Why short-term crustal shortening leads to mountain building in the Andes, but not in Cascadia?

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Subduction-induced crustal shortening, now measured by the GPS across many subduction zones, has led to mountain building in the Andes but not in Cascadia and some other Andean-type convergent plate boundaries. Here we use a two-dimensional viscoelastoplastic finite element model to explore how the GPS-measured short-term strain relates to long-term mountain building. We show that previously proposed causative factors of mountain building can be represented by two model parameters: the strength of mechanical coupling on the plate interface, and the yield strength of the overriding plate. The critical condition for producing permanent (plastic) crustal shortening, hence mountain building, is for the plastic yield strength of the plate interface to be higher than that of the overriding plate. Strong trench coupling and a weak lithosphere explain the Andean mountain building, whereas weak trench coupling in Cascadia allows short-term crustal shortening to be restored periodically by trench earthquakes and aseismic slips. Citation: Luo, G., and M. Liu (2009), Why short-term crustal shortening leads to mountain building in the Andes, but not in Cascadia?, Geophys. Res. Lett., 36, L08301, doi:10.1029/2009GL037347.

1. Introduction

Most workers have concluded that the Andes, the archetype of the Andean-type orogens formed along active continental margins (Figure 1a), have resulted mainly from subduction of the oceanic Nazca plate under South America [Allmendinger et al., 1997; Isacks, 1988]. Recent Global Positioning System (GPS) measurements in the central Andes show that present-day crustal shortening absorbs roughly half of the ~65 mm/yr Nazca-South American plate convergence [e.g., Kendrick et al., 2001; Khazaradze and Klotz, 2003; Norabuena et al., 1998] (Figure 1a). This rate is higher than that estimated from geological record (<15 mm/yr) [Kley and Monaldi, 1998], implying that much of the present-day crustal shortening is transient and will be recovered during future earthquakes [Liu et al., 2000; Norabuena et al., 1998].

If all crustal shortening in the subduction zone is transient, then no permanent crustal shortening, and hence no mountain building, can be produced. This situation seems applicable to the Cascadia subduction zone, where the Farallon plate (now the Juan de Fuca plate) has been subducting under the North American plate since the Mesozoic (Figure 1b). Despite the GPS measurements of ~10–20 mm/yr crustal shortening in Cascadia, no significant mountain building has occurred there since the Eocene [Gabrielse and Yorath, 1992].

Previous studies have suggested many factors contributing to mountain building in subduction zones, including the change of plate convergence rate, trenchward absolute motion of the overriding plate, age and angle of the subducting slab, subduction of oceanic ridge and seamounts, the amount of subducted sediments, and thinning of mantle lithosphere of the overriding plate [e.g., Allmendinger et al., 1997; Lamb and Davis, 2003; Silver et al., 1998]. We show here that these factors can be represented by two competing parameters in a mechanical model: the strength of trench coupling versus the strength of the overriding plate. Using a two-dimensional viscoelastoplastic finite element model, we explore how these two parameters control whether or not short-term strain associated with the cycles of trench earthquakes leads to long-term mountain building in subduction zones.

2. Numerical Model

Previous models of the Andean-type plate boundaries have focused either on short- or long-term deformation. The models of short-term deformation, using elastic or viscoelastic rheology, simulate transient strain during seismic cycles without linking the strain to mountain building [e.g., Bevis et al., 2001; Norabuena et al., 2004]. The models of mountain building, using viscous or viscoplastic rheology, assume continuous and permanent crustal shortening [e.g., Sobolev and Babeyko, 2005], and thus do not explain why subduction-induced crustal shortening leads to mountain building in the Andes but not in other subduction zones.

We have developed a two-dimensional viscoelastic-plastic finite element model for the Andes [Luo and Liu, 2009]. Here we generalize this model for Andean-type plate boundaries (Figure 2). The oceanic plate subducts under the continental plate at the angle of 30° with relative convergence rate of 50 mm/yr. We take the top 30 km of the subducting plate to be elastic, and the top 20 km of the overriding plate to be brittle (elastoplastic). A viscoelastic layer is used for the lower crust and the upper mantle of both plates. The subduction interface is modeled with two layers of special finite elements: the bottom layer with relatively low and depth-variable viscosity to simulate interseismic creeping, and the top layer with elastoplastic rheology to simulate the stick-slip seismogenic zone. The plastic deformation, both within the continental crust and on the subduction interface, is simulated with the Mohr-
Coulomb yield criterion and non-associated flow rule [Luo and Liu, 2009].

3. Model Results

[7] In the reference case, we allow ~15 mm/yr of the 50 mm/yr plate convergence to be accommodated by creeping on the subduction interface; the remaining ~35 mm/yr is absorbed by crustal shortening in the overriding plate during interseismic periods. The plate interface is allowed to slip every ~200 years by choosing a proper stress drop (a few MPa), with ~7 m coseismic slip averaged on the subduction thrust, which is typical for M_w > 8.0 events in the Peru-Chile trench [Nishenko, 1991]. Once the model has reached a quasi-steady state, we calculate strain distribution and evolution during each seismic cycle. The results for each cycle are similar; one cycle is shown in Figure 3a. The interseismic displacement is evenly distributed across the overriding plate (Figure 3a), whose rheology is laterally homogeneous in this case. About half of the interseismic crustal shortening is restored during the trench earthquake; the rest is restored by postseismic viscous relaxation. In this case no plastic deformation, hence no mountain building, occurs in the overriding plate. Allowing the plate interface to slip in shorter intervals would produce smaller coseismic slips, but will not necessarily affect the partitioning between transient and permanent strains. Note that the co- and post-seismic displacement for one earthquake cycles is too small (~10 m) in comparison to the model domain (~1000 km), hence cannot be seen in Figure 3a.

[8] Permanent (plastic) crustal shortening needs to be produced and accumulated through the cycles of trench earthquakes to lead to mountain building. This requires 1) strong mechanical coupling on the plate interface to transmit sufficient compressive stress to the overriding plate, and 2) sufficiently low plastic yield strength in the overriding plate for plastic shortening to occur. With higher plastic yield strength on the plate interface than that in Figure 3a but otherwise the same parameters, the model produced both viscoelastic and plastic strains in the overriding plate (Figure 3b). While the viscoelastic strain is largely restored during coseismic rebound and postseismic relaxation, about 2 m of plastic shortening is accumulated.

Figure 1. GPS velocities, selected focal mechanism solutions, and topographic relief in the (a) central Andes and (b) Cascadia. The Andean GPS data are from Chlieh et al. [2004], Kendrick et al. [2001], Khazaradze and Klotz [2003] and Klotz et al. [2001]. WC, Western Cordillera; AP, Altiplano plateau; EC, Eastern Cordillera; FTB, Subandean fold and thrust belt. The Cascadia GPS data are from Mazzotti et al. [2003], McCaffrey et al. [2007] and Miller et al. [2001]. SAF, San Andreas Fault.

Figure 2. Mesh and boundary conditions of the finite element model. The subduction interface is modeled with two layers: an elastoplastic layer for the stick-slip seismogenic thrust (red) and a viscous layer for aseismic slip (yellow).
over each seismic cycle of ~200 years (i.e., 10 mm/yr). The same results can be produced by reducing the plastic yield strength of the overriding plate. Furthermore, reducing the viscosity in the lower crust and upper mantle causes higher deviatoric stress in the brittle crust, and hence has similar effects.

Note the values of trench coupling and the yield strength of the overriding plate in the models are unconstrained. What matter here are their relative values: when the plastic yield strength on the plate interface is lower than that for the overriding plate, interseismic crustal shortening will be recovered by coseismic slip of the trench earthquake and associated postseismic viscous relaxation, hence no permanent crustal shortening occurs in the overriding plate.

4. Trench Coupling and Lithospheric Strength: Andes Versus Cascadia

The values of plastic yield strength on the plate interface and for the overriding plate, unfortunately, cannot be directly measured or constrained. Nonetheless, geological observations may provide useful proxies for these values. Interplate shear stress in the subduction zone, for example, is a good indicator of the plastic yield strength there, and force-balancing calculations indicated high shear stress in the Andean trench: 25–110 MPa from a viscous model [Husson and Ricard, 2004], and ~37 MPa from an analytic model [Lamb, 2006]. Lamb and Davis [2003] attributed the high shear stress to the lack of trench sediments, which reduces lubrication on the subduction interface. Strong trench coupling in the central Andes is consistent with deep trench and negative gravity anomalies there [Iaffaldano and Bunge, 2008].

Similar force balance calculations, as well as forearc heat flow data, argue for low shear stress (<20 MPa) on the Cascadia subduction interface [Lamb, 2006; Wang and He, 1999; Wang et al., 1995]. The thick (2–3 km) Cascadia trench sediments [Flueh et al., 1998] imply lubrication, high pore fluid pressure and low shear stress on the plate interface [Lamb, 2006]. The stress state in the Cascadia forearc shown by small earthquakes also indicates weak plate coupling [Wang et al., 1995].

Following the force balance analysis of Lamb [2006], we include the effects of sea water and found that the shear stress on subduction interface, \( \tau \), can be determined as

\[
\tau = \frac{\sin \theta}{4} \left( \frac{g h - \rho_p gd^2}{h} \right)
\]

where \( \theta \) is the shallow dipping angle of the subducting slab; \( \rho \) and \( \rho_p \) are the densities of forearc wedge and sea water, respectively; \( g \) is gravitational acceleration; \( h \) is the elevation difference between mountain and trench; and \( d \) is the trench depth (inset in Figure 4). The high elevation of the Andes and the deep Peru-Chile trench demand high interplate shear stresses; the opposite is true for Cascadia (Figure 4).
The interplate shear stress in the subduction zone of the central Andes and Cascadia from an analytic solution of force balance. The shadowed area and dashed lines respectively show effects from variations of $h$ and $d$, compared with those from their average values (solid lines). Thick segments of the lines correspond to shallow subduction angles for the two subduction zones; stars indicate averaged shear stress on subduction thrust from the models, $\rho = 2.8 \times 10^3$ kg/m$^3$, $\rho_s = 1.0 \times 10^3$ kg/m$^3$; $g = 10$ m/s$^2$. Other parameters are from Lallemand et al. [2005], Lamb [2006], and Oleskevich et al. [1999]. Inset shows the force balance in the forearc of a subduction zone.

The plastic yield strength of the Andean and the Cascadian lithosphere is hard to quantify, but high surface heat flow, low seismic velocities, crustal partial melting, high temperature by xenolith thermobarometry, occurrence of ignimbrites and basaltic volcanism, and high electronic conductivity all point to a thermally weakened lithosphere of ignimbrites and basaltic volcanism, and high electronic conductivity all point to a thermally weakened lithosphere in the Andes [Asch et al., 2006; Currie and Hyndman, 2006; Schilling et al., 2006]. This has been used to explain why significant mountain building in the Andes occurred in the past ~30 Ma, although the Nazca plate has been subducting beneath the South American plate at least since early Jurassic [Allmendinger et al., 1997; Isacks, 1988]. The Cascadian lithosphere has also been thermally weakened [Currie and Hyndman, 2006], but likely not as much as the Andean lithosphere. The GPS-measured crustal shortening is concentrated to the forearc in Cascadia (Figure 1b), which can be predicted in the model with a strong backarc lithosphere.

5. Discussion and Conclusions

The importance of trench coupling and yield strength of the overriding plate is self-evident in mechanics. For subduction to produce permanent (plastic) crustal shortening, hence mountain building, it needs strong trench coupling to transmit sufficient compressive stress to the overriding plate, and the overriding plate to be sufficiently weak to deform plastically. By simulating strain evolution during cycles of trench earthquakes, our model provides a useful tool to explore how short-term crustal deformation as measured by space geodesy links to long-term tectonics and mountain building.

In summary, we found that for subduction to lead to significant mountain building, the yield strength on the plate interface needs to be higher than that in the overriding plate. Strong interplate coupling in the Peru-Chile trench and thermal weakening of the Andean lithosphere explain the Andean mountain building, whereas weak trench coupling in Cascadia allows interference crustal shortening to be fully restored through the cycles of trench earthquakes and aseismic slip, leaving no significant plastic strain for mountain building.

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