Constraints on present-day shortening rate across the central eastern Andes from GPS data

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Abstract. Two years of continuous GPS data from several sites in South America indicate that Arequipa in the southern Peruvian Andes has a velocity of 13±3 mm/yr (two standard errors) to the northeast with respect to stable South America. We interpret these data as reflecting a combination of elastic strain accumulation associated with a locked Nazca-South America subduction zone and a small amount of crustal shortening across the fold and thrust belt on the eastern margin of the Andes. Models of elastic strain accumulation for fully locked and partly locked subduction zones constrain shortening in the eastern Andes to 0-3 mm/yr (fully locked) and 0-12 mm/yr (partly locked), slower than some geologic estimates averaged over millions of years.

Introduction

The central part of the Nazca-South America plate boundary zone includes the Altiplano, a high plateau with a thick (>55 km) crustal root [James, 1972; Fukao et al., 1989] and the active Sub-Andean fold and thrust belt to the east, 600 km or more from the trench. Post-Oligocene (<26 Ma) crustal shortening is probably the dominant process for thickening central Andes crust [Isacks, 1988; Sempere et al., 1990]. Finite element models suggest that crustal shortening could have generated the entire topography of the Altiplano in less than 30 Ma [Wdowinski and Bock, 1994]. The locus of shortening has migrated east with time, from the Altiplano in the Miocene to its present position in the Sub-Andean zone [e.g., Suarez et al., 1983; Isacks, 1988]. Balanced cross sections across the central Bolivian fold and thrust belt suggest 210 km minimum shortening [Sheffels, 1990], all post-Oligocene. Schmitz [1994] estimated 320 km post-Cretaceous shortening for the entire south central Andes (trench to foreland). Assuming all shortening happened in the last 25 million years, these estimates translate to average shortening rates of 8.4-12.8 mm/yr. Cross sections in northern Bolivia suggest 135 km shortening in the last 5 Ma [Roeder, 1988], i.e., 27 mm/yr average rate.

Suarez et al. [1983] used rates of seismic energy release to suggest present-day shortening across the eastern Andes and sub-Andean belt at 1-2 mm/yr. If the seismic rate is indicative of longer term rates, it implies a dramatic slowdown of crustal shortening in the geologically recent past. Alternatively, if faster geological rates are indicative of present-day rates, significant deformation must be occurring aseismically, or the seismic estimate is biased by its short time interval, i.e. an "earthquake deficit" is accumulating.

Another noteworthy feature of the central Andes is the lack of recent large or great underthrusting earthquakes along the shallow subduction zone plate interface. The last great (M>8.0) earthquake was in 1877 while a large (M=7.8) earthquake occurred near Arequipa in 1913. In contrast, a series of large and great earthquakes have ruptured virtually the entire length of the southern Andes subduction zone over the last 50 years [e.g., Nishenko, 1991]. Characterizing the rate and locus of elastic strain accumulation in the region may shed light on some of the physical parameters (locking depth, plate coupling) affecting strain accumulation and release during the interplate seismic cycle.

This paper discusses new space geodetic data for the central Andes. Our data are consistent with strain accumulation near the subduction zone and slow to moderate shortening across the eastern fold and thrust belt.

Space Geodetic Data

GPS. We used data from five permanent GPS (Global Positioning System) sites spanning the actively deforming sub-Andean fold and thrust belt: Arequipa (Peru) and four sites in stable South America: Fortaleza and Brasilia
TABLE 1: GPS Site Velocities (ITRF-94)

<table>
<thead>
<tr>
<th>Site</th>
<th>North (mm/yr)</th>
<th>East (mm/yr)</th>
<th>Vertical (mm/yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arequipa</td>
<td>13.1±0.2</td>
<td>12.4±0.6</td>
<td>2.6±1.3</td>
</tr>
<tr>
<td>Brasilia</td>
<td>12.8±0.6</td>
<td>-2.8±1.8</td>
<td>16.5±4.8</td>
</tr>
<tr>
<td>Fortaleza</td>
<td>11.6±0.2</td>
<td>-2.5±0.5</td>
<td>-2.5±1.5</td>
</tr>
<tr>
<td>Kourou</td>
<td>10.3±0.2</td>
<td>-1.1±0.6</td>
<td>6.6±1.7</td>
</tr>
<tr>
<td>La Plata</td>
<td>13.4±1.0</td>
<td>2.0±2.2</td>
<td>3.7±5.3</td>
</tr>
</tbody>
</table>

(Brazil), Kourou (French Guiana) and La Plata, Argentina (Figure 1). Data from these and other globally distributed sites are analyzed daily at the University of Miami’s Geodesy Lab [Dixon et al., 1997], most from mid-1994 onwards. Brasilia and La Plata became operational in 1995. We used high precision non-fiducial satellite orbits and the GIPSY analysis software from the Jet Propulsion Laboratory [Zumberge et al., 1997]. Station velocities are defined in ITRF94 [Boucher et al., 1996]. Table 1 lists site velocities based on least squares fits to daily position estimates, weighted by 1/σ², where σ is scaled formal error. We define stable South America by minimizing velocities of Fortaleza, Kourou, Brasilia and La Plata. The velocity of Arequipa relative to stable South America is 13±3 mm/yr (95% confidence) in a direction 72°±6°E (Table 2).

SLR. Robaudo and Harrison [1993] used satellite laser ranging (SLR) to estimate Arequipa’s velocity. Since Arequipa was the only SLR station in South America, they defined Arequipa’s velocity with respect to stable North America, and used the NUVEL-1 plate motion model [DeMets et al., 1990] to estimate Arequipa’s motion with respect to stable South America. We updated this analysis to reflect the revised NUVEL-1A model velocities [DeMets et al., 1994], obtaining a velocity for Arequipa of 20±6 mm/yr in a direction 80°E (Table 2), similar in rate and direction to the GPS result.

Our GPS result is based on direct measurement. Unlike the SLR result, it employs no assumptions concerning relative motion of North and South America or equivalence between velocities on geodetic and geologic time scales. The similarity between GPS and SLR results (equivalent at 95% confidence in rate and azimuth) suggests that the NUVEL-1A model is a reasonable predictor of current North America-South America motion [Dixon and Mao, 1997]. The azimuths of the NUVEL-1A velocity for Nazca-South America and Arequipa’s velocity (GPS or SLR) are also very similar. We take the GPS velocity and its 95% confidence bounds (10-16 mm/yr) as the likely range for Arequipa’s present day motion relative to stable South America, and explore tectonic implications.

Discussion

Arequipa’s motion likely reflects both elastic (recoverable) strain accumulation associated with the seismic cycle at the locked subduction zone, and permanent shortening in the sub-Andean fold and thrust belt several hundred kilometers east. Shortening probably occurs via seismic strain accumulation and release, but unlike the first mode results in a permanent change in the position of Arequipa with respect to stable South America.

If we estimate that part of Arequipa’s motion due to elastic strain associated with a locked subduction zone, then remaining motion yields an estimate of permanent shortening in the fold and thrust belt. Elastic strain due to a locked subduction zone can be estimated via elastic half space models for a locked, dipping thrust fault [Savage, 1983]. Dixon [1993] used a similar approach to model GPS data in Central America near the Costa Rica trench. This approach requires estimates of depth and average dip of the locked zone.

Tichelaar and Ruff [1991] studied earthquakes along the subduction interface for the central and southern Andes and found locking depths in the range 36-53 km. Since our region of interest (~17°S) has not ruptured in a large earthquake since modern seismic instrumentation was deployed, locking depth must be estimated indirectly. Tichelaar and Ruff [1991] suggested that for 18°S to 24°S, south of our study area, the coupled zone extends at least to 43-48 km. We use 50 km as an estimate of locking depth, and use 46, 50 and 54 km to span the plausible range.

Tichelaar and Ruff [1991] estimated dips for locked seismic zones in two ways. The dip of the plane connecting the deepest thrust earthquake and the trench has

TABLE 2: Velocities Relative to S. America

<table>
<thead>
<tr>
<th>Source</th>
<th>Rate (mm/yr)</th>
<th>Azimuth (deg E of N)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arequipa/GPS</td>
<td>13±3</td>
<td>072°±6°</td>
</tr>
<tr>
<td>Arequipa/SLR</td>
<td>20±6</td>
<td>080°</td>
</tr>
<tr>
<td>Nazca plate</td>
<td>78</td>
<td>079°</td>
</tr>
</tbody>
</table>

1. This study; all errors at 95% confidence
2. SLR+NUVEL-1A model [Robaudo and Harrison, 1993]
3. NUVEL-1A [DeMets et al., 1994]
a range of 12°-19°, whereas focal mechanisms for the deepest thrust earthquake give dips of -20°-30°. The latter are about 10° steeper than the former, due to steepening of the subduction zone with depth. South of our study area (18°-24°) overall geometry gives a slab dip of about 18°, implying a dip at the deepest portion of about 26°. We modeled the locked subduction zone as three segments with progressively steeper dips: 10° (0-15 km), 18° (15-35 km), and 26° (35 km to the maximum locking depth). Single segment models give similar results.

To model surface strain due to a multi-segment locked subduction zone with or without additional thrust faults, we combined the boundary element modeling program "3D-DEF" [Gomberg and Ellis, 1994] and the approach of Savage [1983]. Surface strain for a subduction zone is modeled by imposing back slip at the plate convergence rate [78 mm/yr; DeMets et al., 1994] to simulate the locked segment. We compared models where the entire convergence rate is accommodated on a locked plate interface west of Arequipa ("subduction only," Figure 2) to models where most convergence is accommodated there, but some convergence is accommodated on thrust faults east of Arequipa ("subduction+shortening"). To model the sub-Andean fold and thrust belt, we used faults 600 and 650 km from the trench, each accommodating half the geologic shortening (results are not sensitive to the actual location of the faults, their geometry or the partitioning of motion between them). Thrust faults were assumed to dip 30°W and be locked to 20 km depth, typical values for continental crust. We used shortening rates of 3, 12 and 27 mm/yr, spanning seismic and geologic estimates. We also ran models where 50% of slip on the subduction zone is accommodated aseismically (partly decoupled plate interface) for comparison with the fully locked models (strongly coupled interface). The fully locked models with no to slow (0-3 mm/yr) shortening are in good agreement (95% confidence) with the GPS data. Locked subduction zone models incorporating moderate to fast shortening (>6 mm/yr) disagree with the GPS data at 95% confidence. The high shortening rate model (27 mm/yr) can be excluded at 99% confidence. Models with 50% aseismic subduction allow faster (up to 12 mm/yr) shortening, but exclude the high (27 mm/yr) shortening rate. The high shortening rate model can be excluded at 99% confidence even for completely decoupled (100% aseismic slip) models, i.e., with no elastic strain accumulation in the subduction zone, the maximum shortening rate in the eastern Andes is just Arequipa’s velocity relative to the craton, 13±3 mm/yr.

Assuming a fully locked subduction zone, our data are consistent with slow rates of eastern Andes shortening suggested by seismic data, implying that faster geological estimates averaged over 5-25 million years (8-27 mm/yr) are not indicative of the modern rate. If correct, there is no earthquake deficit in this region, nor any significant aseismic deformation. This is a surprising result, since seismic strain rates are usually much less than total strain rates. It also implies a recent slowdown in the rate of crustal shortening, for which there is no firm evidence. Changes in shortening rate have been tied to changes in Nazca-South America convergence rate. Although the average convergence rate for the last 10Ma may have been less than the previous 10Ma, the timing of the slowdown is poorly constrained [Pardo-Casas and Molnar, 1987]. If Roeder’s [1988] high shortening rate estimate is correct, our data and the locked model imply a more dramatic and more recent (post-5Ma) slowdown.
Assuming a partly decoupled subduction zone, our data allow higher shortening rates (up to 12 mm/yr), allow for aseismic deformation in the eastern Andes, and do not require a recent slowdown in shortening. Additional site velocity data west of Arequipa would better constrain subduction zone models, while data east of Arequipa would directly measure geologic shortening. These data are currently being collected by regional GPS experiments.

The rate of strain accumulation near the trench implied by our data, and the pattern of rupture in previous events, allow estimates of the size of future events. It has been 82 years since the last large earthquake in the area (1913). Assuming a locked plate interface, about 7 meters of slip has accumulated. Tichelaar and Ruff [1991] suggest that interplate thrust events south of our area have rupture areas ~10^4 km^2. If similar rupture areas characterize our study area, release of half the accumulated slip in a single future event would produce a seismic moment of 10^{34} dyne-cm, equivalent to an M_w-8 event [Kanamori, 1983]. Accumulated slip is reduced by half for a 50% aseismic subduction model, reducing the corresponding estimate of earthquake magnitude. Our analysis indicates where GPS data need to be collected and to what accuracy to distinguish subduction zone coupling models (Figure 2). Seismic vs 50% aseismic models should be distinguishable after a few years of observations.

Note added in proof. Preliminary data from 25 sites observed in 1994 and 1996 confirm results presented here and are consistent with partly aseismic subduction.

Acknowledgments. This work was supported by NASA's Geodynamics program. We thank Shimon Wdowinski for helpful discussions, Mike Hefflin and Ken Hurst for advice on GPS data analysis, two anonymous reviewers for helpful comments, and the geodetic community for maintaining a global network of permanent GPS stations.

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(Received February 12, 1996; revised October 22, 1996; accepted March 7, 1997.)