, and 1946 events for three different viscosity combinations. Right panels: corresponding postseismic velocity fields for the summed 1842 and 1946 events only. A steeply dipping fault for the 1946 earthquake is assumed.

1946 (Shallowly Dipping Fault) +1842   1946 (Steeply Dipping Fault) +1842

η_c=3.0 X 10^{19} Pa-s
η_m=1.0 X 10^{19} Pa-s

η_c=3.0 X 10^{19} Pa-s
η_m=1.0 X 10^{19} Pa-s
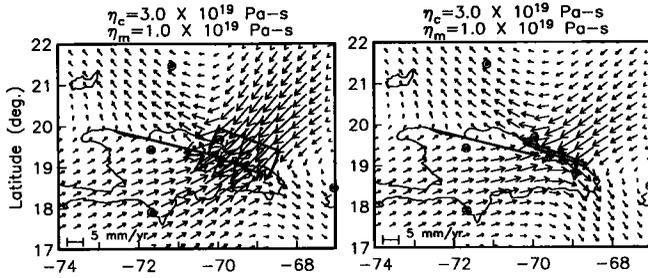
Figure 3. Postseismic velocity field averaged over 1986-1995 for the summed 1842 and 1946 events on both the shallowly dipping [D&W97] and steeply dipping [R&V95] fault geometries for the 1946 sequence.

Caribbean islands, where the crust-mantle transition is much deeper than in the Caribbean oceanic basin [*Edgar et al.*, 1971].
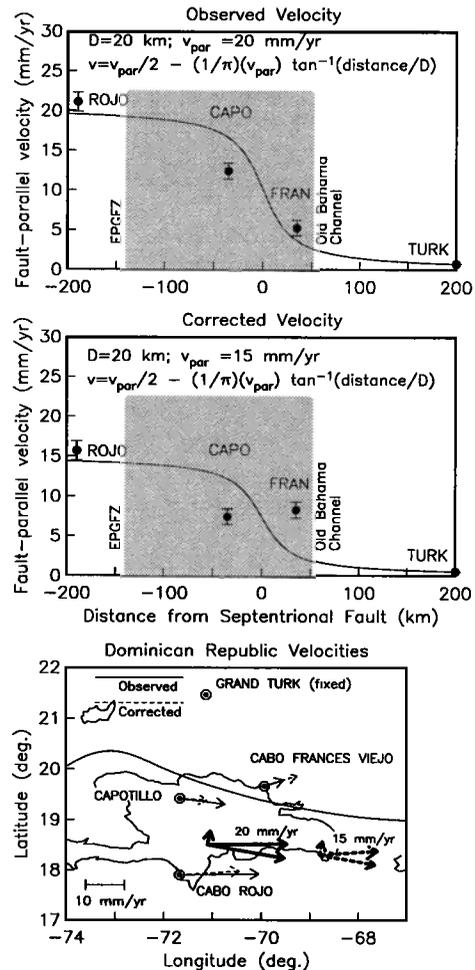
Postseismic Velocity

We calculated the total postseismic horizontal velocity fields averaged over the period 1986-1994 using the method of *Pollitz* [1992] for several different combinations of viscosities η_c and η_m, summing over the historic earthquakes described in the previous section. We first searched for those viscosity combinations which lead to the largest postseismic velocity contribution from the 1751/1770 sequence alone, finding that a lower crustal viscosity of about 10^{20} Pa s was optimal for producing the largest postseismic effects, regardless of the mantle viscosity. We then added the 1842 and 1946 sequences to the calculation. In these calculations we assumed the R&V95 fault geometry for the 1946 sequence. Figure 2 shows the resulting velocity fields for three viscosity combinations when 1751+1842+1946 events are summed (left hand panels) and when only 1842+1946 events are summed (right hand panels). Comparing the left hand and right hand panels it is clear that the 1751 event is responsible for about one-half of the total 3-4 mm/yr predicted at ROJO. At TURK the largest contributions come from the 1842 and 1946 sequences. It is apparent that the highest postseismic effects result from relatively low mantle viscosity.

In order to increase calculated post-1946 relaxation effects, it was found sufficient to fix the mantle viscosity at $\eta_m = 1.0 \times 10^{19}$ Pa s and choose the crustal viscosity in the range from 1.0 to 10.0×10^{19} Pa s. We compared the post-1946 velocity patterns obtained with the shallowly dipping fault model [D&W97] and the steeply dipping fault model [R&V95] for several viscosity combinations and found that the maximum velocity achieved at ROJO did not exceed 2 mm/yr with the steep fault geometry or 3 mm/yr using the shallow fault geometry. Figure 3 shows the postseismic velocity patterns for the summed 1842+1946 events for the maximal viscosity combination using both the shallow and steep fault geometries for the 1946 event. It shows that the summed velocity reaches about 3-4 mm/yr at ROJO. Allowing these absolute velocities to represent Caribbean to stable North American relative plate motion, the maximum amount of excess Caribbean plate velocity relative to stable North America which can be explained through postseismic relaxation effects is thus about 3-4 mm/yr. Since TURK exhibits a small relative velocity with respect to stable North America, the maximum amount of excess Caribbean plate velocity relative to TURK which can be accounted for by the same model is slightly higher at 5-6 mm/yr.

Discussion and Conclusions

We restrict attention to those Caribbean-North America (CAR-NAM) GPS observations relative to TURK, rather than a broader North American rigid plate framework, because the former observations are associated with smaller uncertainties [*Dixon et al.*, 1998]. The preceding results suggest that we can explain at most 5-6 mm/yr of the excess 10 mm/yr CAR-NAM velocity with a model of postseismic relaxation from historic earthquakes. Figure 4 shows results of inversion of horizontal GPS velocities for CAR-NAM velocity, assuming that all boundary deformation is concentrated on the SFZ, with a locking depth of 20 km. This approach appears justified by the fact that the SFZ is more active than the EFZ to the south [*Mann and*

Figure 4. GPS velocities resolved parallel to the SFZ are inverted for CAR-NAM relative plate motion with fault-parallel rate v_{par} assuming a locking depth of 20 km, both with and without correction for maximum possible postseismic relaxation effects. An analytic formula appropriate for strain accumulation in the strike-slip sense on an infinitely long vertical fault [*Savage and Burford*, 1973] is employed. Shaded area indicates the extent of the local plate boundary deformation zone between the Enriquillo-Plantain Garden Fault Zone and the Old Bahama Channel. Light solid/dashed lines in the lower panel indicate predicted velocities relative to fixed TURK without/with correction for postseismic relaxation effects. Heavy solid/dashed lines give the corresponding CAR-NAM relative motion decomposed into its fault-parallel and fault normal components; the fault-normal component is taken to be equal to the fault-normal component observed at ROJO.

Burke, 1984; *Prentice et al.*, 1993]. The corrected velocity corresponds to subtracting out postseismic relaxation effects for a particular viscosity model ($\eta_c = 3.0 \times 10^{19}$ Pa s, $\eta_m = 1.0 \times 10^{19}$ Pa s) and assuming the R&V95 fault geometry for the 1946 event. Use of the D&W97 1946 fault geometry would increase the size of the correction at ROJO by about 1 mm/yr and lead to a somewhat larger correction for FRAN. The value $v_{par} = 15$ mm/yr obtained after correction for postseismic relaxation effects shows that the fault-parallel velocity is underestimated by at least 4 mm/yr by Model NUVEL–1A.

Although the velocity observations of neither FRAN nor CAPO require a CAR-NAM relative velocity greater than 11 mm/yr, the observations at sites more distant from the plate boundary deformation zone appear to require a significantly greater rate of relative motion. In particular, site ISAB on Puerto Rico (Figure 1) is not associated with any obvious post-seismic relaxation effects from historic earthquakes (Figures 2 and 3). Its eastward velocity of 20 mm/yr suggests that $v_{par} = 20$ mm/yr is appropriate for the region. A similar inter-seismic loading analysis applied to site GTMO in eastern Cuba (Figure 1) also points towards $v_{par} \approx 20$ mm/yr. Taken together, these observations suggest that the postseismic relaxation effects manifested at ROJO are much smaller than the maximum possible, suggesting a lithospheric rheology producing only locally significant relaxation effects. An indeterminately high viscosity mantle combined with a crustal viscosity of $\sim 10^{20}$ Pa s, as considered in the top panels of Figure 2, then appears appropriate for the Hispaniola region. The postseismic relaxation predicted on such a model does not exceed 1-2 mm/yr at any of the highest-velocity GPS sites.

If postseismic relaxation can be ruled out as a significant contributor to the excess CAR-NAM velocity determined by GPS, then an alternative explanation is that the CAR-NAM rate predicted by the NUVEL–1A model is too small. Recent reinterpretation of marine magnetic anomaly data in the Cayman Trough [*Rosencrantz*, 1995] yields a rate of about 19 mm/yr. It is also recognized that the Cayman Trough represents not directly CAR-NAM motion but rather the motion between NAM and the Gonave microplate wedged between Cuba and Hispaniola [*Rosencrantz and Mann*, 1991]. The possibility of significant long term Gonave-CAR relative motion along the southern boundary of the Gonave plate (Rosencrantz, personal communication, 1997), suggests that geologic estimates of the relative motion rate are equal to or may exceed 20 mm/yr, in agreement with GPS estimates.

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