Uplift of the western margin of the Andean plateau revealed from canyon incision history, southern Peru

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ABSTRACT

We explore canyon incision history of the western margin of the Andean (Altiplano-Puna) plateau in the central Andes as a proxy for surface uplift. (U-Th)/He apatite data show rapid cooling beginning at ca. 9 Ma and continuing to ca. 5.1 Ma in response to incision. A minimum of 1.0 km of incision took place during that interval. The youngest apatite date and a volcanic flow perched 125 m above the present valley floor dated at 2.261 ± 0.046 Ma (^{40}Ar/^{39}Ar) show that an additional ~1.4 km of incision occurred between ca. 5.1 and 2.3 Ma. Thus, we infer that a total of at least 2.4 km, or 75% of the present canyon depth was incised after ca. 9 Ma. (U-Th)/He zircon data collected along the same transect imply that the western margin of the plateau was warped upward into its present monoclinal form, rather than uplift being accommodated on major surface-breaking faults.

Keywords: Altiplano, geochronology, tectonics, Peru, geomorphology, helium.

INTRODUCTION

Understanding the development of the Central Andean plateau is crucial to evolutionary models of both Andean geodynamics and regional climate patterns. (Note: We use "Central Andean plateau" as defined by Allmendinger et al., 1997, to represent the region above the 3 km elevation contour between 13°S and 27°S. This includes the Altiplano and Puna plateaus and portions of the Western and Eastern Cordilleras.) Although episodes of central Andean deformation are constrained in many regions, it is often difficult or impossible to discern the magnitude of plateau uplift based on deformation history alone. Different approaches for estimating paleoelevation or the existence of high topography have led to a broad range of proposed uplift histories, but precise constraints are lacking. Oligocene uplift probably generated less than half of the central Andean relief seen today (e.g., Gubbels et al., 1993; Kennan, 2000). Numerous lines of evidence from the plateau and its eastern margin point to additional surface uplift starting at ca. 10 Ma, with magnitudes ranging from at least 1 km to as much as 3.5 km (e.g., Kennan et al., 1997; Lamb and Hoke, 1997; Barke and Lamb, 2006; Garzione et al., 2006; Ghosh et al., 2006). In northern Chile, Hoke (2006) and Nestor et al. (2006) estimate 1–1.4 km of western margin uplift after 10 Ma, and Wörner et al. (2000) argue for termination of uplift by 2.7 Ma. More recently, interpretations of oxygen and clumped isotope data have led to estimates that the plateau reached its present height by ca. 6 Ma (Garzione et al., 2006; Ghosh et al., 2006). Uncertainties in all these estimates require that independent measures of surface uplift be made before we can accurately constrain uplift history.

In southwestern Peru, large rivers cut deep canyons through the western margin of the Central Andean plateau. Cotahuasi-Ocoña Canyon is the deepest of these, incising more than 3 km below the plateau surface (Fig. 1). The western margin of the plateau has been characterized by a semiarid-hyperarid climate for at least the past 15 m.y. (e.g., Hartley, 2003; Rech et al., 2006). As expected, given the arid climate and gentle slopes of the region, Kober et al. (2007) found that millennial-scale interfluve erosion rates are uniformly low, despite a gradual increase in erosion rate with elevation (<0.001–0.05 mm/yr). As predicted from these studies, and as seen in the field, little surface erosion has occurred on canyon interfluves in our study area, making canyon incision an excellent proxy for surface uplift. Incision depth provides a minimum estimate for the amount of surface uplift, while the onset of incision provides a minimum age for the timing of uplift. We use ^{40}Ar/^{39}Ar dates of volcanic flows and bedrock (U-Th)/He thermochronologic data to explore canyon evolution. Detailed information regarding geochronologic data and their interpretation are provided in the GSA Data Repository[1] (Tables DR1–DR4).

VALLEY-FILLING VOLCANICS

Dates of valley-filling volcanic flows give a minimum time by which the canyon had incised to the depth of the sampled flow. In the middle reaches of Cotahuasi-Ocoña Canyon, a whole-rock ^{40}Ar/^{39}Ar date of 2.261 ± 0.046 Ma for a basaltic andesite flow sampled 125 m above the present valley floor (05TS38, Fig. 1) shows that ~96% of the incision in that section of the canyon (3.2 km total local incision) happened before 2.3 Ma. Farther upvalley, an ignimbrite (05TS25) perched ~400 m above the present valley shows that ~75% of the canyon depth (1.6 km total local incision) was cut before 3.825 ± 0.016 Ma. Thouret et al. (2005) obtained a ^{40}Ar/^{39}Ar date of 3.76 ± 0.14 Ma for the same flow and interpreted it as a maximum age for most of the valley incision.

LOW-TEMPERATURE THERMOCHRONOLOGY

In this setting where background erosion rates are very slow, we expect thermochronometers to yield very old dates. However, canyons are sites of localized, potentially rapid exhumation. When a canyon is incised, perturbations to near-surface isotherms result in rapid cooling of bedrock below the canyon bottom (Fig. 2). Because the closure temperature isotherm for the apatite (U-Th)/He system is only ~2–3 km below the surface (assuming a geothermal gradient of 20–30 °C/km, Farley, 2000), we expect to find young, rapidly cooled apatites below this level in the deepest reaches of Cotahuasi-Ocoña Canyon. Their young ages should reflect both downward movement of the closure isotherm and relative upward advection of rock due to erosion of the surrounding surface, though the latter component is small in this setting as discussed above. The oldest date among this young suite of ages should reflect the initiation of the thermal response to canyon incision.

We collected most samples for apatite (U-Th)/He dating along a 75 km, canyon-bottom transect, far from the influence of volcanic activity (Figs. 1 and DR1). Additional samples were collected from the upper-

[1]GSA Data Repository item 2007124, Figure DR1 (map of volcanic units and generation of paleosurface), Table DR1 (apatite-He data), Table DR2 (zircon-He data), Table DR3 (^{40}Ar/^{39}Ar data), Table DR4 (^{40}Ar/^{39}Ar summary), and Table DR5 (thermal model parameters and results), is available online at www.geosociety.org/pubs/ft2007.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

most catchment where there are abundant volcanic flows dated at ca. 1.4–3.8 Ma (Thouret et al., 2005). We constrained canyon depth by measuring the distance below a paleosurface that is preserved today as bedrock remnants beneath the regionally blanketing 14–16 Ma Huaylillas Ignimbrite (Thouret et al., 2005, and dates reported here). Positions of remnants exposed in valley walls were digitized and a spline surface warped to fit the points, giving an approximation for the bedrock surface (Fig. DR1). Samples were collected from ~1 km below the paleosurface near the coast, down to ~3.1 km below in the deepest reaches of the canyon.

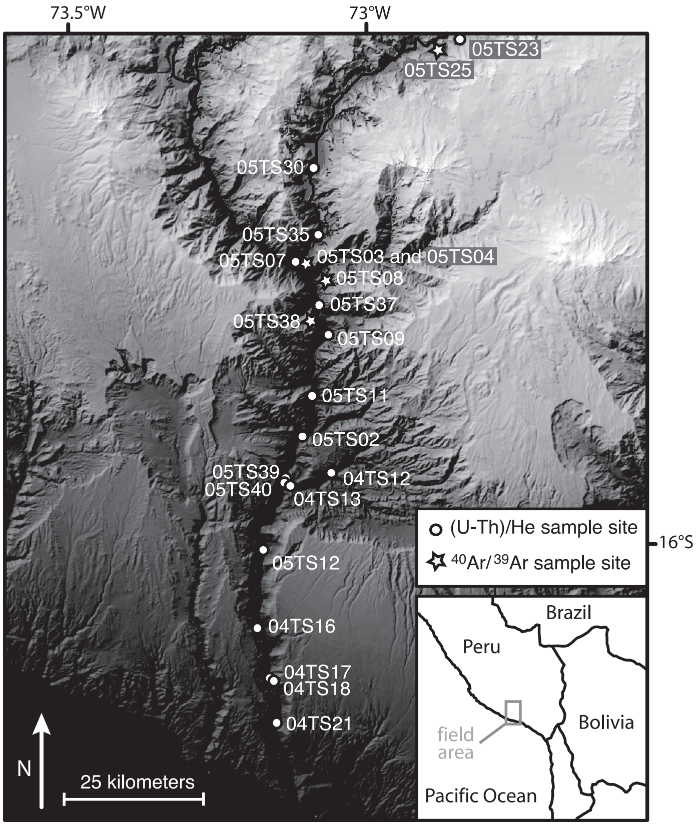

Figure 1. Location map and sample sites. Topography is shown as a gray-scale digital elevation model (DEM) draped over shaded relief from 90-m-resolution Shuttle Radar Topography Mission (SRTM) data. Inset shows location relative to South American geography. (U-Th)/He samples are collected from near river level. $^{40}Ar/^{39}Ar$ samples are from elevations noted in GSA Data Repository (see footnote 1).

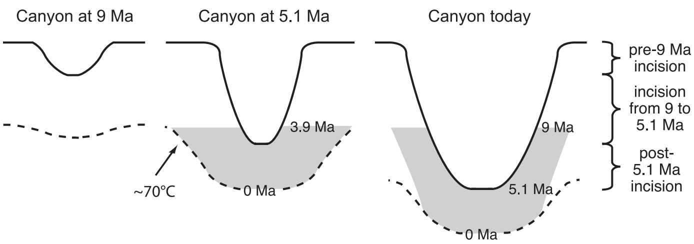

Figure 2. Schematic interpretation of incision history based on (U-Th)/He apatite data. Gray zones are rapidly cooled due to migration of the closure temperature isotherm (~70 °C) in response to incision. At least 1.0 km of incision occurred between 9 and 5.1 Ma, since this is the thickness of the rapidly cooled zone between those ages on the age-depth plot (Fig. 4). Post–5.1 Ma incision is equal to the depth of the closure temperature isotherm below the valley bottom at 5.1 Ma.

The resulting dates are ca. 60 Ma near the coast and rapidly decrease to ca. 9 Ma in the middle reaches of the canyon. Farther upvalley, there is a much more gradual change in dates with distance, with the youngest sample away from the volcanic-dominated upper catchment dated at 5.1 Ma (Fig. 3). Samples from the uppermost catchment (05TS07 and samples upstream from there) range from <1 to 3.8 Ma. Because these samples are as young as the abundant volcanic flows in that region, we suspect their ages have been reset or partially reset by recent volcanism. For this reason, we focus the remainder of our discussion on samples from the middle and lower catchment, which we believe have not been affected by reheating.

A plot of canyon depth versus sample cooling age (Fig. 4) can be interpreted the way that vertical profiles of thermochronologic data are typically interpreted: as indicative of how exhumation rates changed over the time interval represented by the measured ages. Although such an interpretation of Figure 4 may be compromised by variations in bedrock thermal structure, there is very little evidence of large lateral variations in upper-crustal temperatures in this region of southern Peru. For example, reported variations in measured surface heat flow range only from 32 to 44 mW/m^2 (Henry and Pollack, 1988; Hamza and Muñoz, 1996; Springer and Förster, 1998).

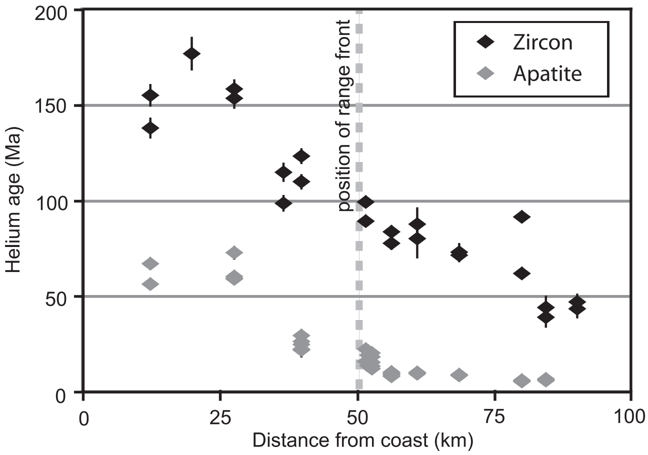

Figure 3. Zircon and apatite results plotted as distance measured perpendicular to coastline. Errors in age show 2σ uncertainty.

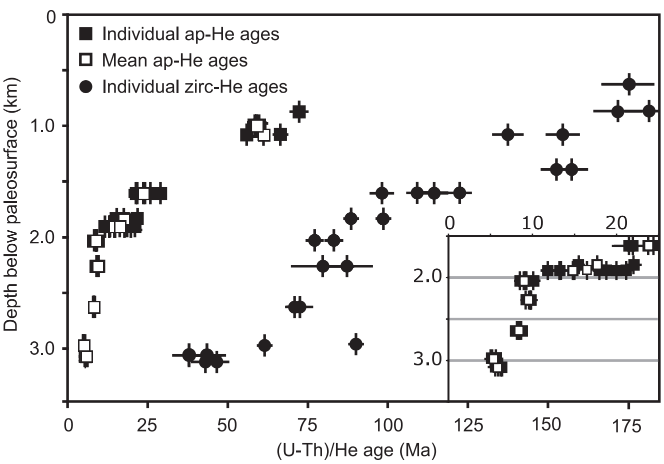

Figure 4. Apatite and zircon (U-Th)/He ages collected along valley at river level. Mean ages are weighted by the inverse of the variance of individual crystal ages. Errors in age are plotted as 2σ analytical errors only. Errors in depth below paleosurface are estimated at ±100 m. Inset shows apatite data over the last 25 m.y.

Ignoring the minor influences of such variations in heat flow, our apatite data imply slow background erosion of only ~0.7 km from ca. 60 to 9 Ma (0.01 km/m.y.) and a change to rapid incision of at least 1.0 km from ca. 9 to 5.1 Ma (0.26 km/m.y.). The latter is a minimum estimate because isotherm movement is damped compared to changes in surface topography. Additional incision must have occurred after 5.1 Ma to exhume this youngest apatite. Because a 2.3 Ma volcanic flow is perched 125 m above the valley floor in the middle reaches of the canyon, we know that the young apatite must have been exhumed from its closure depth to the near surface between 5.1 and 2.3 Ma. We explored probable erosion rates and depths to the closure temperature for this scenario using M. Brandon's computer code AGE2EDOT (summarized in Ehlers et al., 2005). Assuming typical thermal properties of the crust (Table DR5) and an initially uncompressed geothermal gradient between 20 and 30 °C/km, exhumation of the youngest apatite requires incision between 0.7 and 0.5 km/m.y., an effective closure temperature of 72 °C, and corresponding depths of 2.0–1.4 km to the closure temperature. Because we expect even greater compression of isotherms below a canyon compared to the compression expected from simply increasing erosion rate over flat topography as AGE2EDOT assumes, we choose the lower end of this range as an estimate. Thus, we infer that ~1.4 km of material was removed in the middle reaches of the canyon between 5.1 and 2.3 Ma (average 0.5 mm/yr) to exhume the youngest apatite.

Collectively, the data suggest that the middle section of Cotahuasi-Ocoña Canyon was incised at least 1.0 km between ca. 9 and 5.1 Ma and ~1.4 km between 5.1 and 2.3 Ma. This represents incision of at least 75% of the total canyon depth since 9 Ma, and 44% since 5.1 Ma.

INTERPRETING SURFACE UPLIFT FROM THERMOCHRONOLOGIC DATA

Deciphering surface uplift from thermochronologic interpretations of canyon incision is complicated by several factors. First, incision can reflect a transient response to tectonically driven uplift and/or climatically driven changes in sediment supply and fluvial discharge, provided prior surface uplift had produced sufficient topographic relief. We interpret the latest phase of incision to be a response to uplift because it is the simplest explanation, and there is no evidence for a climate change that could induce a pulse of incision at this time. The interpreted drying of climate starting at ca. 15 Ma (e.g., Hartley, 2003; Rech et al., 2006) should have the opposite effect of deterring incision, rather than generating a pulse of incision into pre-existing topography. Also, our estimate for the onset of incision correlates well with tectonic evidence in northern Chile for surface uplift after ca. 10 Ma (Nestor et al., 2006). Second, although incision magnitude is limited by uplift magnitude, incision rates are not necessarily closely tied to uplift rates. The incision history we interpret from thermochronologic data is offset from tectonically driven uplift by lag times defined by both the geomorphic response time to uplift and the thermal response time to incision. Constraining these response times is critical in order to quantitatively extract uplift history from canyon incision.

By "geomorphic response time," we mean the time scale over which rivers respond to a pulse of uplift. When uplift occurs, a river steepens at its outlet. Over time, this steep segment migrates upstream at a rate that depends principally on whether incision is transport- or detachment-limited (Whipple and Tucker, 2002). In a detachment-limited system, the pulse of incision will move through the system vertically at a rate that approximates the new uplift rate (Niemann et al., 2001), which appears to be ~0.5 mm/yr in Cotahuasi-Ocoña Canyon. In transport-limited systems, the incision pulse migrates upstream much more rapidly, resulting in essentially uniform onset of incision throughout the catchment on geologic time scales (Whipple and Tucker, 2002). Because we sampled from the trunk stream at large drainage area and thus both relatively low in elevation and most likely to approach transport-limited conditions, we expect incision across the sampled region to have a relatively short response time. Even if incision were detachment-limited, the highest bedrock sampled for apatite (852 m) would imply a maximum lag time of 2 m.y..

We investigated the thermal response time to incision using M. Brandon's RESPTIME computer code (summarized in Ehlers et al., 2005). Although this code was written to evaluate thermal field transients resulting from changes in broad-scale erosion rates, it is also useful for exploring general effects of localized incision. We tested the thermal response time to a sudden increase in erosion rate from 0.01 mm/yr to 0.5 mm/yr. Using typical values for thermal properties of the crust (Table DR5), results show that after an initial slow response, the migration rate of the apatite-He closure temperature isotherm reaches 75% of the incision rate 0.7 m.y. after incision starts. Thus, isotherm migration is significantly slower than incision for the first ~1 m.y.

Response time estimates show two ways in which our thermochronologic data do not directly reflect uplift. First, the age of the "kink" in the age-depth profile (Fig. 4) is likely to lag behind onset of uplift primarily as a result of geomorphic response time, which we expect to be a maximum of 2 m.y. Second, the damped response of isotherm movement in response to incision means that we derive only a minimum estimate of uplift rate and magnitude. Given these limitations, our estimate for the onset of uplift is likely good to ~2 m.y., and our estimate for uplift magnitude should be a robust minimum.

STYLE OF TECTONIC DEFORMATION

Compared to apatite, the higher closure temperature to helium diffusion in zircon (~180 °C, Reiners et al., 2004) means that (U-Th)/He zircon ages are set much deeper in the crust (~7 km) where they are not significantly affected by localized, near-surface thermal effects of canyon incision. The pattern of zircon cooling ages should therefore reflect only regional exhumation and postclosure bedrock deformation. Zircon dates of samples collected along the same valley-bottom transect show a regular but scattered progression from older dates near the coast to younger dates farther upvalley (Fig. 3).

If the latest pulse of uplift recorded by canyon incision were accommodated through localized, surface-breaking fault movement, we would expect to see a step-change in the pattern of zircon cooling ages (Fig. 5). Although we see no evidence for a large step, the present data set cannot rule out the possibility of uplift accommodated on a series of small faults distributed over a wide region. We have not seen distributed faulting in the field, but we mapped a focused shear zone at the range front. A 16.12 ± 0.04 Ma ($^{40}Ar/^{39}Ar$) undeformed ignimbrite crossing what appears to be the equivalent shear zone 100 km to the southeast of the field area suggests that significant movement on this structure predates the inferred ca. 9 Ma uplift. These thermochronologic data lead us to interpret the latest pulse of uplift as resulting from monoclinal warping of the western margin, consistent with interpretations presented by Isacks (1988).

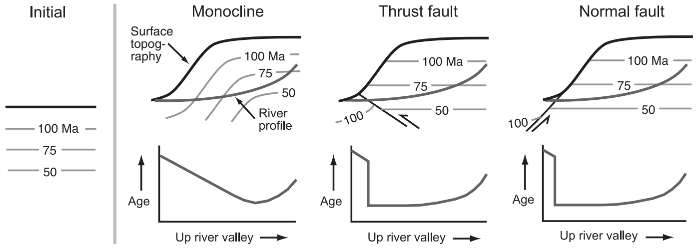

Figure 5. Predictions for patterns of (U-Th)/He zircon ages for samples collected along river profile in various tectonic settings, assuming initially flat isochrons. The pattern of ages presented in this study best matches the prediction resulting from a monocline.

GEODYNAMIC IMPLICATIONS AND CONCLUSIONS

One of the most important characteristics of the data presented here is the evidence they provide for onset of at least 2.4 km of river incision at ca. 9 Ma, in broad agreement with the magnitude and timing of recent uplift of the Central Andean plateau estimated by different methods. If, as we suggest, this incision corresponds to a phase of surface uplift, what can the age range tell us about plausible causes of uplift? Upper-crustal thickening from shortening at the plateau margins mostly occurred prior to 10 Ma (e.g., Gubbels et al., 1993; Victor et al., 2004; Farías et al., 2005; Elger et al., 2005). The argument for uplift through lithospheric delamination is well supported at the southern end of the plateau, where thinned lithosphere and widespread mafic volcanism point to delamination as a plausible cause (Kay and Kay, 1993; Schurr et al., 2006). In the north, however, strong evidence is lacking. Seismic data (Beck and Zandt, 2002) suggest there may be thinned lithosphere beneath the eastern portion of the plateau, but not under the western region. There is also no evidence for a volcanic flare-up close to the time of uplift (James and Sacks, 1999). Ignimbrite volcanism in southern Peru from ca. 16 to 14 Ma may be related to a small delamination event, although we expect canyon incision would not lag more than 2 m.y. behind significant uplift. Ductile thickening of the middle to lower crust provides a likely alternative, possibly triggered by displacement of material due to underthrusting of the Brazilian Shield beneath the eastern margin, as proposed by Isacks (1988). The ca. 10 Ma initiation of underthrusting (Gubbels et al., 1993) correlates well with the latest phase of uplift inferred from our data. For these reasons, although the data we present on uplift timing and monocline deformation are consistent with either lithospheric delamination or lateral flow of lower crust, we favor the latter hypothesis to explain this latest phase of uplift.

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