Large along-strike variations in the onset of Subandean exhumation: Implications for Central Andean orogenic growth

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ABSTRACT

Plate tectonics drives mountain building in general, but the space-time pattern and style of deformation is influenced by how climate, geodynamics, and basement structure modify the orogenic wedge. Growth of the Subandean thrust belt, which lies at the boundary between the arid, high-elevation Central Andean Plateau and its humid, low-elevation eastern foreland, figures prominently into debates of orogenic wedge evolution. We integrate new apatite and zircon (U-Th)/He thermochronometer data with previously published apatite fission-track data from samples collected along four Subandean structural cross-sections in Bolivia between 15° and 20° S. We interpret cooling ages vs. structural depth to indicate the onset of Subandean exhumation and signify the forward propagation of deformation. We find that Subandean growth is diachronous south (11±3 Ma) vs. north (6±2 Ma) of the Bolivian orocline and that Subandean exhumation magnitudes vary by more than a factor of two. Similar north-south contrasts are present in foreland deposition, hinterland erosion, and paleoclimate; these observations both corroborate diachronous orogenic growth and illuminate potential propagation

mechanisms. Of particular interest is an abrupt shift to cooler, more arid conditions in the Altiplano hinterland that is diachronous in southern Bolivia (16-13 Ma) vs. northern Bolivia (10-7 Ma) and precedes the timing of Subandean propagation in each region. Others have interpreted the paleoclimate shift to reflect either rapid surface uplift due to lithosphere removal or an abrupt change in climate dynamics once orographic threshold elevations were exceeded. These mechanisms are not mutually exclusive and both would drive forward propagation of the orogenic wedge by augmenting the hinterland backstop, either through surface uplift or spatially variable erosion. In summary, we suggest that diachronous Subandean exhumation was driven by piecemeal hinterland uplift, orography, and the outward propagation of deformation.

1. INTRODUCTION

In the simplest sense, mountain belts grow when the addition of mass outpaces its removal by erosion (Montgomery et al., 2001; Willett and Brandon, 2002). Mountain belts have been described as critically tapered orogenic wedges which grow by lengthening and forward propagation of the deformation front (Davis et al., 1983). In this view, factors such as erosion and rock/fault strength determine whether mountain belts either propagate outward (i.e., towards the foreland) or deform internally to rebuild taper (Dahlen and Suppe, 1988; Willett et al., 1993). Even where such an overall framework is applicable, the timing of outward orogenic growth and its variation along strike are typically poorly known. These space-time data are needed to evaluate the mechanisms driving orogenic wedge growth and the length scales over which these processes are active.

The outward expansion of mountain belts can reflect changes in tectonics, geodynamics, and/or climate that occur during orogenesis. The degree to which the deformation front is localized by either surface processes (Barnes et al., 2012; Hirschmiller et al., 2014; Wobus et al., 2003) or deeper lithospheric processes (Clark, 2007; Decelles et al., 2009; Kay et al., 2005; Wells and Hoisch, 2008) is a topic of particular interest. For instance, the locus of orogenic deformation can advance or retreat because of factors like lithosphere density

(Molnar et al., 1993), orography (Willett, 1999), or subduction angle (Jordan et al., 1983). In the last decade, specific predictions of how the topography, width, erosion, and accretion of a mountain belt changes in response to tectonic and climatic forcing have been made using numerical and analytical models (Roe et al., 2008; Stolar et al., 2006; Whipple and Meade, 2006). However, few geologic examples exist to test these predictions (Whipple, 2009; Willett et al., 2006; Enkelmann et al., 2015a).

The development of orogenic plateaus like the Tibetan Plateau and the Central Andean Plateau represents a case where accretion of mass has outpaced erosion and high topography approaches a mechanical limit (Montgomery et al., 2001; Willett, 1999). Recent research on the Central Andean Plateau has attempted to quantify past changes in elevation (Bershaw et al., 2010; Canavan et al., 2014; Garzzone et al., 2014; 2008; 2006; Ghosh et al., 2006; Leier et al., 2013; Quade et al., 2015; Saylor and Horton, 2014), climate (Ehlers and Poulsen, 2009; Fiorella et al., 2015a; 2015b; Insel et al., 2010a; 2012; 2013; Jeffery et al., 2012; Poulsen et al., 2010; Rohrmann et al., 2014), and their interrelationship. Hinterland paleoelevation has proven a key parameter in evaluating geodynamic models of orogenic growth (Clark, 2007; England and Molnar, 1990) and their tectonic and climatic consequences. For example, an increase in hinterland elevation can both drive outward propagation of deformation (Decelles et al., 2009; Garzzone et al., 2006) as well as construct orographic barriers that focus rainfall at the mountain front (Mulch et al., 2010; Poulsen et al., 2010; Strecker et al., 2007) and lead to aridification of the hinterland (Pingel et al., 2014).

Proposed linkages between hinterland and foreland processes in Cordilleran orogenic systems like the Central Andes (Decelles et al., 2015; Garzzone et al., 2008; Kay et al., 2005; Strecker et al., 2007) are difficult to assess without timing constraints on mountain belt propagation (Barnes et al., 2012; Carrapa and Decelles, 2015). In this study we use low-temperature thermochronology to date the emergence of the frontal zone of the Central Andes: the Subandes. We then discuss the coeval hinterland processes that contributed to widening of the mountain belt.

2. GEOLOGICAL SETTING

Prolonged Nazca Plate subduction and upper plate deformation has produced the Central Andean orogen between 13 and 27°S (Allmendinger et al., 1997; Barnes and Ehlers, 2009; Isacks, 1988). Orogenic deformation extends 600-900 km eastward from the trench and has created a plateau of >3.5 km elevation and 60-75 km crustal thickness (Beck, 2002) that is flanked by forearc and retroarc thrust belts. Crustal shortening has been focused in the retroarc thrust belt since 50 Ma with >300 km of shortening (McQuarrie et al., 2005). In the region between 15° and 20°S (Fig. 1), shortening that initiated in the Eastern Cordillera (50-30 Ma) propagated eastward into the Interandean Zone (25-18 Ma), and finally, into the Subandean zone (15-0 Ma; Barnes et al., 2012; 2006; 2008; Decelles and Horton, 2003; Ege et al., 2007; Gillis et al., 2006; McQuarrie et al., 2008a).

The overall pattern of Cenozoic outward propagation of Andean deformation is well established, however less is known about the along-strike (north to south) pattern of deformation within any given zone of the orogen (Barnes and Ehlers, 2009). Despite similar plate convergence rates (Somoza, 1998), significant along-strike differences in Andean orogenic growth exist and have been interpreted to reflect along-strike variations in Andean climate, geodynamics, and geologic heritage (Barnes et al., 2012; Hilley and Coutand, 2010; Horton, 1999; Kley et al., 1999; Masek et al., 1994; McQuarrie et al., 2008b; Montgomery et al., 2001; Ramos, 2009). Even where along-strike variations are prominent, the timing of their influence on later deformation and erosion patterns is often poorly-constrained. A region of particular interest is the Subandean thrust belt, the youngest portion of the orogen (Brooks et al., 2011; Oncken et al., 2006) and an area where structural relationships and the foreland basin are preserved.

The Subandean tectonomorphic zone is characterized by active, generally thin-skinned deformation involving relatively shallow structural levels that has generated a basin and range morphology with an average elevation of ~1 km (McQuarrie, 2002; Kley, 1999). The sedimentary rocks involved in the Subandean thrust belt are a 2-10 km thick continuous

Paleozoic succession of Cambrian through Devonian marine siliciclastic rocks overlain by a 1-3 km thick discontinuous section of Carboniferous to Cretaceous nonmarine rocks (Roeder and Chamberlain, 1995; Sempere, 1994; 1995; Welsink et al., 1995). Tertiary synorogenic sediments up to 2-5 km in thickness are preserved in synclinal Subandean basins (Uba et al., 2009; Baby et al., 1995; Dunn et al., 1995).

Miocene to recent deformation and erosion in the Subandean thrust belt has been cited as evidence for various geodynamic (Garzzone et al., 2008) and climatic (Barnes et al., 2012; Uba et al., 2007) phenomena, yet few reliable exhumation chronologies exist. Many previous authors have extrapolated the ~9 Ma onset of Subandean deformation from growth strata in one well-studied locality in northern Argentina (Echavarria et al., 2003) to widespread areas 100s of kilometers to the north. These extrapolations do not take into account how prominent along-strike differences in structural heritage (Sheffels, 1995), sedimentary thicknesses (Allmendinger and Gubbels, 1996; McGroder et al., 2015), and climate (Bookhagen and Strecker, 2008; Garzzone et al., 2014) may have regulated the timing and style of Bolivian Subandean thrust belt growth. Testing tectonic-climatic-geodynamic linkages in the Subandes (Horton, 1999; McQuarrie et al., 2008b) has thus been limited by an inability to clearly define the onset of Subandean deformation and characterize its along-strike variation.

3. METHODS:

We quantify the timing and magnitude of exhumation of the Subandean thrust belt with low-temperature thermochronology. To resolve the timing of regional exhumation, we focus on the relationship between ages and structural depths determined from integrating multiple thermochronometers within balanced structural cross-sections. We identify those samples that were thermally reset because of sediment burial and record cooling that is due to Subandean thrust-induced erosion. Synchronous cooling across a range of structural depths provides a minimum age for when deformation had propagated into the zone and thrust sheet erosion began.

We collected samples from thrust fault hanging walls along structural cross-sections in four regions of Bolivia between 15° and 20°S (Figs. 1, A1-A2). All samples had well-defined structural/stratigraphic positions based on previous geologic mapping and balanced structural cross-sections in the Subandean thrust belt (Eichelberger et al., 2013; McQuarrie, 2002; McQuarrie et al., 2008a). In addition, cross-sections by others in these areas (Baby et al., 1992; 1995; Dunn et al., 1995; Roeder and Chamberlain, 1995) suggest similar sample depths and relationships as those reported here.

We focused on apatite (U-Th)/He thermochronology (AHe: closure temperature $T_c \sim 65^\circ\text{C}$; Farley, 2000; Shuster et al., 2006; Flowers et al., 2009) since previous apatite fission track thermochronology (AFT: $T_c \sim 110^\circ\text{C}$; Green et al., 1986; Ketcham et al., 2007) has largely failed to ascertain the timing of relatively shallow Subandean exhumation. We present results from 6 new AHe and 6 new zircon (U-Th)/He (ZHe: $T_c \sim 180^\circ\text{C}$; Reiners et al., 2002; Guenthner et al., 2013) thermochronometer samples from Paleozoic sedimentary rocks (Tables 1, S1-S2). Only 7 of the 35 Paleozoic quartzite samples we processed yielded apatite suitable for (U-Th)/He analysis, precluding analysis of a more comprehensive dataset. Furthermore, we compiled 25 previous, primarily AFT, ages (Table 1; Barnes et al., 2012; 2008; 2006; Eichelberger et al., 2013). Grain selection followed by Helium measurement on a Patterson Instruments gas extraction line equipped with a 960 nm diode laser was conducted at the University of Tübingen, Germany (Enkelmann et al., 2015b; Ehlers et al., 2015). Grain digestion and dissolution followed by Uranium, Thorium, and Samarium measurements on an Element 2 inductively coupled plasma mass spectrometer were performed at the University of Arizona, USA (Reiners, 2005). (see Supplementary Information for complete analytical details).

Samples are from Cambrian – Cretaceous strata and their thermochronometric ages can be classified as either reset, mixed reset, partially reset, or unreset by post depositional burial and heating (Barnes et al., 2006; Enkelmann et al., 2015b). Reset samples contain one age component that is younger than deposition; samples were buried and heated to temperatures high enough to erase the thermochronometric signal of the source area and the reset age may record cooling that is due to subsequent exhumation. Mixed reset

samples contain multiple age components younger than deposition. Partially reset samples contain multiple age components, which both predate and postdate deposition. Unreset samples have one or more age components older than deposition and record cooling in the sediment source area(s).

To estimate the onset of rapid cooling for individual reset samples, we conducted inverse thermal modeling using the HeFTy software (Ketcham, 2005). HeFTy uses a Monte Carlo approach to generate possible candidates of time-temperature paths for which it calculates the resulting thermochronometric ages and fission track length data. The modeled data are then compared with observations. Those time-temperature paths that resulted in modeled data that agree within statistical limits with the observations are retained, while others are discarded. We report “good” and “acceptable” thermal history envelopes according to the commonly used probability of a worse fit calculated with a Kuiper’s statistical test (good: 0.5; acceptable: 0.05; Ketcham, 2005). The modeled range of possible thermal histories was used to test the consistency between individual sample cooling histories and the regional age vs. depth plot.

In contrast to previous thermal modeling of AFT data (Barnes et al., 2012; 2008; 2006), the thermal history models reported in this study synthesize the new AHe and ZHe ages with the published AFT data, resulting in more precisely delimited cooling paths that are constrained over a larger temperature range (see Supplementary Information for complete modeling details). Finally, we resolved separate age components (Fig. A3) within mixed reset AFT age distributions in the South Subandes (Barnes et al., 2008) using the RadialPlotter software (Vermeesch, 2009), which implements mixture modeling from Galbraith (2005). Younger, reset age components may record the timing of exhumation related to Subandean thrusting, in contrast to older age components which do not.

Similar to many studies (e.g., Barnes et al., 2006; 2008; 2012; Carrapa and Decelles, 2015; Ege et al., 2007; Eichelberger et al., 2013; Enkelmann et al., 2015b; Gillis et al., 2006; McQuarrie et al., 2008a; Oncken et al., 2006), we assume that the early stages of

erosionally-driven exhumation are the result of deformation because thrust faulting facilitates erosion by generating topography and relief (e.g., Willett, 1999; Willett and Brandon, 2002). We recognize that erosion can continue because of duplexing at depth (Lock and Willett, 2008) and also that erosion can continue long after deformation has ceased (e.g., Lease and Ehlers, 2013). Thus, thermochronometer data from thrust belts are sensitive to the onset time of deformation, but may poorly constrain when deformation has terminated because of protracted erosional exhumation. The early stage of exhumation provides a minimum age for the onset of deformation. The Subandes have subdued topography despite 10 mm/yr of active shortening (Brooks et al., 2011) and the exposure of >3-7-km-deep structural levels. This suggests that the high 1-4 mm/yr precipitation rates (Bookhagen and Strecker, 2008) and 0.3-1 mm/yr cosmogenically-derived erosion rates (Insel et al., 2010b) across the Subandes efficiently erode structural relief generated by thrust faulting soon after its creation. Thus, we assume a short, <1 m.y. lag time between the onsets of Subandean deformation and exhumation.

4. RESULTS and INTERPRETATION

We present new and previous thermochronometer results in their structural context for each of the four studied regions of the Subandean thrust belt. First, sample locations are shown in both deformed and restored cross sections (part A in Figs. 2, 3, 4 and 5) to display sample paleodepths. Depths below the Cenozoic are reported because of uncertain local thicknesses of Cenozoic strata deposited before deformation commenced. Then all thermochronometer ages are compiled in a regional age vs. structural depth plot (part B in Figs. 2, 3, 4 and 5), with an inset plot for reset ages that record rapid cooling (part C). Sparse sample coverage and poor mineral abundances within individual transects precluded more detailed local reconstructions. Finally, new inverse thermal models are presented for individual reset samples (part D in Figs. 2, 3, 4, and 5). Below, the four regions of the Subandes are discussed separately, proceeding from south to north.

4.1 South Subandes 20°S

A synthesis of 10 age analyses from 6 samples is presented. Samples are from Devonian, Carboniferous, and Mesozoic strata at 1-2.9 km depth below the Cenozoic (Fig. 2A). Cenozoic strata are approximately 3 km thick in the adjacent foreland basin. Mixed reset and reset AFT and AHe thermochronometer data show a wide 264 to 10.6 Ma range of ages that decrease with depth (Fig. 2B). Barnes et al. (2008) interpreted an exhumed Partial Annealing Zone (PAZ; e.g. Fitzgerald et al., 1995) based on the shallow AFT age-depth gradient, reduced AFT mean track lengths of 10-11.6 μm whose variance increases with depth, and increasing percentages of reset AFT grains with depth. In agreement with this, AHe samples at 1 and 1.3 km depth show large ranges of AHe single grain ages (Fig. 2B) that are typical of rocks that spent a long time in the AHe Partial Retention Zone (PRZ) before exhumation to the surface. In contrast, samples with greater structural depths of 2.6 and 1.6 km have tightly clustered AHe single grain ages (Table S1) and AHe mean ages of 10.6 ± 2.2 Ma and 11.1 ± 3.9 Ma for these samples show little variation with depth. This pattern suggests that these samples resided at higher temperatures prior to the onset of rapid exhumation, at which point they were exhumed rapidly through the PRZ. Similar ages are found in the youngest AFT age components of 11.7 ± 2.8 and 10.5 ± 5.9 Ma from samples at 2.6 and 2 km depths (Figs. 2C, A3).

We interpret the consistency of 12-10 Ma reset cooling ages at deep levels of the South Subandes thrust belt to indicate the onset of erosion. This interpretation is supported by thermal models from individual reset samples that suggest the onset of cooling at 12-10 Ma and since 11 Ma (Fig. 2D). Overall, we interpret the age-depth and thermal model results to suggest that thrust-induced erosion of the South Subandes began at 11 ± 3 Ma.

4.2 South-Central Subandes 18°S

Six thermochronometric ages from 3 samples were previously reported (Eichelberger et al., 2013). Here they are presented in a new age vs. structural depth plot. Samples are from Devonian, Carboniferous, and Mesozoic sedimentary rocks at depths of 0.3-4.2 km below the Cenozoic (Fig. 3A). Cenozoic strata in the adjacent foreland basin are approximately 2

km thick. Paleozoic ZHe ages from 2.3-4.2 km depths and an AFT age from 0.3 km depth are unreset and predate the time of sediment deposition (Fig. 3B). AFT and AHe ages from greater structural depths of 2.3-4.2 km, however, have reset ages of 12.1 ± 3 , 8.0 ± 2.0 , and 7.5 ± 1.9 Ma that decrease with depth (Fig. 3C).

We interpret the reset ages to indicate that rapid cooling from deep levels of the thrust belt was underway by 12-10 Ma and continued after 8 Ma. Time-temperature modeling of ZHe, AFT, and AHe data from the most deeply buried sample suggests the onset of cooling since 10 Ma (Fig 3D; Eichelberger et al., 2013). Overall, we interpret the age-depth pattern to indicate the onset of rapid thrust-induced erosion of the South-Central Subandes at 11 ± 3 Ma.

4.3 North-Central Subandes / Interandes 17°S

A synthesis of 11 age analyses from 6 samples is presented. Subandean samples were collected from Cambrian and Ordovician sedimentary rocks at 4.5-7 km depth below the Cenozoic (Fig. 4A). Cenozoic strata are 5 km thick in the adjacent foreland. The North-Central Subandes is characterized by the fault ramp of one principal, basement-involved structure. ZHe ages display an increasing degree of burial resetting with depth: from partially reset (4.5 km depth) to mixed reset (5.5 km) to reset (7 km; Fig. 4B). For this same suite of samples, all AFT ages are reset (mean track lengths of 14.2-14.3 μm) with ages that decrease from 5.6 ± 2.0 to 1.8 ± 2.6 Ma between depths of 4.5 and 7 km (Fig. 4C).

We interpret the age-depth pattern to indicate the onset of rapid cooling in the North-Central Subandes at 6 ± 2 Ma. This interpretation is supported by inverse modeling of AFT and ZHe data that suggest consistent time-temperature pathways for individual samples within this hanging-wall depth transect. The onset of rapid cooling is recorded at 7-5 Ma, 5-3 Ma, and 4-2 Ma (Fig 4D) for successively deeper samples, suggesting progressive erosion along the principal thrust starting at 6 Ma. The reset ZHe age for the structurally deepest sample constrains the magnitude of North-Central Subandes exhumation, though its 8.9 ± 1.2 Ma ZHe age pre-dates the 4-2 Ma onset of rapid cooling for this sample that is

based on its 1.8 Ma AFT age and long track length with low variance. Thermal modeling of this sample depicts an initial slow cooling rate of $\sim 5^{\circ}\text{C}/\text{m.y.}$ between 9 and ~ 3 Ma, followed by a rapid $\sim 50^{\circ}\text{C}/\text{m.y.}$ cooling rate from ~ 3 Ma to the present (Fig. 4D).

The timing of thrust-induced erosion in the Interandean zone immediately to the west was also examined. Interandes samples are from Ordovician sedimentary rocks at 3.3-5.6 km depth below the Cenozoic (Fig. 4A). The best timing constraint comes from thermal modeling of AFT and ZHe data from the deepest sample at 5.6 km depth, which is reset (mean track length of $14\ \mu\text{m}$) and shows the onset of cooling starting at 14-21 Ma or earlier (Fig. 4D). Other Interandean ages are partially reset or do not show a consistent age vs. depth pattern. Overall, we interpret a shift from thrust-induced erosion in the Interandes since 14-21 Ma to the onset of erosion in the North-Central Subandes zone starting at 6 ± 2 Ma.

4.4 North Subandes 15°S

A synthesis of 9 ages from 6 samples in the medial part of the North Subandean zone is presented. Samples are from Devonian, Carboniferous, and Mesozoic sedimentary rocks at 0-3.5 km depth below the Cenozoic (Fig. 5A). The thickness of Cenozoic strata exceeds 5 km in the adjacent foreland basin. Samples from the top of pre-Cenozoic section (depth = 0 km) have unreset ZHe ages and partially and mixed reset AFT ages (Fig. 5B) with mean track lengths of 10.7-12 μm . These ages indicate incomplete burial resetting prior to Subandean deformation. At depths of 1-3.5 km, however, mixed reset AFT (mean track length of 13 μm) and reset AHe thermochronometers have ages of 6.0 ± 0.5 , 6.6 ± 0.9 to 5.0 ± 1.1 Ma that show little variation with depth (Fig. 5C).

We interpret the consistent 6.6-5 Ma cooling ages over a wide range of depths in the North Subandes to indicate the onset of erosion. This interpretation is supported by time-temperature modeling of individual samples that suggest the onset of cooling at 8-4 Ma and 7-3 Ma (Fig. 5D). Overall, the age-depth pattern and thermal modeling from the North Subandes suggest the onset of cooling at 6 ± 2 Ma due to thrust-induced erosion.

4.5 Exhumation magnitudes

We calculate the magnitude of Subandean exhumation based on the depth of the transition from reset to partially/mixed reset ages in each region and the closure temperature of the thermochronometer that records this transition. We use the modern Subandean geothermal gradient ($18 \pm 5^\circ\text{C}/\text{km}$; Springer and Forster, 1998; Barnes et al., 2008). From this, we calculate Subandean exhumation magnitudes of 4.0 ± 0.7 km in the South, 7.9 ± 1.7 km in the South-Central, 9.7 ± 2.6 km in the North-Central, and 7.3 ± 1.9 km in the North (Fig. 6E). These calculations refine previous estimates based on more limited thermochronometric data (Barnes et al., 2012). Our data indicate that exhumation magnitudes vary by more than a factor of two along the Subandes between 15 and 20°S .

5. DISCUSSION

5.1 North-south variations in foreland and hinterland datasets

5.1.1 Subandes thrust belt exhumation

The thermochronologic results suggest a 5 ± 3.5 m.y. difference in the timing of thrust belt exhumation in the northern vs. southern portions of the Subandes. We compiled all reset low-temperature thermochronologic ages and depths associated with Subandean exhumation from the four studied regions (Fig. 7C). The aggregated results for the two southern sections located at 18 - 20°S indicate the onset of exhumation at 11 ± 3 Ma: most ages are between 12 and 10 Ma and deep samples, indicating continuing exhumation since this onset, are as young as 8 Ma. In contrast, for the two northern sections located at 15 - 17°S , exhumation began later, at 6 ± 2 Ma: most ages are between 7 and 5 Ma and deeper samples, indicating continuing exhumation, are as young as 2 Ma. We interpret the onset of exhumation to herald the timing of deformation and thrust belt propagation into the Subandes.

5.1.2 Subandes foreland deposition

Our interpretation of diachronous thrust belt exhumation is corroborated by basin analyses in the northern vs. southern foreland. As summarized by Uba et al. (2009), the onset of southern Subandean deformation is generally regarded as coeval with deposition of the Yecua Formation (Gubbels et al., 1993; Marshall et al., 1993; Dunn et al., 1995; Uba et al., 2005; 2006; 2009). Subandean deposition commenced in the Oligocene (Sempere et al., 1990; Marshall et al., 1993) with slow accumulation of the Petaca Formation in a forebulge-backbulge setting (Uba et al., 2006; 2009) when the deformation front was located far to the west in the Eastern Cordillera and Interandean zones. In contrast, rapid Miocene accumulation of the Yecua formation occurred in a foredeep characterized by accelerated subsidence, a low-angle basal unconformity, and underfilled accommodation (Uba et al., 2006). Rapid increases in both shortening and subsidence rates during Yecua deposition have been interpreted to reflect active southern Subandean deformation (Uba et al., 2009; Echavarria et al., 2003). Tephra from the basal and lower Yecua formation are dated at 12.4-10.5±0.5 Ma (U/Pb zircon; Uba et al., 2007; 2009) and provide chronostratigraphic constraints for this deformation. This timing is consistent with our thermochronologic interpretation of an 11±3 Ma onset of exhumation of southern Subandean thrust sheets (Fig. 7C).

Deposits of the northern Subandean foreland suggest a Late Miocene onset of deformation, but are less well studied. The Cangalli conglomerate blanketed irregular topography of the Interandean zone and extended across the relatively undeformed Subandean zone during 10-7 Ma (Mosolf et al., 2011; Strub et al., 2005). Mosolf et al. (2011) studied deposits of the Interandes and interpreted (1) the cessation of Interandean deformation by the ~10 Ma onset of deposition and (2) that the 10-7 Ma sediment load promoted deformation propagation. We interpret the post-7 Ma incision of Cangalli basin fill to suggest propagation of deformation into the Subandes was complete. In addition, proprietary industry data have been interpreted to indicate the major phase of northern Subandean deformation and deposition at ~6 Ma (Roeder et al., 1995; Baby et al., 1995). The onset of

exhumation of northern Subandean thrust sheets at 6 ± 2 Ma (Fig. 7C) is consistent with this basin evolution.

5.1.3 Pre-Andean stratigraphy of the Subandes

Miocene-Recent Subandean deformation has occurred across a variable architecture of Paleozoic strata (Baby et al., 1994; Sheffels, 1995). A regional stratigraphic section along the boundary between the Subandes and its undeformed foreland highlights major along-strike changes in Paleozoic basins (Fig. 6A; McGroder et al., 2015). A >5 km thick Paleozoic section in the south is truncated and pinches out across the Chapare Arch before resuming in the north at <2 km thickness. Original Paleozoic thicknesses are greater towards the thrust belt; within the medial portion of the Subandes, the Ordovician-Carboniferous section is 8-10 km thick in the southern cross-sections and 5-6 km thick in the northern cross-sections. The edge of the Chapare basement high has been interpreted to have limited thrust belt propagation (Baby et al., 1994; Hinsch et al., 2002; Sheffels, 1995). Cross-sections show basement-involved structures along this trend (Figs. 4A and 6C) in contrast to a more detached, thin-skinned thrust belt in the southern Subandes with greater overall propagation (Fig. 6D). In addition, the Chapare high is adjacent to an anomalous topographic and structural high in the Eastern Cordillera (Fig. 6B), suggesting rock uplift focused by the basement high (Whipple and Gasparini, 2014).

The timing of Subandean exhumation, however, appears to transcend these structural and exhumational boundaries. For example, a 6 ± 2 Ma onset of exhumation is observed in both of the northern Subandes sections, despite a more detached thrust belt in the North Subandes with 4.1 km of exhumation (Fig. 5A) vs. a basement-involved structure in the North-Central Subandes with 7.3 km of exhumation (Fig. 4A). We conclude that the *timing* of Subandean deformation advance was driven by factors other than lateral variations in Paleozoic stratigraphy, though these variations surely affected the deformation *style*.

The along-strike pattern of Subandes exhumation magnitude correlates with patterns of the Subandes width, shortening percentage, and average precipitation (Fig. 6E). The

consistency of these patterns suggests that erosion played a key role in the propagation of the Subandes thrust belt (McQuarrie et al., 2008b; Horton, 1999, Masek et al., 1994), as expected from critical wedge theory (Dahlen and Suppe, 1988; Willett, 1993). The wetter, more erosive climate in the north limited overall propagation length and created a narrower belt with significant out-of-sequence deformation (Horton, 1999; Masek et al., 1994; McQuarrie et al., 2008b), in contrast to in-sequence deformation in the wider, more arid southern belt.

5.1.4 Hinterland erosion

The Central Andean hinterland has experienced a spatially variable timing and magnitude of Miocene-Recent erosion. The southern Bolivian hinterland has been largely shielded from erosion since ~12 Ma as evinced by preservation of the extensive San Juan del Oro erosion surface (Kennan et al., 1997). Inasmuch as erosion is driven by precipitation (Ferrier et al., 2014), this limited erosion suggests establishment of orographic barriers by ~12 Ma that would promote deformation propagation (Sobel et al., 2003). Indeed, a ca. – 5‰ decrease in the $\delta^{18}\text{O}$ of southern foreland paleosols starting at 12 Ma suggests increased rainout due to growing hinterland topography (Mulch et al., 2010). Furthermore, the cessation of hinterland deformation since erosion surface formation at ~12 Ma heralds the transfer of deformation eastward, into the southern Subandes (Gubbels et al., 1993).

In contrast to little to no erosion of the southern hinterland since 12 Ma, a large 3.5-km-magnitude of erosion has occurred across the northern hinterland since ~11 Ma (Barnes et al., 2006; Gillis et al., 2006). In the adjacent Interandean wedgetop basin, sediments deposited between 10 and 7 Ma were derived from the rapidly-exhumed northern hinterland (Mosolf et al., 2011). Enhanced hinterland erosion would have stalled orogenic wedge growth in northern Bolivia during the same time that little to no hinterland erosion promoted wedge growth in southern Bolivia. Forward propagation in northern Bolivia would resume once the wedge geometry changed (Garzione et al., 2006) or critical taper was rebuilt via out-of-sequence thrusting (Dahlen and Suppe, 1988) and/or deposition on top of the wedge (Fillon et al., 2013).

5.1.5 Altiplano hinterland paleoclimate

The Central Andean hinterland experienced large, abrupt changes in paleoclimate during the Miocene. Decreases in average paleosol carbonate $\delta^{18}\text{O}$ by ca. -3.5‰ , Δ_{47} -derived soil temperatures by ca. -15°C , and authigenic clay δD by ca. -15‰ suggest that the Bolivian Altiplano became cooler in the middle to late Miocene (Garzzone et al., 2014; 2008; 2006; Ghosh et al., 2006). A decrease in fluvial deposits and an increase in paleosol carbonate $\delta^{13}\text{C}$ by ca. $+3\text{‰}$ suggest that Miocene Altiplano cooling was accompanied by increased aridity (Garzzone et al., 2008; Quade et al., 2007; Smith et al., 2011).

Miocene cooling and increased aridity was diachronous along the length of the Bolivian Altiplano. Cooling occurred earlier in the southern Bolivian Altiplano, between 16 and 13 Ma (Garzzone et al., 2014), than in the northern Bolivian Altiplano, where it occurred between 10 and 7 Ma (Fig. 7B) (Garzzone et al., 2008; 2006; Ghosh et al., 2006). Our thermochronometric data show that the onset of exhumation in the low-elevation foreland followed Altiplano hinterland climate cooling; Subandean exhumation commenced in the south at 11 ± 3 Ma and in the north at 6 ± 2 Ma (Fig. 7C).

An earlier phase of cooling has been observed in the northern Bolivian Eastern Cordillera and is indicated by a decrease in average paleosol carbonate $\delta^{18}\text{O}$ by ca. -5‰ and Δ_{47} -derived soil temperatures by ca. -18°C between ca. 24 Ma and 15 Ma (Leier et al., 2013). The degree to which Miocene cooling reflects either rapid surface uplift (Garzzone et al., 2014; 2008; 2006; Ghosh et al., 2006) or threshold climate effects (Ehlers and Poulsen, 2009; Insel et al., 2012; Poulsen et al., 2010) is an area of ongoing debate (see section 5.2).

5.2 Potential mechanisms linking hinterland uplift/climate and foreland exhumation

The abrupt changes in Altiplano paleoclimate proxy data described in section 5.1.5 have elicited competing interpretations of rapid surface uplift due to lithosphere removal (Garzzone et al., 2014; 2008; 2006; Ghosh et al., 2006) versus abrupt changes in climate

dynamics driven by gradual surface uplift above orographic threshold elevations (Ehlers and Poulsen, 2009; Insel et al., 2012; Poulsen et al., 2010). Comparing Altiplano paleosol data to modern, linear lapse rates suggests rapid surface uplift of ~ 2.5 km at a rate of ~ 1 km/Myr (Garzione et al., 2008). Climate modeling, however, suggests that the same paleosol data are consistent with steady surface uplift at a rate of ~ 0.1 km/Myr because paleo-lapse rates change non-linearly as surface uplift progresses (Insel et al., 2012). In particular, changes in moisture source, transport, and availability occur during surface uplift (Poulsen et al., 2010) and the largest isotopic changes occur once a threshold elevation of 3 km is exceeded (Insel et al., 2012; Poulsen et al., 2010). Though the mechanisms driving the isotopic shifts and the rates of Miocene surface uplift are debated (Ehlers and Poulsen, 2009; Garzione et al., 2008), the contrasting interpretations both require absolute elevations >3 km to explain the observed isotopic shifts. In addition, the consistent, sequential timing of Altiplano hinterland surface uplift and Subandean foreland exhumation in the Central Andes suggests that these two processes may be linked via orogenic wedge dynamics. Finally, both competing interpretations of the paleoclimate proxy data are compatible with deformation propagation into the Subandean zone. Below these scenarios are discussed in more detail.

5.2.1 Rapid surface uplift due to lithosphere removal

The interpreted large 2.5 km magnitude and rapid ~ 1 km/Myr rate of Altiplano surface uplift to an elevation of 4 km has been suggested to reflect isostatic adjustment to the catastrophic removal of dense lower lithosphere (Fig. 8A; Garzione et al., 2008; 2006). Diachronous Altiplano surface uplift events have been interpreted as separate lithosphere removal events; at 16-13 Ma in southern Bolivia and 10-7 Ma in northern Bolivia (Garzione et al., 2014; Fig. 7). An earlier ca. 24-15 Ma phase of surface uplift in the northern Bolivian Eastern Cordillera to an elevation of 2.5 km has been interpreted as a separate lithosphere removal event (Leier et al., 2013). Rapid hinterland surface uplift is predicted to induce a supercritical phase of orogenic wedge growth and the forward propagation of deformation (Fig. 8A). Propagation is driven by an increase in horizontal force per unit length and

rotation of the orogenic wedge (i.e. a decrease in decollement slope and increase in surface taper angle; Garzione et al., 2006).

5.2.2 Gradual surface uplift above orographic threshold elevations

Alternatively, the Altiplano paleoclimate proxy data have been interpreted to reflect an abrupt change in climate dynamics once a threshold elevation of 3 km was exceeded during gradual surface uplift (Fig. 8B). In this scenario, the Altiplano paleosol isotopic shifts document the intensification of convective rainfall at the windward, eastern range front (Ehlers and Poulsen, 2009; Insel et al., 2012; Poulsen et al., 2010). The diachronous Altiplano cooling and aridification suggest an earlier attainment of 3 km elevations and convective rainfall in southern Bolivia, by 13 Ma, than in northern Bolivia, by 7 Ma (Fig. 7). Most modern orographic precipitation in the Central Andes falls below a ~3 km elevation threshold (Fig. 9). Precipitation rates drop from 4-1 m/yr to <1 m/yr once this threshold is exceeded, suggesting that hinterland surface uplift above 3 km significantly changes precipitation and hence erosion patterns and amounts (Barnes et al., 2012; Norton and Schlunegger, 2011; Bookhagen and Strecker, 2012). As convective rainfall intensifies at the range front, the leeward hinterland would be increasingly shielded from moisture-bearing easterlies and thus erosion (Fig. 8B). Reduced hinterland erosion would promote orogenic wedge growth by allowing mass to accumulate within the hinterland, which would drive forward (i.e., eastward) propagation (Roe et al., 2008; Sobel et al., 2003; Strecker et al., 2007).

5.3 Summary: Punctuated Propagation

Competing interpretations of the paleoclimate proxy data concur that Altiplano hinterland elevations >3 km were attained by the end of the paleosol isotopic shift, though there is disagreement whether this threshold was attained abruptly or gradually. Surface uplift of the orogenic hinterland, whether punctuated or steady, will drive outward propagation of deformation (Hilley and Strecker, 2004; McQuarrie et al., 2008b). Our thermochronologic data indicate a coincident onset of orogenic wedge propagation into the Subandes. Thus,

we conclude that Altiplano hinterland elevations >3 km were required to drive the most recent phase of propagation in the Central Andes (Fig. 10). Earlier, ca. 24-15 Ma surface uplift in the northern Bolivian Eastern Cordillera (Leier et al., 2013) may have driven an earlier phase of propagation into the northern Interandean Zone, where exhumation accelerated after 15 Ma (Barnes et al., 2006). Spatiotemporal variations in both thermochronometric and paleoclimate proxy data (Fig. 7) indicate that orogenic propagation was piecemeal. Irrespective of uplift rate, we interpret that surface uplift above 3 km elevation occurred in the southern Altiplano and drove propagation into the southern Subandes at 11 ± 3 Ma (Fig. 10B). A similar evolutionary sequence occurred later in the northern Altiplano and northern Subandes, at 6 ± 2 Ma (Fig. 10C). The piecemeal history of orogenic wedge propagation in the Central Andes suggests that this process occurs over areas <200-500 km wide. This lengthscale is in accord with those suggested by geophysical (Beck, 2002; Beck et al., 2015) and paleoclimate studies (Garzione et al., 2014; Leier et al., 2013; Saylor and Horton, 2014).

6. CONCLUSION

Along-strike differences in the onset of Bolivian Subandean (Central Andes frontal retroarc) thrust belt exhumation are evident from low-temperature thermochronometry. Cooling ages versus structural depth from balanced cross-sections exhibit an earlier onset of thrust-induced erosion in the south (11 ± 3 Ma) than in the north (6 ± 2 Ma), which we interpret as diachronous thrust belt propagation. Each phase of foreland exhumation is coeval with completion of an abrupt shift to cooler, more arid conditions in the adjacent Altiplano hinterland. Previous interpretations of the hinterland paleoclimate proxy data are either rapid surface uplift due to removal of dense lower lithosphere or an abrupt change in climate dynamics/orography during more gradual surface uplift. Both of these mechanisms would drive forward propagation of the orogenic wedge by augmenting the hinterland “backstop,” either directly through surface uplift or via spatially variable erosion. These processes may have operated in tandem once >3 km elevations were exceeded during growth of the Central Andean Plateau. The piecemeal history of

Subandean propagation suggests a restricted <200-500 km width of deforming regions in response to punctuated hinterland change.

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FIGURE CAPTIONS:

Figure 1

Central Andean orogen topography and tectonomorphic zones: SA-Subandes, IA-Interandes, EC-Eastern Cordillera, AL-Altiplano. Subandes study regions are depicted with white lines for cross-section locations (Figs. 2, 3, 4, 5) and white boxes for geologic maps (Figs. A1, A2).

Figure 2

South Subandes 20°S: (A) balanced structural cross-section (McQuarrie, 2002) displaying thermochronology sample locations and ages; (B) regional cooling history from age-depth plot of 60°C-110°C thermochronometers (This study; Barnes et al., 2008) with inset (C) for reset ages that record rapid cooling; (D) thermal histories of individual samples from inverse modeling of AHe and AFT data. PAZ: Partial Annealing Zone for AFT ages; PRZ: Partial Retention Zone for AHe ages; MR: mixed reset. Percentage of reset grains for AFT young components given in parentheses. Note that widely-spaced samples from multiple thrust sheets are compiled onto regional age-depth plot.

Figure 3

South-Central Subandes 18°S (A) balanced structural cross-section (Eichelberger et al., 2013) displaying thermochronology sample locations and ages; (B) regional cooling history from age-depth plot of 60-180°C thermochronometers (Eichelberger et al., 2013) with inset (C) for reset ages that record rapid cooling; (D) thermal history of individual sample from inverse modeling of AHe, AFT, and ZHe data (Eichelberger et al., 2013). Note that widely-spaced samples from multiple thrust sheets are compiled onto regional age-depth plot.

Figure 4

North-Central Subandes / Interandes 17°S: (A) balanced structural cross-section (McQuarrie, 2002) displaying thermochronology sample locations and ages; (B) regional cooling history from age-depth plot of 110°C-180°C thermochronometers (This study; Barnes et al., 2012) with inset (C) for reset ages; (D) thermal histories of individual samples from inverse modeling of AFT and ZHe data. MR: mixed reset; PR: partially reset.

Figure 5

North Subandes 15°S: (A) balanced structural cross-section (McQuarrie et al., 2008a) displaying thermochronology sample locations and ages; (B) regional cooling history from age-depth plot of 60°C-180°C thermochronometers (This study; Barnes et al., 2006) with inset (C) for reset ages that record rapid cooling; (D) thermal history of individual samples from inverse modeling of AHe and AFT data (top) and AFT and ZHe data (bottom). Percentage of reset grains for AFT young components given in parentheses.

Figure 6

(A) Along-strike (orogen-parallel) stratigraphic section along the eastern terminus of the Subandean thrust belt highlighting north vs. central vs. south differences in geologic substrate (modified from McGroder et al., 2015); (B) Map displaying thicknesses of Paleozoic strata in the Bolivian Subandean foreland (<2 km thin; absent; >5 km thick) and cross-sections lines for A, C, and D; (C) Across-strike (orogen-perpendicular) structural cross-section (Baby et al., 1994) exhibiting influence of Chapare basement arch: narrow, basement-involved thrust belt. Note adjacent topographic high in the eastern Cordillera;

(D) Across-strike (orogen-perpendicular) structural cross-section (McQuarrie, 2002) exhibiting influence of thick Paleozoic stratigraphic section: wide, detached thrust belt; (E) Covariance between Subandean exhumation magnitudes (black), structural depths (green), average precipitation (blue), shortening percentage (red), and thrust belt width (orange). Data from this study; Barnes et al., 2012; 2008; 2006; Eichelberger et al., 2013; McQuarrie, 2002; McQuarrie et al., 2008a; Bookhagen and Strecker, 2008.

Figure 7

(A) Space-time pattern of changes in hinterland climate (clouds, numbered circles; Table 2) and foreland exhumation (boxes) in the Central Andes. Note 5 m.y. timing difference between the northern (blue) and southern (pink) areas; (B) Timing of northern vs. southern Altiplano hinterland cooling and aridification from decreases in average paleosol carbonate $\delta^{18}\text{O}$ by ca. -3.5‰ , Δ_{47} -derived soil temperatures by ca. -15°C , and authigenic clay δD by ca. -15‰ (Garzzone et al., 2014; 2008; 2006; Ghosh et al., 2006) (circles in part A); (C) Timing of northern vs. southern Subandean foreland exhumation from age-depth relationship of all reset AHe and AFT thermochronometers (Figs. 2C, 3C, 4C, 5C; boxes in part A).

Figure 8

Competing interpretations of Altiplano hinterland climate proxy data and potential mechanisms for thrust belt propagation into the Subandean zone (SAZ). (A) Rapid hinterland surface uplift due to catastrophic removal of dense lower lithosphere. High-elevation, buoyant hinterland drives Subandean thrust belt propagation by 1) applying an increased force per unit length to low elevation surroundings, and 2) rotation of the thrust belt, i.e., a decrease in decollement slope and increase in surface taper angle; (B) Abrupt change in climate dynamics once orographic threshold elevations were exceeded during gradual surface uplift. Large change in precipitation/erosion patterns occurs and reduced hinterland erosion augments orogenic wedge backstop and promotes thrust belt propagation into the Subandes. IAZ – Interandean Zone.

Figure 9

Central Andean orogen precipitation and topography. (A) Mean annual precipitation map from 1998-2006 TRMM satellite data (Bookhagen and Strecker, 2008); (B), (C), (D) Swath profiles of precipitation and topography; Solid line is average and shaded envelope denotes maximum and minimum values within 50-km-wide swath. Tectonomorphic zones: SA-Subandean, IA-Interandean, EC-Eastern Cordillera, AL-Altiplano.

Figure 10

Proposed model for Miocene propagation of the Central Andean Orogen: piecemeal attainment of >3 km elevations in the Altiplano hinterland since 16 Ma (A) drive diachronous Subandean thrust-belt exhumation in the south at 11 Ma (B) followed by the north at 6 Ma (C). White lines are cross-section locations (Figs. 2-5) and numbered white circles are locations of paleoclimate studies (Table 2). See figure 7 of Saylor and Horton (2014) for more detailed reconstruction of 16 Ma paleotopography,

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Figure 1

Fig. 1 Lease et al.

Figure 2

Figure 2 Lease et al.

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Figure 3 Lease et al.

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Fig. 4 Lease et al.

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Fig. 9 Lease et al.

Figure 10

Fig. 10 Lease et al.

Table 1

Table 1. Summary of apatite (U-Th)/He [AHe], apatite fission track [AFT], and zircon (U-Th)/He [ZHe] data

Sample	Latitude	Longitude	Elevation (m)	Depth below Cenozoic (km)	Depositional age range (Ma)	AHe mean age (Ma)	AHe error (2σ)	AFT young comp. (Ma)	AFT error (2σ)	AFT pooled age (Ma)	AFT error (2σ)	AFT mean track length (μm)	AFT error (1σ)	ZHe mean age (Ma)	ZHe error (2σ)
South Subandes 20°S															
S-SA1	-19.5138	-64.1710	1825	2.9	360-390	-	-	-	-	^b 48.3	17.4	11.6	2.2	-	-
S-SA2	-19.8041	-64.0211	1236	2.6	360-390	10.6	2.2	11.7	2.8	^b 41.9	6	11.5	2.6	-	-
S-SA3	-19.8538	-63.7285	1367	1	300-360	MR		-	-	^b 264.3	31.6	11.1	1.6	-	-
S-SA4	-19.8085	-63.7181	1260	1.6	300-360	11.1	3.9	26.1	7	^b 117	13.8	10.6	2.3	-	-
S-SA5	-19.8428	-63.4849	1137	2	300-360	-	-	10.5	5.9	^b 103.9	14.6	10.0	1.9	-	-
S-SA6	-19.7813	-63.2485	844	1.3	300-360	MR		-	-	^b 246.7	28.6	10.8	1.5	-	-
South-Central Subandes 18°S															
SC-SA1	-63.55	-18.18	684	2.3	300-360	-	-	-	-	^d 12.1	3.0	-	-	^d 458	75
SC-SA2	-63.72	-18.18	1285	0.3	65-200	-	-	-	-	^d 316	22	-	-	-	-
SC-SA3	-64.05	-17.99	1178	4.2	360-390	^d 7.5	1.9	-	-	^d 8	2.0	-	-	^d 475	7.0
North-Central Subandes/Interandes 17°S															
NC-SA1	-17.1581	-65.7350	1816	4.5	490-445	-	-	-	-	^c 5.6	2	14.2	1.5	PR	
NC-SA2	-17.1031	-65.6813	911	5.5	490-445	-	-	-	-	^c 4.1	1.4	14.2	1.4	MR	
NC-SA3	-17.0570	-65.6464	619	7	490-540	-	-	-	-	^c 1.8	2.6	14.3	1.2	8.9	1.2
NC-IA1	-17.2321	-65.8875	3288	3.3	490-445	-	-	-	-					47.1	11
NC-IA2	-17.1728	-65.8987	2875	4.7	490-445	-	-	-	-	^c 125.7	100	10.5	2.4	MR	
NC-IA3	-17.1919	-65.8175	1856	5.6	490-445	-	-	-	-	^c 8.1	3.6	14.0	1.8	21	5.3
North Subandes 15°S															
N-SA2	-15.5260	-67.2570	801	0	65-200	-	-	-	-	^a 75.5	37.6	12.0	3.0	-	-
N-SA3	-15.3950	-67.1370	820	2.5	360-390	-	-	^a 5	0.8	^a 6.6	0.9	13.0	2.7	^a 330	35
N-SA4	-15.3700	-67.1080	640	0	65-200	-	-	^a 9.5	4	^a 51.5	5.6	11.8	2.6	^a 440	311
N-SA4B	-15.3662	-67.1053	586	1	300-360	5	1.1	-	-	-	-			-	-
N-SA4C	-15.3521	-67.0881	562	3.5	390-415	6	0.5	-	-	-	-			-	-
N-SA5	-15.2350	-67.0720	790	0	65-200	-	-	^a 39.5	6.7	^a 91.4	9.4	10.7	2.1	^a 315	25

MR = mixed reset. No mean age reported
PR = partially reset. No mean age reported
AHe analyses reported in Table S1
ZHe analyses reported in Tables S2

	this paper
^a	Barnes et al., 2006
^b	Barnes et al., 2008
^c	Barnes et al., 2012
^d	Eichelberger et al., 2013

Table 2

Table 2. Locations of paleoclimate proxy studies

Map ID	Region	Reference(s)	Latitude	Longitude	Location	Age (Ma)	Time (Ma) when >3 km elevations exceeded
1	northern Bolivian Altiplano	Garzione et al., 2006; 2008; Ghosh et al., 2006	-17.619	-68.3	Corque	12-5	7
2a	southern Bolivian Altiplano	Garzione et al., 2014	-20.883	-66.317	Cerdas	16	13
2b			-21.967	-65.133	Quebrada Honda	13	
2c			-19.933	-66.833	Quehua	9-7	
3	northern Bolivian Eastern Cordillera	Leier et al., 2013	-17.177	-67.62	Salla	29-15	