(U-Th)/He thermochronology records late Miocene accelerated cooling in the north-central Peruvian Andes

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ABSTRACT

The along-strike morphology of the South American Andes varies significantly, but it remains unclear if the timing and magnitude of rock exhumation are similarly varied. We used low-temperature (U-Th)/He thermochronology to constrain the exhumation history of the Eastern Cordillera in the north-central Peruvian Andes (7°S–8.5°S). Nine zircon (U-Th)/He mean ages range from 28 ± 4.5 Ma to 281 ± 60 Ma, and seven apatite (U-Th)/He mean ages yield tightly clustered late Miocene ages of 6.4 ± 3.8 Ma to 10.7 ± 1.8 Ma. These results document slow rock cooling followed by abrupt, accelerated cooling in the mid–late Miocene. Model time-temperature histories show slow cooling throughout the late Paleozoic to early Miocene, followed by an increase in cooling initiating between 14 and 10 Ma and continuing to the present. This rock cooling signal is regionally synchronous along the strike of the range and occurs at a time of a shift to a more humid and erosive climate in the eastern Andes. We suggest that a mid–late Miocene climate shift, following or synchronous with topographic growth, was responsible for the acceleration of rock exhumation in the north-central Peruvian Andes.

INTRODUCTION

Orogenic belts in convergent margins are formed by subduction of lithosphere, building topography by crustal shortening and thickening. As deformation progresses, mountainous topography interacts with atmospheric processes at Earth’s surface producing complex feedbacks that can perturb the climate system and potentially spatially focus deformation, possibly influencing relief generation (e.g., Molnar and England, 1990; Horton, 1999; Wobus et al., 2003; Willett and Brandon, 2002; Reiners et al., 2003; Vannay et al., 2004; Roe, 2005; Finnegan et al., 2008; Whipple, 2009). The formation of the Andean orogen is intimately linked to long-lived subduction along the western margin of South America (James, 1971; Isacks, 1988; Gephart, 1994; Kley et al., 1999; Roperch et al., 2000); however, both tectonic forces and climate are potentially reflected in the morphology of the Andes (Isacks, 1988; Masek et al., 1994; Montgomery et al., 2001, Pelletier et al., 2010; Capitanio et al., 2011). In Peru, the morphology of the range varies along strike; the southern Peruvian Andes contain the northernmost portion of the Altiplano, a broad, internally drained, low-relief, high plateau, whereas to the north, the spines of the Eastern and Western Cordillera merge in a narrower configuration of high topography (Fig. 1). This transition is spatially coincident with the location of the subducted Nazca Ridge, the transition from normal subduction in the south to flat-slab subduction in the north, and the absence of active volcanic centers (Barazangi and Isacks, 1976; Leeman, 1983; Gutscher, 2002). Morphological differences along strike have been explained by both tectonic and climatic variations, such as the influence of inherited structure (e.g., Kley et al., 1999), subduction characteristics (e.g., Isacks, 1988), and climatological variations (e.g., Masek et al., 1994; Montgomery et al., 2001).

Most Andean studies investigating surface uplift have focused on the Central Andes and recognized that surface uplift of the Altiplano has occurred since 40 Ma (Allmendinger et al., 1997; Isacks, 1988; Sempere et al., 1990; Capitanio et al., 2011). While some have suggested protracted uplift since Eocene time (McQuarrie et al., 2005; Barnes and Ehlers, 2009), recent work supports a period of very rapid uplift between 10 and 4 Ma (Gregory-Wodzicki, 2000; Ghosh et al., 2006; Hoke et al., 2007; Garzione et al., 2008; Lamb, 2011). However, the data that constrain these models for surface uplift are mostly from the Central Andes, and it is not clear to what degree these ideas apply elsewhere in the Andean chain, for example, in the north-central Peruvian Andes, where the morphology, tectonics, and geology differ.

Early work in the Andes identified punctuated, intense phases of tectonism that were tied to major changes in plate rates and/or dynamics (Mégard, 1984; Isacks, 1988; Noblet et al., 1996). In this view, deformational events included late Oligocene–Miocene crustal deformation and thickening via shortening or lower-crustal flow, which uplifted the Altiplano, followed by phases of tectonism in the late Miocene to present, deforming the eastern Subandes (Mégard, 1987; Isacks 1988; Sempere et al., 1990). More recent work has advocated for an alternative model of continuous deformation, propagating eastward from Late Cretaceous to the present (Noblet et al., 1996; Hartley et al., 2000; McQuarrie et al., 2005). In the Central Andes, this view is well supported by magnitudes, timing, and rates of exhumation from thermochronologic data, shortening estimates based on cross-section reconstructions, and estimates of foreland basin propagation (e.g., Noblet et al., 1996; McQuarrie et al., 2005; Ege et al., 2007; Carrapa and DeCellettes, 2008; Barnes et al., 2008). However, throughout north-central Peru, testing of models of punctuated versus continuous deformation using data that can quantify shortening and exhumation is only beginning to be addressed (Espurt et al., 2008; Giovanni et al., 2010; Gautheron et al., 2013; Eude et al., 2015; Margirier et al., 2015).

Specifically, in the north-central Peruvian Andes (Fig. 2), the tectonic chronology has classically been described as punctuated pulses of compressive deformation separated by longer periods of quiescence or extension...
Figure 1. Morphostructural map of South America. Altiplano and Puna Plateaus are labeled, along with: WC—Western Cordillera; EC—Eastern Cordillera; SA—Subandes; SB—Santa Barbara thrust system; SP—Sierra Pampeanas thrust system (modified from Gregory-Wodzicki, 2000; Garver et al., 2005). Flat-slab subduction is shown in dashed lines (based on Leeman, 1983), and active volcanic centers are shown by triangles (from Barazangi and Isacks, 1976). Elevation above 1500 m is shown in gray.

(McKee and Noble, 1982; Mégard, 1984; Noble et al., 1990). Several studies suggest that an Eocene deformational event (Incaic phase) resulted in shortening and was accompanied by a phase of mountain building, rock uplift, and perhaps surface uplift (e.g., Mégard, 1984; Sebrier et al., 1988; Noble et al., 1990). Because the timing of deformation in these studies relied on disparate geochronologic ages and geologic relationships, the apparent punctuated tectonic history may be an artifact of an incomplete geologic record. Further, the estimates of rock and surface uplift are based on geochronologic data and structural studies and not on data that can directly quantify rates and timing of rock exhumation. Low-temperature thermochronometer systems, such as (U-Th)/He in apatite and zircon, offer information about the cooling of the shallow crust, recording tectonic and/or erosional exhumation (e.g., Stockli et al., 2000; Ehlers and Farley, 2003; Reiners and Brandon, 2006). While they do not directly measure surface uplift, thermochronometers may provide constraints on relief generation and topographic growth. Here, we constrain the timing and magnitude of rock cooling from two sites in the Eastern Cordillera and discuss the potential processes causing Mesozoic through Cenozoic rock cooling in the north-central Peruvian Andes.

TECTONIC AND GEOMORPHIC SETTING OF THE PERUVIAN ANDES

Long-lived subduction of the Nazca plate beneath the South America plate (e.g., James, 1971) has produced the classic, “Andean-style” magin, consisting of a subduction trench, forearc basin, magmatic arc, and foreland fold-and-thrust belt (e.g., Dewey and Bird, 1970). In the Peruvian Andes between ~5°S and 15°S, the structural zones from west to east (Figs. 1 and 2B) are (1) the Western Cordillera, including the Coastal Batholith, associated with Mesozoic magmatism, which intruded Precambrian metamorphic rocks and is overlain by Cenozoic continental sediments and volcanics, (2) the Eastern Cordillera, composed of folded and faulted Precambrian–Paleozoic low-grade metamorphic basement and minor intrusive rocks, and (3) the Subandean fold-and-thrust belt, involving Mesozoic through Cenozoic marine and continental sedimentary deposits (James, 1971; Cobbing and Pitcher, 1972; Mégard, 1984; Sebrier et al., 1988; Hermoza et al., 2005).

The classic view of deformation and orogenesis in this region is described by five to six compressive phases, based on stratigraphic relationships and limited geochronology (Mégard, 1978, 1984; Sebrier et al., 1988; Noble et al., 1990). The Peruvian phase began ca. 85–70 Ma, affecting Paleozoic rocks of the eastern side of the Western Cordillera, and was coeval with volcanism along the entire Andean arc (Mégard, 1978). During the Paleocene–Eocene Incaic phases (I and II), the deformation front propagated eastward, producing the Marañón fold-and-thrust belt. The Marañón fold-and-thrust belt is characterized by Paleozoic basement overlain by folded Mesozoic metasedimentary rocks. The Miocene–Pliocene Quechua phases primarily affected the Eastern Cordillera and Subandes zone and have traditionally been reported as three punctuated phases (Quechua I at ca. 19 Ma, Quechua II at ca. 10 Ma, and Quechua III at ca. 5 Ma; Mégard, 1984; Noble et al., 1990; Sebrier et al., 1988). Since the late Miocene, the majority of crustal deformation has been focused within and along the eastern margin of the Subandean zone (Hermoza et al., 2005; Espurt et al., 2011; Gautheron et al., 2013). The Rio Marañón incises into the Mesozoic–Paleozoic metasedimentary and plutonic rocks of the Marañón fold-and-thrust belt, the location of this study (Fig. 2).

Early workers in north-central Peru identified three regional-scale erosional surfaces. From oldest to youngest, these are the Puna, Valle, and Cañon surfaces. The low-relief Puna surface, now at 4200–4400 m above sea level (masl), is cut on the ca. 30–15 Ma Calipuy volcanics, lying unconformably on Cretaceous sediments deformed during the Eocene Incaic II phase of deformation (Hollister and Sirvas, 1978; Cobbing and Pitcher, 1972). These units, presumed to be deposited near sea level, were uplifted and eroded ca. 15 Ma, during and just after Quechua I deformation (Bowman, 1906; McLaughlin, 1924; Coney, 1971). Similarly, the Valle stage is characterized by ~2 km of river incision into the Puna surface before deposition of the Fortaleza volcanics beginning at ca. 6 Ma, just postdating the Quechua II phase of deformation (Myers, 1976). Finally, during the Cañon stage, up to 2–3 km of localized incision into both the Puna and Valle surfaces reflects a recent episode of relative base-level fall and/or surface uplift during the Pliocene–Holocene (Myers, 1976, 1980; Cobbing et al., 1981, 1997; Garver et al., 2005). A recent study using thermochronologic techniques in this region recorded exhumation in the Cordillera Negra at 15 Ma coincident with regional surface uplift, the onset of Nazca Ridge subduction, and eastward propagation of magmatism (Margirier et al., 2015).

Additional constraints on the surface uplift history of the Peruvian Andes come from southern Peru in the northern extent of the Altiplano. Studies of river canyon incision rates using low-temperature thermochronology suggest late Miocene through Pliocene incision along the plateau margins (Thouret et al., 2007; Schildgen et al., 2007, 2009; Gunnell et al., 2010; Lease and Ehlers, 2013). In north-central Peru, the morphology lacks the presence of a broad, low-relief plateau such as that characterizing the Central Andes, including southern Peru. Here, the mechanisms and timing of deformation, surface uplift, and rock exhumation, and whether
they are similar to those proposed in southern Peru and the Central Andes are open questions.

METHODS

Apatite and Zircon (U-Th)/He Thermochronology

Thermochronology offers quantitative information about how rocks in Earth’s crust cool, typically due to advection to the surface. In general, the (U-Th)/He isotopic system is based on the radiogenic production of helium by alpha decay. At high temperatures, $^4$He diffuses out of the mineral, but below some temperature, diffusion slows such that the mineral grain is essentially a closed system. The closure temperature is the temperature at the time corresponding to the mineral’s apparent cooling age (Dodson, 1973). Cooling ages are calculated from the measured ratios of $^4$He/$^3$He, and parent nuclides $^{238}$U/$^{233}$U, $^{236}$U/$^{233}$U, and $^{232}$Th/$^{229}$Th (Farley, 2002; Reiners et al., 2004), and these are corrected for alpha ejection based on mineral volume and shape (Hourigan et al., 2005). Mineral thermochronometers vary in closure temperature and, thus, sensitivity to changes in topographically induced thermal perturbations; the zircon U-Th/He system (ZHè; ~180 °C; Reiners et al., 2004) records midcrustal cooling, while the apatite U-Th/He system (AHè; ~60 °C; Farley, 2002) records upper-crustal cooling. The AHè system is sensitive to variations in topography and erosion rates (Braun, 2002; Reiners and Brandon, 2006), while the ZHè system records midcrustal thermal effects. The use of both apatite and zircon (U-Th)/He thermochronology allows us to obtain more robust time-averaged exhumation rates across a thicker section of crust. Indeed, by using two different mineral thermochronometers from a single surface sample, it is possible to calculate acceleration or deceleration of rock cooling rates over time (e.g., Reiners and Brandon, 2006). In order to derive internally consistent time-temperature histories, we used the HeFTy® (Ketcham, 2005) program to produce inverse models. We determined the best-fit time-temperature histories based on prior geological information such as depositional ages, new U-Pb zircon crystallization ages of igneous samples, and our AHè and ZHè data. The thermal histories were modeled using each sample’s weighted mean zircon and apatite grain ages, where $n$ = 3–7 grains, for both ZHè and AHè systems (see supplemental Figs. A5–A10 in the Data Repository).³

From north to south, our sample locations are (Fig. 2): (1) the Balsas region (~7°S; Fig. 2B), and (2) the Sihuas region (~8.5°S; Fig. 2B). Both localities are within the Rio Marañon corridor. At each location, we sampled a transect of 4–7 individual rock samples and analyzed these using apatite and zircon (U-Th)/He thermochronology. Our sampling strategy included sampling across the folded and faulted Mesozoic sedimentary and Precambrian basement that comprise the Marañon fold-and-thrust belt, as well as 1–3 km of relief.

U-Pb Zircon Geochronology

At our sample locations, outcropping plutonic rocks are mapped as Carboniferous–Permian to Tertiary (Wilson and Reyes, 1964; Wilson et

³GSA Data Repository Item 2015375, tables of detailed chemical and geometric properties of grains analyzed for (U-Th)/He thermochronology, U-Pb geochronology, age-elevation relationships of thermochronologic data, and parameters used for HeFTy models, is available at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org.
al., 1967; INGEMMET, 1995, 1998). We quantified the timing of pluton emplacement by using in situ laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) zircon geochronology to further constrain our thermal inverse models. We used standard mineral separation techniques on granitoid samples from the Balsas section (RM07-1, RM07-2, RM07-3, and RM07-6), and Sihuas section (RM07-P3 and RM07-P10). Analyses of 7–20 individual crystals per sample included 26 μm spot sizes and 30 s of baseline data collection, followed by 30 s of on-peak data collection. Ablated zircon material was carried from the ablation cell by helium gas, mixed with argon sample gas, and injected into an ICP stream and measured using an XR single-collector ICP-MS. We used Iolite, a free software available on the Igor Pro platform (Paton et al., 2011) for data reduction and downhole fractionation and common-Pb corrections (see supplemental Table A3 [see footnote 1]).

RESULTS

The weighted mean zircon and apatite (U-Th)/He cooling ages were calculated from three to seven individual grain ages per sample (Figs. 3A and 3C; Table 1). For samples that were likely partially reset (i.e., containing a spread of ages), we report the range of individual grain ages (Figs. 3B and 3D). AHe ages are shown in an age-elevation plot (Fig. 4). Selected igneous samples were also analyzed for zircon $^{208}$U-$^{206}$Pb ages and are reported as weighted mean ages from 7 to 20 individual grain analyses (Fig. 5; Table 1; Table A3 [see footnote 1]). Both sample localities yielded U-Pb zircon results suggesting Paleozoic-age plutonic rocks. ZHe grain ages range from Mesozoic to early Cenozoic. All AHe grain ages in this study are tightly clustered from 10 to 6 Ma. Neither ZHe nor AHe ages are strongly correlated with effective U concentration (eU). Details of the geometry and chemistry of individual grains are presented in the supplemental data (Table A1; Figs. A3–A4 [see footnote 1]).

Balsas

The Balsas transect spans 900 vertical meters, consisting of four samples from granitoid units mapped as late Paleozoic (Wilson et al., 1967; INGEMMET, 1998; Figs 3A and 4). ZHe ages are scattered from Paleozoic to Mesozoic. In some samples, the ZHe samples exhibit a spread of individual grain ages (Fig. 3B), resulting in large errors of the mean weighted grain age (Fig. 3A). AHe ages are mid–late Miocene, ranging from 10.7 ± 1.8 Ma to 8.1 ± 1.2 Ma. RM07-1, RM07-2, RM07-3, and RM07-6 were also analyzed for zircon $^{208}$U-$^{206}$Pb ages, yielding mean weighted grain ages of 335.2 ± 2.3 Ma, 329.3 ± 6.2 Ma, 356.5 ± 6.1 Ma, and 329.1 ± 8.2 Ma, respectively (Table 1; Fig. 5).

Sihuas

In the Sihuas transect, located ~200 km south of Balsas, we collected six samples spanning 1800 m of elevation (Fig. 4). Sample lithologies vary from Precambrian meta-igneous rocks of the Marañon complex in the east, to folded and faulted Mesozoic sedimentary rocks in the west (Wilson et al., 1967; INGEMMET, 1995; Fig. 3B). Four of the five ZHe mean grain ages range from Jurassic to Oligocene in age (Figs. 3B and 3D). Individual grain ages scatter in some samples (i.e., RM07-P6, RM07-P13). In contrast, all AHe mean grain ages are tightly clustered in late Miocene time, ranging from 9.5 ± 2.3 Ma to 6.4 ± 3.8 Ma. Two samples from plutonic rocks, RM07-10 in the west, and RM07-P3 in the east, yield zircon $^{208}$U-$^{206}$Pb ages of 44.7 ± 0.5 Ma and 407 ± 12 Ma, respectively (Fig. 5).

DISCUSSION

Cooling Patterns across the North-Central Peruvian Andes

Our data document slow cooling throughout the Mesozoic and a mid–late Miocene to present period of accelerated cooling. In order to determine the cooling history at the sample localities, we ran model inversions using HeFTy with additional constraints from field relations and our new U-Pb zircon geochronology (Fig. 5). In plutonic units, we used zircon $^{208}$Pb Pb crystallization ages reported in this study, indicated by solid line boxes on Figure 5. Further, where sedimentary sections nonconformably overlie these plutonic rocks, we included the depositional age range as a constraint, plotted as dashed boxes on Figure 5. These field relations are based on stratigraphic sections and geologic maps by Benavides-Cáceres (1956), Wilson and Reyes (1964), Wilson et al. (1967), Janjou et al. (1981), INGEMMET (1995, 1998), and Scherrenberg et al. (2012). Models from Balsas are shown in Figures 5A–5D, and those from Sihuas are shown in Figures 5E–5F. The dark-gray envelopes encompass the model paths that pass a “good” fit (goodness of fit [GOF] >0.05), and the light-gray envelopes show paths that are “acceptable” (GOF >0.5). In each model, we have expanded the late Cenozoic time-temperature history as an inset where the dashed lines represent the older and younger bounds of modeled accelerated cooling within the range of “good” paths.

At the Balsas site, four samples come from Carboniferous plutonic rocks (Table 1; Figs. 5A–5D). The U-Pb zircon ages of these plutonic units presented here are consistent with previous studies of zircon U-Pb ages from the Marañon corridor ranging from 320 to 309 Ma (Mišković et al., 2009). Thus, the ZHe age spectrum, 280 ± 60 Ma, 213 ± 20 Ma, 137 ± 80 Ma and 101 ± 34 Ma, does not reflect crystallization ages. The scatter in the ZHe ages (Fig. 3B) suggests that these samples spent an extended time in the partial retention zone (PRZ). The age of the Mitu Group is constrained to Permian–Triassic by marine fossils (Dunbar and Newell, 1946), stratigraphic relationships (Laubacher, 1978), and K-Ar geochronology (Kontak et al., 1990). In the Balsas region, the Mitu Group (Ps-m) nonconformably overlies the Carboniferous plutonic units (Wilson et al., 1967; INGEMMET, 1995). These structural, stratigraphic, and chronologic relationships allow us to estimate pluton unroofing during ca. 300–270 Ma. The Mesozoic section in this region thins to the east. Throughout central Peru, the Mesozoic stratigraphic section above the late Paleozoic erosional nonconformity varies from ~2.5 to ~4.5 km thick (Benavides-Cáceres, 1956; Wilson and Reyes, 1964; Wilson et al., 1967; Scherrenberg et al., 2012). To avoid forcing the thermal models by constraining the paleothermometric gradients or making assumptions about the detailed stratigraphic thicknesses, we defined a broadly acceptable burial history (indicated by dashed line boxes; Figs. 5A–5D) and let the inversion determine the best-fit paths. In our time-temperature models, we assumed that the Carboniferous plutonic rocks were buried deepest at the end of the Mesozoic, and we loosely constrained the range of possible burial temperatures in the Late Cretaceous from 60 °C to 300 °C. All four inverse models from the Balsas samples RM07-1, RM07-2, RM07-3, and RM07-6 yield similar thermal histories. After pluton emplacement, the samples cool on the order of ~12 °C/m.y, as they are unroofed to the surface in the Permian and then buried throughout the Mesozoic and exhibit slow cooling in the early-to-mid-Cenozoic. In the late Miocene, the area experienced rapid cooling to the present (Figs. 5A–5D). The onset of rapid cooling is slightly older for RM07-1 (Fig. 5A) and RM07-6 (Fig. 5D), compared to RM07-2 and RM07-3 (Figs. 5B and 5C). If we assume that the crustal cooling signal recorded was broadly regional, we expect that samples at higher modern elevations would have cooled through the AHe isotherm first. RM07-6 (Fig. 5D, inset), the highest-elevation sample,
Late Miocene accelerated cooling in the north-central Peruvian Andes

**TABLE 1.** (U-Th)/He APATITE AND ZIRCON AND U-Pb ZIRCON AGES

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<th>Site</th>
<th>Latitude (°S)</th>
<th>Longitude (°W)</th>
<th>Sample name</th>
<th>Lithology</th>
<th>Elevation (masl)</th>
<th>Mean ZHe age (Ma)</th>
<th>Mean AHe age (Ma)</th>
<th>Mean 238U-206Pb age (Ma)</th>
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Note: Mean ages are mean-weighted from 3–7 grains for He analyses, and 7–20 grains for U-Pb analyses. Errors reported are 2σ. Abbreviations: masl—m above sea level; ZHe—zircon (U-Th)/He thermochronology; AHe—apatite (U-Th)/He thermochronology.
Figure 4. Mean-weighted apatite (U-Th)/He ages with 2σ errors from (A) Balsas section and (B) the Sihuas section plotted against sample elevations in meters. Each mean age is calculated from three to seven individual ages. Individual ages are shown in Figures 3B and 3D. Elevations were determined using a handheld global positioning system (GPS) unit.

Figure 5. Time-temperature (T-t) model results for selected samples using HeFTy (Ketcham, 2005). Goodness-of-fit T-t paths are shown in dark gray (GOF > 0.05) and light gray (GOF > 0.5). Black solid line boxes show 238U-206Pb zircon geochronologic ages from this study, and dashed line boxes show additional time-temperature constraints based on burial history and depositional age. We show the late Cenozoic time-temperature history as an inset in each model, where dashed lines represent the older and younger bounds of modeled accelerated cooling, and we report this time range as text within the inset. Abbreviation: m.a.s.l.—m above sea level.
shows an older range of good fits for the onset of mid–late Miocene cooling, from 18 to 12 Ma. Samples RM07-2 and RM07-3, at elevations of 1135 and 1162 masl, respectively, yield younger ranges for the onset of rapid cooling, from 11.5 to 10 Ma, consistent with the idea that higher elevations cooled through the apatite isomorph first. RM07-1 is the lowest elevation sample at 900 masl, with an onset of rapid cooling estimated from 14 to 12.5 Ma. When considering “acceptable” paths (GOF > 0.5) shown in light gray, the onset of rapid cooling extends to a younger bound of ca. 7 Ma for RM07-1 (Fig. 5A, inset). In summary, from good paths, all four thermal models from Balsas demonstrate accelerated cooling initiated between 14 and 10 Ma.

In the Sihuas transect, 200 km south of the Balsas site, we see a similar cooling pattern from ZHe and AHe chronometers: a period of slow Mesozoic cooling and rapid late Miocene cooling, respectively. However, unlike the Balsas section, the sample locations are distributed over a greater horizontal distance, and samples vary in lithology, including Cretaceous sedimentary units, Precambrian basement, and an Eocene pluton. The folded and faulted Mesozoic sedimentary units that crop out in the western section of our map lie in fault contact with Precambrian basement in the east (Fig. 3C). Four samples, RM07-P13, RM07-P10, RM07-P6, and RM07-P5, are from an upper thrust sheet, and three samples, RM07-P3, RM07-P2, and RM07-P1, are from Precambrian basement in a structurally lower thrust sheet to the east. Three of the Sihuas sample locations were collected from sedimentary rocks; thus, it is possible that the age scatter within samples RM07-P5, RM07-P6, and RM07-P13 is due to nonreset detrital zircons within the metasedimentary units. However, Michalak (2013) has shown that the provenance for the zircons in these Cretaceous sedimentary units is Permian or older, indicating the samples were at least partially reset, requiring burial depth to at least the ZHe PRZ (cf. Scherrenberg et al., 2012).

For sample RM07-P3 from Sihuas, we assumed that the broader region experienced a similar geologic history and used the same unroofing and burial constraints as for the Balsas basement rocks (Fig. 5E). Accelerated cooling initiates between 14 and 10 Ma (Fig. 5E, inset). The time-temperature model for RM07-P10 is constrained by a U-Pb age of 44.7 ± 0.5 Ma (Fig. 5F). Unlike all other samples modeled in this study, RM07-P10 displays fast cooling (~30 °C/m.y.) in the early Cenozoic. We attribute this signal to localized postemplacement cooling. While there are no known plutons of this age range (ca. 46 Ma) in the Eastern Cordillera of Peru, west of the study site, in the Coastal Batholith, Eocene ages (49.7 Ma) have been measured in the Santa Rosa Group (Mukasa, 1986). This may suggest that the extent of magmatism associated with the Coastal Batholith extends significantly farther east than previously recognized. However, aside from this period of early Cenozoic cooling, the cooling rate of RM07-P10 (3553 m) from mid–late Miocene to the present is similar to all the other models, with an acceleration of cooling bracketed between 12 and 7.5 Ma (Fig. 5F, inset). Compared to RM07-P3 (1864 m), where onset of accelerated cooling is 14–10 Ma, the range of onset of cooling for RM07-P10 is younger. At Sihuas, we see a slightly older range of good fits for onset of rapid cooling in the lower-elevation sample, RM07-P3 (Fig. 5E, inset). Despite differences in elevation, these two models are generally consistent with an onset of rapid cooling initiating between ca. 14 and 10 Ma.

While the U and Th (ppm) content (Table A1; Figs A3–A4 [see footnote 1]) of our mineral separates are within a range of values (~100–1000 ppm) where diffusion characteristics behave similarly (Reiners et al., 2004), if grains ranging in U and Th spend extended time in or near the PRZ (Wolf et al., 1998), the apparent ages will scatter due to differing He loss (Reiners and Farley, 2001; House et al., 2002; Shuster et al., 2006; Guenthner et al., 2013). Indeed, from some samples, we observe significant scatter of ZHe ages of individual grains (Figs. 3B and 3D). However, HeFTy (Ketcham, 2005) can take into account geometric, chemical, kinetic, and thermal parameters to determine a best-fit time-temperature path for a population of grains. In the data from the central Peruvian Andes we present here, all the thermal models demonstrate slow cooling near the PRZ (Fig. 5) throughout the Mesozoic and early Cenozoic.

In summary, the three-stage thermal histories derived from zircon U-Pb, ZHe, and AHe data suggest that Precambrian basement was unroofed by Permian–Triassic time and was buried near the ZHe closure isotherm in the Late Cretaceous coeval with deposition of Mesozoic sedimentary rocks of the Rio Marañon corridor (Fig. 6). With the exception of the localized cooling from the Eocene pluton, all HeFTy models show
slow cooling throughout the Mesozoic and early Cenozoic prior to a mid–late Miocene acceleration in cooling rate. The timing of onset of rapid cooling ranging from 14 to 10 Ma is congruous between the two sample sites. The slight variations in the range of onset of accelerated mid–late Miocene cooling can be explained by the elevation range of samples or are within error of paths of acceptable fit. Moreover, the similarity in cooling history across the sample sites suggests that the signal we document here was regional in scale and broadly synchronous.

Cenozoic Rock Exhumation and Orogenic Growth of the North-Central Peruvian Andes

Thermochronometers record the timing of rock cooling. Cooling may be generated by regional relaxation of thermal gradients, or by the advection of rock toward the surface (e.g., Ring et al., 1999; Reiners and Brandon, 2000). Thus, the late Cenozoic cooling pattern we measure in central Peru might be due to thermal perturbations resulting from postmagmatic cooling. Widespread Cenozoic magmatism is well documented in the Western Cordillera of Peru (Noble et al., 1974; Tosal et al., 1981; McKee and Noble, 1982), and the locus of magmatic activity is represented by the widespread Eocene–mid-Miocene Calipuy volcanics in the Western Cordillera between 7.5°S and 8.5°S (Cobbing and Pitcher, 1972; Hollister and Sirvas, 1978; Cobbing et al., 1997; Noble et al., 1999). Plutonism in the Western Cordillera was largely confined to the Coastal Batholith (~100 km to the west of the study region) with emplacement ages older than 35 Ma (Stewart et al., 1974; Cobbing et al., 1981; McKee and Noble, 1982; Mukasa, 1986). The only other known plutonic unit in the area is the late Miocene Cordillera Blanca Batholith (Gilette and Day, 1968; Landis and Rye, 1974; Stewart et al., 1974; McNulty and Farber, 2002; Giovanni et al., 2010). Garver et al. (2005) suggested that cooling determined from zircon and apatite fission-track data in the central Cordillera Huayhuash at ca. 11 Ma was due to postmagmatic cooling of the Cordillera Blanca Batholith. However, the plutons of the Cordillera Blanca are confined to intrude along the Cordillera Blanca detachment fault (McNulty and Farber, 2002) to the west and are not known to extend into the Rio Marañon corridor in the Eastern Cordillera. In summary, there is no geologic evidence suggesting that the Rio Marañon corridor experienced significant thermal perturbations during Mesozoic and Cenozoic time. Moreover, our geochronologic data largely show Paleozoic-aged intrusives, and ZHe data indicate that there was no regional heating past the ZHe closure isotherm after the late Paleozoic. Thus, based on the lack of evidence for significant Cenozoic thermal perturbations, we suggest that the mid–late Miocene cooling signal from our thermochronologic data records rock exhumation.

In north-central Peru, the two main belts of Cenozoic deformation are the Marañon and Subandean fold-and-thrust systems, formed during the Eocene and late Miocene, respectively (Jordan et al., 1983; Mégard, 1984; Horton and DeCelles, 1997; Hermoza et al., 2005). The Marañon fold-and-thrust belt formed during the Eocene Incaic orogeny, which has been considered the main contractional orogenic phase in the region (McKee and Noble, 1982; Mégard, 1984; Noble et al., 1990). Incaic structures are interpreted as reactivated normal faults that originally formed in response to Paleozoic rifting (Wilson and Reyes, 1964; Mégard, 1987). The timing of activity of Incaic structures in the Marañon fold-and-thrust belt is constrained by scattered geochronologic studies. Noble et al. (1990) showed that volcanics dated at ca. 42 Ma rest unconformably on Incaic structures, suggesting that shortening in the Marañon fold-and-thrust belt was complete by Eocene time (Noble et al., 1974, 1990). Paleocene–Eocene syntectonic sedimentary deposits in the Subandean retroarc foreland basin (i.e., Huallaga and Marañon Basins in north-central Peru) support the interpretation that at least some topographic relief existed in the area of the Marañon fold-and-thrust belt by the late Eocene, but these stratigraphic sections are not widely preserved. Our results show that the thermochronologic signal of rock cooling during this time frame is slow; if significant relief was generated during the Incaic orogeny, it is undetectable using the ZHe and AHe systems.

To determine exhumation rates, we used thermal constraints based on the HeFTy inverse models. Using a typical geothermal gradient (30 °C/Km) and the temperature of maximum burial depth indicated by our thermal models, ~180 °C (Figs. 5A–5E), we calculated a crustal thickness between the ZHe and AHe closure isotherms of 4 km. Thus, the maximum time-averaged exhumation rate from the Late Cretaceous (the time of maximum burial) to the mid–late Miocene (onset of rapid cooling) is ~0.08 mm/yr. The restriction of this section of crust to the zircon PRZ throughout the Mesozoic and early Cenozoic implies that Mesozoic and Upper Paleozoic sections exposed today were not buried to significant depths during the Incaic orogeny to warm above the zircon closure temperature.

Evidence for early Cenozoic relief generation in the north-central Peruvian Andes is found in the sedimentary record within the eastern flank of the Andes. We see the transition from marine sedimentary rocks (e.g., Late Cretaceous Celendin Formation; Benavides-Cáceres, 1956) to terrestrial sediments in the latest Cretaceous (e.g., latest Cretaceous Chota Formation; Benavides-Cáceres, 1956) and the preservation of these marine sedimentary rocks several kilometers above sea level today. Terrestrial sedimentation in the Marañon corridor continued through the Paleocene–Eocene, generating hundreds of meters of clastics preserved in localized basins (Hermoza et al., 2005; Espurt et al., 2011). These data suggest we might expect to see a signal of regional topography generation during the early Cenozoic. Balanced cross sections across the Marañon fold-and-thrust belt from recent structural studies suggest most crustal shortening in the Eastern Cordillera and Subandes has occurred since the mid–late Oligocene, in the range of 27%–41% (Mégard, 1984; Baby et al., 1997; Scherrenberg et al., 2014; Eude et al., 2015), providing some constraint on surface uplift. For a simple crustal block in isostatic balance, 30% crustal shortening would result in surface uplift on the order of 1–2 km over this entire period, assuming surface uplift and denudation are in steady state. Despite geological field evidence for crustal shortening and thickening in the early–mid Cenozoic, crustal shortening estimates based on balanced cross sections suggest that shortening during the Incaic phase was not enough to expect to see evidence of related surface uplift in our cooling histories (Eude et al., 2015). Further, these data suggest most of the surface uplift and relief generation in the Rio Marañon corridor were achieved from the late Oligocene to present day. In summary, in north-central Peru, we documented an extended period of slow cooling in early-to-mid-Cenozoic time during formation of the Marañon fold-and-thrust belt. Our results can be explained with current estimates of shortening, but they require that time-averaged exhumation be less than 0.08 mm/yr from Late Cretaceous to mid–late Miocene time.

Late Miocene Rock Cooling in the North-Central Peruvian Andes

While our ZHe results imply little surface denudation and erosion throughout the early-to-mid-Cenozoic, the AHe grain ages are all ca. 10 to 6 Ma. Six thermal inverse models are consistent with a rapid increase in cooling rate initiating between 14 and 10 Ma. Such a regionally synchronous signal of rock cooling requires a mechanism capable of broadly exhuming the entire eastern Peruvian Andes (7°S–8.5°S). Next, we discuss the processes that may explain this signal, such as oxygen-scale extension, reactivated thrusting (inducing regional exhumation), and/or focused surface denudation resulting from a regional shift to a wetter climate.
While there is evidence for late Miocene extension in central Peru, it is localized to a few small basins in the Western Cordillera of the Peruvian Andes and restricted primarily to the west of the Marañón fold-and-thrust belt (i.e., the Rio Santa, Cajamarca, San Marcos, and Namora Basins; Bonnot et al., 1988; Bellier et al., 1989; McNulty and Farber, 2002). Thus, the absence of any regional-scale Cenozoic extension in the Rio Marañón corridor suggests that the Miocene cooling was not related to regional crustal thinning. More plausible mechanisms for late Miocene cooling are (1) rock exhumation of the Eastern Cordillera from crustal thickening, and subsequent increased erosion due to steepened topography, and/or (2) increased surface denudation driven by a shift to a wetter regional climate.

Many studies have pointed out the importance of basement-involved thrusting in the Marañón fold-and-thrust belt and Subandes (Suárez et al., 1983; Dorbath et al., 1991; Espurt et al., 2008; Gautheron et al., 2013; Scherrenberg et al., 2012, 2014; Eude et al., 2015). Based on historical seismicity in the eastern Peruvian Andes, some authors have suggested active underthrusting of the Brazilian Shield beneath the eastern Andes, where the crust is brittle and thinner (Suárez et al., 1983; Dorbath et al., 1991). In the Ucayali and Camisea Basins, Espurt et al. (2008) and Gautheron et al. (2013) emphasized the role of reactivated Paleozoic structures in generating shortening and associated surface denudation recording a Miocene cooling signal. Scherrenberg et al. (2012, 2014) interpreted structural sections as being accommodated by thick-skinned, basement-involved thrusts. Recently, Eude et al. (2015) argued that mid-Miocene shortening in the Marañón fold-and-thrust belt was controlled by thick-skinned thrusts underlying thin-skinned structures, which resulted in steepening topography and increased erosion rates, recorded by Miocene apatite fission-track ages and AHe ages. Specifically, they suggested that thrust-wedge propagation favored growth via basement structures in north-central Peru. Indeed, Suárez et al. (1983) reported that earthquake data suggest that faults along the eastern flank of the Andes are characterized by reverse slip at depth.

The mechanism that caused reactivation of basement thrusts in mid-late Miocene time in north-central Peru could be linked to changes in plate rates or dynamics. In the early Miocene (ca. 25 Ma), the convergence rate between the Nazca–South American plates accelerated (Wortel and Cloetinig, 1981; Parodo-Casas and Molnar, 1987). These events were coincident with arc magmatism in the Western Cordillera (Jordan and Alonso, 1987), and structural, paleontological, and geologic evidence suggests a significant and protracted shift in tectonism throughout the Central Andes during the Miocene (e.g., Sempere et al., 1990; Lamb et al., 1997; Jordan et al., 2001). In the northern Andes of Ecuador, spatial and temporal patterns of exhumation are linked to changes in plate kinematics, and possibly subduction of oceanic ridges (Spikings et al., 2000, 2010). However, it is difficult to tie a causal link from changes in plate dynamics to the Miocene rock exhumation documented in this and other studies throughout north-central Peru (Gautheron et al., 2013; Eude et al., 2015).

Conversely, accelerated exhumation does not require shortening or deformation. For example, in the Eastern Cordillera of northern Bolivia, an onset of rapid exhumation 15–10 Ma is not associated with internal deformation and is instead attributed to erosional exhumation (Gillis et al., 2006; Barnes et al., 2012). Mora et al. (2014) noted the presence of passive roof duplex structures in the Colombian and Bolivian Subandes and suggested that their formation required erosion of the duplex crest and deposition at the front and was therefore influenced by climate. In Peru, Mosolf et al. (2011) proposed that rapid exhumation ca. 11 Ma was decoupled from upper-crustal shortening and was achieved via lower-crustal dynamics, or a climate change that increased the supply of wedge-top sediment, thereby loading the Subandes and promoting eastward orogenic growth. In the Sierra Pampeanas of Argentina, rock exhumation and surface uplift have in part been explained by climatic forcing from late Miocene orographic focusing of precipitation (Sobel and Strecker, 2003). In the Patagonian Andes, it is suggested that late Miocene rock cooling may be explained by changes in moisture patterns in the late Cenozoic (Fosdick et al., 2013).

**Relationships between Tectonics and Climate in the Andes**

Over the past two decades, there has been a large effort using both model-based and field-based studies in orogens around the world to identify the feedbacks in the dynamics of the tectonic-climate system (e.g., Koons 1990; Willett, 1999; Beaumont et al., 2001; Reiners et al., 2003; Whipple and Meade, 2004; Vannay et al., 2004; Finnegan et al., 2008; McQuarrie et al., 2008). Recent work has emphasized the role of orographic focusing of precipitation in perturbing the internal dynamics of orogens and has shown that surface processes may influence where deformation occurs, the width and the rate of wedge growth, and the rock exhumation rate (e.g., Dahlen and Suppe, 1988; Koons, 1990; Beaumont et al., 1992, 2001; Masek et al., 1994; Reiners et al., 2003; Thiede et al., 2004; Whipple and Meade, 2004; Roe and Baker, 2006). In this view, mountain belts are constructed by a coupling of tectonic and climatic processes, yet it remains challenging to deconvolve the relative influence of each on the orogenic system (e.g., Whipple, 2009). For example, in the north-central Peruvian Andes, the mid–late Miocene acceleration from slow to rapid cooling in the Marañón fold-and-thrust belt in the Eastern Cordillera of north-central Peru could be explained by a passive erosional response to a primary tectonic driver such as a change in plate rate and reactivation of basement-involved thrusts that steepen topography (Baby et al., 1997; Gautheron et al., 2013; Eude et al., 2015). Alternatively, the cooling signal could be caused by regional topographic growth sufficient to induce a change in regional climate patterns, such as a shift to a wetter climate, which may focus orographic rainfall and enhance rock exhumation, with or without internal deformation (Horton, 1999; Strecker et al., 2007; Barnes et al., 2012).

The mid–late Miocene accelerated rock cooling signal we find in the Marañón Basin is similar in timing to several other geologic changes in the Subandean and Amazon Basins. Thick deposits of late Miocene clastics in the Subandes (Hermoza et al., 2005) suggest increased erosion rates on the eastern flank of the Andes. Rapidly increased sedimentation rates in the Subandean foreland basin (Uba et al., 2007) and Amazon Basin (Figuereido et al., 2009) during the mid–late Miocene, the development of megafans in the foreland basin (Horton and DeCelles, 2001), and Amazonian flora and fauna diversification due to increased nutrient supply (Hoorn et al., 2010) all point to a wetter, more erasive climate in late Miocene time. These changes in depositional patterns and organization of the modern Amazon Basin system were also coincident with changes in stable isotopes recorded in the Subandean foreland deposits of the Chaco Basin of Bolivia, which indicate an increase in paleoprecipitation between 12 and 8 Ma (Mulch et al., 2010). Recent regional climate modeling indicates that the late Miocene was characterized by intensification of convective rainfall along the eastern flank of the north-central Andes, due to its steady rise (Poulsen et al., 2010; Insel et al., 2012). Finally, studies elsewhere in the Andes using thermochrometers suggest that patterns of rapid rock exhumation spatially and temporally correlate with patterns of greater precipitation (Gillis et al., 2006; Barnes et al., 2012). Together, these data imply that the mid–late Miocene regional climate was characterized by (1) more intense convective precipitation along the eastern margin of the Andes, (2) increased rock exhumation rates in the eastern Andes, and (3) higher sedimentation rates into the Subandes and Amazon Basin. These observations are consistent with the mid–late Miocene increase in rock exhumation we document in the Rio Marañón corridor of the Eastern Cordillera.
Late Miocene canyon incision in the southern Peruvian Andes is attributed to focused erosion in response to surface uplift (Thuoret et al., 2007; Schöndeggen et al., 2007, 2009; Gunnell et al., 2010). In the north-central Peruvian Andes, there are outcrops of at least three regional-scale incised paleosurfaces, the Puna, Vallé, and Cañon surfaces. These paleosurfaces were abandoned during the Miocene, late Miocene, and Pliocene–Pleistocene, respectively, and some workers have attempted to correlate these surfaces with the phases of deformation articulated in Peru (McLaughlin, 1924; Myers, 1976; Garver et al., 2005). Margirier et al. (2015) recorded accelerated rock exhumation in the Cordillera Negra of the Western Cordillera at 15 Ma coincident in time with incision into the Puna surface. Importantly, while the magnitude of incision into these surfaces was controlled by surface uplift (base-level fall), the timing of the incision into the surfaces might be more reasonably correlated with a change in climate resulting in increased river incision (Hall et al., 2008). Our data suggest rock cooling and exhumation initiating between 14 and 10 Ma, similar in timing to the incision and abandonment of paleosurfaces that have mainly been described in the Western Cordillera and western flank of the Andes of north-central Peru. Contemporaneous incision in spatially disparate localities throughout the north-central Peruvian Andes perhaps suggests a regional cause for the timing of the initiation of incision such as a shift to a more humid climate.

CONCLUSIONS

Throughout the Eastern Cordillera of north-central Peru, we have documented a signal of slow cooling in the early-to-mid-Cenozoic and an acceleration of cooling in the mid–late Miocene. Thermal modeling using HeFTy shows that cooling accelerated between 14 and 10 Ma and was regionally synchronous between the two study sites (Balsas, 7°S and Sihuas, 8.5°S). Zircon and apatite thermochronometers do not record a detectable signal from the effects of the major Eocene orogenic events (e.g., Mégard, 1984). We suggest that surface denudation rates remained low throughout most of the Mesozoic and early Cenozoic. However, several kilometers of rock uplift could occur in the Eastern Cordillera of the north-central Peruvian Andes, forming the Marañon fold-and-thrust belt, without a cooling signal that would be detectable using the AHe and ZHe thermochronometers. Importantly, we show that deformation, shortening, and mountain building may not be accompanied by increased cooling, particularly if surface denudation is low, precluding rock exhumation.

The regional synchronicity of the mid–late Miocene cooling signal documented here suggests a broad, widespread mechanism. This rock cooling of the Marañon fold-and-thrust belt occurred in the absence of postmagmatic cooling, oxygen-scale crustal extension, or major crustal deformation. It is possible that the cooling signal we document could have been driven by far-field tectonic forcing such as an increase in plate convergence rates (e.g., Wilson, 1996), reactivating Incaic structures (Mégard, 1984) and/or Paleozoic faults in thick-skinned style (Baby et al., 1997; Gautheron et al., 2013; Scherenberg et al., 2014; Eude et al., 2015). However, structural evidence suggests that the Miocene Quechua phase produced minor shortening in the Marañon fold-and-thrust belt and most late Miocene shortening occurred in the Subandes (Mégard, 1984; Scherenberg et al., 2014; Eude et al., 2015).

A change to a wetter, more erosive climate along the eastern flank of the Andes in the mid-to-late Miocene is suggested in studies of rock exhumation (e.g., Gillis et al., 2006; Barnes et al., 2012), regional climate modeling (e.g., Poulsen et al., 2010; Insel et al., 2012), stable isotopes (Mulch et al., 2010), sedimentation in the foreland basins (Horton and DeCelles, 2001; Uba et al., 2007; Figueiredo et al., 2009), and Amazonian species diversity (Hoorn et al., 2010). This regional shift in climate conditions was likely influenced by topographic growth of the Andes since the mid-late Miocene. Our results are consistent with the timing of a regional shift to a more humid climate, and we suggest that increased rates of surface denudation were most likely responsible for the mid–late Miocene increase in regional rock cooling in the north-central Peruvian Andes. Finally, while Eocene sedimentary thicknesses support the notion that minor surface uplift occurred during the Incaic orogeny, we suggest that the majority of surface uplift in the Marañon fold-and-thrust belt occurred from mid–late Miocene to present.

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Late Miocene accelerated cooling in the north-central Peruvian Andes


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(U-Th)/He thermochronology records late Miocene accelerated cooling in the north-central Peruvian Andes

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