Rise of the Andes

Carmala N. Garzione,1* Gregory D. Hoke,2 Julie C. Libarkin,2 Saunia Withers,3 Bruce MacFadden,4 John Eiler,5 Prosenjit Ghosh,6 Andreas Mulch7

The surface uplift of mountain belts is generally assumed to reflect progressive shortening and crustal thickening, leading to their gradual rise. Recent studies of the Andes indicate that their elevation remained relatively stable for long periods (tens of millions of years), separated by rapid (1 to 4 million years) changes of 1.5 kilometers or more. Periodic punctuated surface uplift of mountain belts probably reflects the rapid removal of unstable, dense lower lithosphere after long-term thickening of the crust and lithospheric mantle.

The surface uplift of mountain belts, such as the central Andes plateau, has long been thought to be the isostatic response of shortening and thickening of the continental crust. Recently developed isotopic techniques allow us to determine the uplift history of the central Andes independently from the shortening history. These results show that shortening and uplift are temporally decoupled, with shortening and thickening happening over protracted periods of time, whereas uplift occurs geologically rapidly. Thus arises a paradox: Why does slow, continuous shortening and thickening not produce slow, continuous isostatic uplift in the central Andes?

Both crustal thickening and the removal of relatively dense mantle or lower crust can generate isostatic surface uplift (1, 2). Paleoelevation studies help resolve the geodynamic evolution of mountain belts because the rate and lateral extent of surface uplift depends on the processes involved. Here, we synthesize the elevation history of the central Andes, Earth’s second largest mountain belt. We then compare paleoelevation estimates to histories of regional incision, sedimentation, shortening, and volcanism within the mountain belt to characterize lithospheric evolution and the geodynamic mechanisms that led to surface uplift.

The central Andean plateau (Fig. 1), with a width of ~400 km and an average elevation of ~4 km, is a typical example of an active plate margin where oceanic lithosphere is subducted beneath continental lithosphere. At its widest, the central Andean plateau consists of the internally drained Altiplano basin at an elevation of ~3800 m that is bounded by the Western and Eastern Cordilleras, where peak elevations exceed 6 km. The Western Cordillera is a chain of volcanic edifices associated with the modern Andean magmatic arc, whereas the Eastern Cordillera and Altiplano basin record a history of folding and faulting. The central Andes have a protracted crustal shortening history spanning the last 50 million years (My) (3–5) that has generated crustal thicknesses of ~70 km below the highest topography in the Eastern and Western Cordillera and 60 to 65 km below the central Altiplano (6). Geophysical observations suggest that eclogitic lower crust is absent beneath much of the plateau (6). The mantle between 16°S and 20°S shows the lowest P wave velocities below the Altiplano/Eastern Cordillera transition, suggesting that virtually all of the mantle lithosphere has been...
removed (7–9). In addition, high $^{3}He$/$^{4}He$ ratios in hydrothermal fluids and gases across much of the Altiplano and Eastern Cordillera indicate the degassing of mantle asthenosphere-derived magmas (10). Together, these observations support previous suggestions for the southern Altiplano and Puna (11, 12) that both mantle lithosphere and eclogitic lower crust were removed below much of the Altiplano and the western part of the Eastern Cordillera (6). Removal of lower lithosphere might occur rapidly by delamination or convective removal (1, 13) or gradually by ablative subduction of foreland cratonic lithosphere (14, 15). Either case results in a influx of lighter asthenosphere, generating surface uplift of several kilometers. However, rapid removal of lower lithosphere would result in surface uplift in as little as several million years, whereas gradual removal by ablative subduction would generate surface uplift over tens of millions of years, coincident with crustal shortening.

**Climate Trends and Subsidence History**

The Altiplano and Eastern Cordillera contain thick accumulations of Oligocene through late Miocene fluvial, floodplain, and lacustrine deposits (16). The oxygen ($^{18}O$) and hydrogen ($^{2}H$) isotopic composition of paleosol carbonates and authigenic clays from volcaniclastic units provide a record of meteoric water composition that is the basis for stable isotope paleoelevation estimates (Fig. 2). Stable carbon isotope ($^{13}C$) values of paleosol carbonates provide a record of plant respiration rates that can be used as a proxy for aridity (i.e., lower plant respiration reflects a more arid climate). Depositional environments within a 3.6-km-thick succession preserved in the eastern limb of the Corque syncline include fluvial channel sandstones and floodplain mudstones in the lower 1.4 km and upper 700 m, as well as a widespread freshwater-playa lake system that can be traced more than 100 km along strike. Both sedimentology and carbon isotopes in this section suggest that the central Altiplano became more arid between ~10 and 6 million years ago (Ma) (17). Fluvial channel deposits decreased in thickness and lateral extent upsection, further suggesting a decrease in discharge. Over the same time interval, $^{13}C$ values of pedogenic carbonates increased by ~3 per mil (%), suggesting a decrease in plant-respired CO$_2$ (17). Despite evidence for increasing aridity, the $^{18}O$ values of palustrine and paleosol carbonates ($^{18}O_c$) decrease by ~3 % (Fig. 2), a change opposite of the expected trend for higher rates of surface-water evaporation. More positive $^{18}O$ values observed in the older part of the record are synchronous with observations of wetter conditions, suggesting that evaporative enrichment of $^{18}O$ is an unlikely cause for the trend to more negative $^{18}O$ values over time.

Sedimentation rates in the Altiplano reflect rates of subsidence relative to the surrounding topography. Sedimentation rates in the central Altiplano and Puna dramatically decreased after 10 Ma (18) (Fig. 3). Between 13 and 9 Ma, sedimentation rates were extremely high, averaging 880 m/Ma (Fig. 3). During this time period, widespread lacustrine deposition suggests underfilled basin conditions. By 8.6 Ma, fluvial deposition resumed and subsidence decreased dramatically, averaging 0.12 mm/year.

**Paleoelevation Constraints**

Shallow marine deposits of the El Molino Formation require that the Altiplano lay at sea level at the end of Cretaceous time (19). Paleotemperature estimates derived from fossil-leaf physiognomy in the northern Altiplano and Eastern Cordillera (Fig. 1) suggest that paleoelevations were <1.3 km at ~15 to 20 Ma (20) and <2 km by ~10 Ma (21). Both the oxygen isotopic composition of rainfall and surface temperatures vary as a function of elevation. Stable oxygen isotope values of pedogenic carbonate and carbonate cement should reflect the composition of soil water and shallow groundwater, which is a reflection of rainfall composition and near surface temperature. The abundance of $^{13}C$ and $^{18}O$ bonds relative to a random distribution of carbon and oxygen isotopes in carbonate (measured by the $^{18}O_c$ value of CO$_2$ extracted from carbonates) should record the soil carbonate precipitation temperature (22). Carbonate $^{18}O$ values and $^{18}O_c$ temperature estimates decrease with time (17, 22, 23), suggesting that elevations increased by 2.5 ± 1 km during late Miocene time, consistent with low elevation estimates from fossil leaves.

We supplement these elevation records with $^{18}O$ data from authigenic clays in late Miocene ash deposits in the Callapa section and $^{18}O$ data from late Oligocene to early Miocene pedogenic carbonates from the Salla and Huayllapucara/Totora Formations (16) (Figs. 1 and 2 and tables S4 and S6). One challenge in interpreting stable isotope records of elevation is that they can be biased toward lower elevation estimates (i.e., more positive values) by increased surface-water evaporation associated with climate change. $^{18}O$ data of authigenic clay minerals in combination with the oxygen isotope carbonate record ($^{18}O_c$) provide a qualitative assessment of this bias because of the retention and fractionation behavior of hydrogen and oxygen in soil and lake water during evaporation (24). $^{18}O$ values of smectite from volcanic ashes of the Callapa Formation parallel $^{18}O$ values (Fig. 2) and decrease by about 10 to 20 % during the late Miocene. Despite increasingly arid climate, the combined $^{18}O_c$ and $^{18}O$ data show trends toward more negative isotopic compositions of meteoric water, supporting the inference that the decrease in $^{18}O_c$ reflects a change in surface elevation of the Bolivian Altiplano.

Before the late Miocene, $^{18}O_c$ values and paleotemperature estimates suggest a long history of fairly stable surface temperatures and isotopic compositions of surface waters (Fig. 2), perhaps reflecting minimal surface elevation change between ~25 and 10 Ma. Using the paleotemperature estimates from an early middle Miocene fossil-leaf assemblage (21), we converted the $^{18}O_c$ values of late Oligocene to early Miocene soil carbonates to surface-water values ($^{18}O_w$) (25). For the modern isotopic lapse rate of $h = 472.5$ $^{18}O_{rainfall} – 2645$ [where $h = 18$ in elevation in meters (17)], paleoelevation is <2.3 ± 1 km (26) between 10 and 25 Ma, broadly consistent with fossil-leaf estimates (20) (Fig. 4). Before 25 Ma, the only paleoelevation estimates come from pedogenic carbonates of the Salla Formation. Both high $^{18}O_c$ values and high paleotemperature estimates, based on $^{18}O_c$ values (table S5), suggest elevations close to sea level. [Assuming modern temperature and $^{18}O_c$ lapse rates, reconstructed $^{18}O_c$ are similar to modern values in the Amazon foreland, and temperatures are slightly warmer than those in the foreland (Fig. 4).] The relatively positive late Oligocene to early Miocene $^{18}O_c$ values cannot be explained by diagenesis of carbonate because both higher temperatures and/or diagenesis in the presence of later fluids should produce more negative $^{18}O_c$ values (27), which are not observed. In addition, $^{18}O_c$ values suggest reasonable temperatures for surface environments, as opposed to the higher temperatures that might result from burial diagenesis.

The $^{18}O_c$, $^{18}O$, and $^{18}O$ compilation (Figs. 2 and 4) suggests that there was at least one discrete
pulse of rapid surface uplift of ~1.5 to 3.5 km (2.5 ± 1 km) between ~10 and 6 Ma and perhaps an earlier phase of surface uplift at ~25 Ma. However, limited data between 30 and 20 Ma preclude an understanding of the nature and extent of the older event. In the following discussion, we review geologic histories of the magmatism, shortening, and incision within the central Andean plateau that, when viewed with sedimentation rates and surface uplift history, shed light on the regional geodynamic processes that induced late Miocene surface uplift.

**Magmatism and Distribution of Shortening**

Widespread felsic magmatism in the Andean plateau began between 18°S and 24°S at ~25 Ma (29) and has been attributed to steepening of the subducting Nazca slab (29). Despite its wide extent, the volume of pre-late Miocene magmatism was small. Most activity (>85%) between 19°S and 23.5°S occurred between ~8.5 and 4 Ma (30, 31). Mafic lavas erupted throughout the northern and central Altiplano beginning at ~7.5 to 5.5 Ma (32, 33) and at ~7 to 3 Ma in the southern Altiplano and Puna (13, 34) (Fig. 5D). One group of lavas that erupted between 25°S and 26.5°S shows trace element and radiogenic isotopic compositions characteristic of an asthenospheric source, inferred to reflect the removal of eclogitic lower crust and mantle lithosphere beneath the southern Altiplano and Puna plateaus (34).

From 30 to 10 Ma, the Andean plateau experienced east-to-west shortening by ~6 to 12 mm/year across the plateau (3, 5). From 10 to 7 Ma, while elevation increased, shortening ceased and deformation propagated eastward into the Subandean zone (4, 35) (Fig. 5C). This shift in the locus of shortening is consistent with surface uplift, which should decrease the horizontal deviatoric compressive stress under the plateau while applying greater force per unit length to the surrounding lowland.

**Geomorphology**

Changes in the relief structure of the Plateau also imply that surface uplift occurred since ~10 Ma (36–38). Paleosurfaces on both the eastern slope of the Eastern Cordillera (36, 37) and the western slope of the Western Cordillera (38–43) reflect remnants of low-relief drainage systems that were active between ~7 and 12 Ma in the Eastern Cordillera (36, 37) and until ~10 Ma on the western slope (44). Widespread incision of the paleosurfaces in the Eastern Cordillera began by ~6.5 Ma (37) and in the Western slope began between ~9 and 5.5 Ma (40, 42, 43). Cooling ages of minerals in the Eastern Cordillera also imply that rapid incision began between ~15 and 6 Ma in the absence of substantial shortening (45, 46). Reconstructions of the relief in these incised valleys suggest ~2 km of surface uplift of the Eastern Cordillera (47) and ~1 to 2.5 km of surface uplift of the Western Cordillera (40–42). The western slope of the Andes (north of ~30°S) has had an arid-to-hyperarid climate since at least 15 Ma (48–50), indicative of atmospheric circulation patterns similar to the modern pattern, in which rainfall is derived predominantly from the west. Despite dramatically different climates between west and east, the similar timing of incision on the western and eastern slopes supports the notion that incision was induced by surface uplift and rotation of the slopes.

The widespread extent of incision implies that the entire width of the mountain belt—over at least 5° latitude—rose (Fig. 1A) (51). The regional late Miocene paleotopography can be reconstructed using surface uplift estimates in Fig. 4 (for the Altiplano) and the magnitude of relief generated during incision of the Eastern and Western cordilleras (40, 47) (Fig. 1B). There is a large difference in cross-sectional area between the average modern elevation of the paleosurface and the 10-Ma reconstructed topography. For crustal thickening to account for this difference would require shortening rates in excess of 40 mm/year over the 1 to 3 My during which surface uplift occurred. This is four times greater than the observed rates in the Andean plateau over the past 40 My, which seems implausible. Flow of middle-lower crust from the Eastern and Western cordilleras to the Altiplano (52) fails to explain the simultaneous rise of the cordilleras, which should subside or remain at the same elevation as lower crustal material flows laterally. We conclude that the rapid rate, magnitude, and regional extent of surface uplift, in addition to crustal thickening, also require a mantle contribution, most likely the isostatic response to removal of eclogite and mantle lithosphere.

Climatic responses to Andean surface uplift may extend beyond South America. For example, the presence of the Andes deflects the Intertropical Convergence Zone to north of the equator in the Pacific, which influences the strength and distribution of monsoonal climates with Pacific teleconnections (53). It is therefore possible that punctuated Andean surface uplift contributed to the reorganization of south Asian climate observed in the late Miocene.

**Geodynamic Implications**

Crustal thicknesses of 60 to 70 km in the central Andes are the result of protracted shortening and thickening over the past 50 My. Despite extensive crustal thickening, regional reconstruction of paleotopography suggests paleoelevations of <2 km in the Altiplano and 2.5 to 3.5 km in the Eastern and Western cordilleras until ~10 Ma (51). These anomalously low paleoelevations probably reflect the presence of dense eclogitic lower crust, which held the surface down. An analogous region today might be the western Sierras Pampeanas in Argentina, with 60-km-thick crust but average elevations of ~1 km (54).

Geologic observations suggest that the internal structure of the Andean lithosphere changed between ~10 and 6 Ma. During this time, the entire width of the plateau experienced surface uplift, deep incision of low-relief paleosurfaces initiated on the eastern and western slopes of the Andes, sedimentation rates within the Altiplano basin decreased dramatically while the basin...
transitioned from underfilled lacustrine environments to filled fluvial/floodplain environments, shortening ceased in the plateau and propagated into the foreland, and widespread and voluminous ignimbrite eruption began, followed closely in time by mafic volcanism (Fig. 5). Together, these observations are best explained by the removal of dense eclogite and mantle lithosphere, triggering regional surface uplift of ~1.5 to 2.5 km. The amount of surface uplift requires the removal of eclogite and mantle lithosphere of ~80 ±1000 m at elevations of <1 km. Assuming a Rayleigh distillation process for fractionation during condensation from a vapor mass, higher-than-modern global temperature in the late Oligocene to late Miocene would lower the δ18O versus-altitude gradient, which would cause an underestimation of paleoelevation for distillation from a vapor mass with a similar to modern starting composition (56).

33. G. Carlier et al., Geology 33, 601 (2005).
59. We thank D. Foster for 40Ar/39Ar analyses and R. Allmendinger and P. Molnar for suggestions that improved the paper.

Supporting Online Material
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Materials and Methods
SOM Text
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References
27. 2007; accepted 18 March 2008 10.1126/science.1148615
www.sciencemag.org  SCIENCE VOL 320 6 JUNE 2008
ERRATUM

Post date 5 September 2008

Reviews: “Rise of the Andes” by C. N. Garzione et al. (6 June, p. 1304). A minus sign was missing from an equation in the third column on p. 1305. The correct equation should read “\(h = -472.5 \delta^{18}O_{\text{rainfall}} - 2645\).”