**Rapid regional surface uplift of the northern Altiplano plateau revealed
by multiproxy paleoclimate reconstruction**

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Abstract

The central Altiplano is inferred to have experienced $\sim 2.5 \pm 1$ km surface uplift between ~ 10 and 6 Ma, while the southern Altiplano experienced a similar magnitude of surface uplift that began earlier, between ~ 16 and 9 Ma. To properly constrain the along strike timing of the Altiplano plateau surface uplift, it is necessary to know how and when the northernmost part of the Altiplano plateau evolved. We reconstruct the paleoclimate and infer the corresponding paleoelevation from the Miocene-Pliocene deposits of the Descanso-Yauri basin ($14\text{--}15^\circ\text{S}$) in the northernmost part of the Altiplano plateau using 4 different proxies, including carbonate clumped isotope composition (i.e., Δ_{47} values), carbonate $\delta^{18}\text{O}_\text{c}$, leaf wax $\delta\text{D}_\text{wax}$ and pollen assemblages from paleosol, lacustrine and palustrine carbonates and organic-rich sediments. The isotopic signatures reflect past climate conditions of mean annual air temperature (Δ_{47}) and meteoric water isotope values ($\delta^{18}\text{O}_\text{c}$, $\delta\text{D}_\text{wax}$). Our results show that the northernmost plateau remained at low elevation (0.9 ± 0.8 to 2.1 ± 0.9 km) until late Miocene time (~ 9 Ma) characterized by $\sim 15^\circ\text{C}$ warmer than modern temperature (mean annual air temperature of $23 \pm 4^\circ\text{C}$, 2σ), low elevation vegetation and precipitation signature with reconstructed $\delta^{18}\text{O}_\text{mw}$ (VSMOW) of $-8.3 \pm 2.0\text{‰}$ (2σ) from carbonate ($\delta^{18}\text{O}_\text{c}$) and $-8.6 \pm 1.8\text{‰}$ (2σ) from leaf wax ($\delta\text{D}_\text{wax}$). Modern elevations of 4 km were not reached until 5.4 ± 1.0 Ma, as indicated by a negative shift in $\delta\text{D}_\text{wax}$ (VSMOW) from $-143.4 \pm 12.8\text{‰}$ (2σ) to $-209.2 \pm 21.1\text{‰}$ (2σ) between 9.1 ± 0.7 and 5.4 ± 1.0 Ma. The timing of surface uplift of the northernmost Altiplano is consistent with the evidence for late Miocene surface uplift of the central Altiplano ($16\text{--}19^\circ\text{S}$) between 10 and 6 Ma, and indicates that regional scale uplift in the

northern – central plateau significantly postdates the onset of surface uplift in the southern Altiplano (19–22°S) between ~16 and 9 Ma. These results are consistent with piecemeal removal of the lower dense lithosphere, combined with possible lower/middle crustal flow from south to north in the plateau acting as the main mechanisms for the formation of the Altiplano plateau.

1. Introduction

The spatial and the temporal nature of the surface uplift of the Altiplano plateau in the Central Andes and the associated geodynamic mechanisms behind plateau formation have been the subject of debate in recent times (e.g., Ghosh et al., 2006b; Garzzone et al., 2006, Garzzone et al., 2008; Barnes and Ehlers, 2009). Two end-member models argue between gradual surface uplift associated with crustal shortening and rapid, pulsed surface uplift possibly related to lithospheric removal (Ghosh et al., 2006b; Garzzone et al., 2006, Garzzone et al., 2008; Garzzone et al., 2014). Structural cross-section reconstructions have been used to advocate a slow and gradual surface uplift by distributed shortening of the upper crust (McQuarrie et al., 2005; Lamb, 2011) with continuous removal of dense lower lithosphere through ablative subduction (Pope and Willett, 1998). The uplift then is expected to be contemporaneous with major deformation. However, paleoelevation studies from different parts of the central Andes (Fig. 1A) suggest surface uplift postdated the major deformation and had up to an order of magnitude faster rates than can be accounted for by crustal shortening alone (Gregory-Wodzicki, 1998; Ghosh et al., 2006b; Garzzone et al., 2006; Garzzone et al., 2008; Leier et al., 2013; Garzzone et al., 2014, Saylor and Horton, 2014). The rapid surface uplift pulses in the Altiplano have been characterized as surface expressions of the removal of lower

lithosphere (e.g., Houseman et al., 1981) and/or lower crustal flow (e.g., Husson and Sempere, 2003), decoupled from and postdating the timing of major crustal shortening.

Existing paleoelevation constraints (Fig. 1A) come from various climate proxies: leaf physiognomy (Gregory-Wodzicki, 1998), stable isotopes $\delta^{18}\text{O}$ from carbonates (Garzione et al., 2006; Mulch et al., 2010; Leier et al., 2013), δD from volcanic glass (Saylor and Horton, 2014) and clumped isotope Δ_{47} from carbonates (Ghosh et al., 2006b; Garzione et al., 2008; Leier et al., 2013; Garzione et al., 2014). They suggest that the Eastern (Leier et al., 2013) and the Western (Saylor and Horton, 2014) Cordillera witnessed rapid surface uplift between 24–17 Ma and 19–16 Ma. In the Altiplano, a rapid surface uplift of $\sim 2.5 \pm 1$ km has been inferred between ~ 10 –6 Ma (Ghosh et al., 2006b; Garzione et al., 2006; Garzione et al., 2008) in the central part of the plateau and between ~ 16 –9 Ma in the southern plateau (Garzione et al., 2014) and Western Slope of the Andes (Rech et al., 2006). A 2–3 km surface uplift in the last 12 Ma was also proposed by geomorphic study of low-relief landscapes (Hoke and Garzione, 2008; Whipple and Gasparini, 2014) in the Bolivian Andes.

Atmospheric modeling studies with General Circulation Models (GCMs) show that the regional circulation pattern should respond to plateau uplift, causing intensification of precipitation in the Central Andes (Poulsen et al., 2010; Insel et al., 2012) and non-adiabatic surface temperature change (Ehlers and Poulsen, 2009). These studies (Ehlers and Poulsen, 2009; Insel et al., 2012) further suggest that climate change associated with plateau uplift could produce isotopic shifts and isotopic lapse rate changes that would tend to reduce the estimate of magnitude of surface uplift as compared to paleoelevation estimates with modern isotopic lapse rates. Consistent with inferences from climate

models, a positive shift in pedogenic carbonate $\delta^{18}\text{O}$ at ~ 8.5 Ma (Mulch et al., 2010), coeval with increased river discharge (Mulch et al., 2010) and sedimentation rate (Uba et al., 2007) was reported from Subandean foreland deposits and was inferred as a consequence of intensification of precipitation associated with attainment of 60–70% elevation of the modern Andes.

We present new multiproxy-based paleoclimate constraints from the northernmost Altiplano plateau to provide insights into the along-strike changes in the timing and magnitude of surface uplift of the Altiplano (Fig. 1). The only previous study here (Picard et al., 2008) proposed an elevation of 2.0 to 2.5 km in middle to late Miocene based on the genetic divergence of potato cyst-nematodes as this divergence is elevation dependent. While this estimate is consistent with inferences about the nearby Western Cordillera surface uplift history (Saylor and Horton, 2014) and establishes a threshold paleoelevation for the highest parts of the range, it does not provide constraints on the rate and spatial extent of surface uplift. Better characterization of the surface uplift history of the northern Altiplano is especially important in the context of the evidence pointing to rapid but diachronous surface uplift of the south and central plateau. To reduce the potential uncertainties in paleoelevation associated with any single proxy, we use four proxies (i.e., $\delta^{18}\text{O}_c$ and Δ_{47} from carbonates, and $\delta\text{D}_{\text{wax}}$ from leaf wax and pollen assemblages) to reconstruct the Miocene–Pliocene elevation history of the northernmost Altiplano (Fig. 1 and 2). We incorporate the results from climate models (Ehlers and Poulsen, 2009; Insel et al., 2012) to account for the additional climate changes associated with plateau uplift, to provide a more robust and conservative estimate of the magnitude of surface uplift in the northernmost Altiplano plateau.

2. Geological Background of the Descanso–Yauri Basin

Our study area, Descanso–Yauri basin, is situated at the boundary between the Western Cordillera and the Altiplano and extends from 14-15°S and 71-72°W (Fig. 1B). The basin is bounded on the south and west sides by Cenozoic volcanics, and the Langui and Descanso thrusts of the Cusco-Lagunillas-Manazo (CLM) fault system that extends along its northeastern boundary. It is part of the Miocene-Pliocene Maure–Condoroma group of basins (Carlotto 2013). Regionally it corresponds to the NW termination of the ~4 km high Altiplano plateau in the Central Andes (Fig. 1A).

3. Basin Stratigraphy

The ~1800m of sedimentary fill in the Descanso–Yauri basin is divided into members A, B and C (Fig. 1A) from oldest to youngest (Cerpa and Meza, 2001; Carlotto, 2013) with maximum measured thickness of ~500m, ~1100m and <100m respectively (Fig. 2A). Member A is exposed in a narrow zone adjacent to the northeastern basin boundary, whereas Member B covers most of the basin. Member C is observed locally in the southern part of the basin. The contact between members A and B is conformable. A low angle angular unconformity separates members B and C, observed in the Accocunca section (Fig. 1A and 2B).

In the northeastern part of the basin, lithologies of members A and B are characterized by lenticular beds of massive to cross-stratified and imbricated conglomerates and sandstones with scoured bases, interpreted to be deposited in a braided river depositional environment. Well-developed, 1–2 m thick, calcic paleosols horizons (Fig. 2B and A.1) formed in the overbank floodplain settings of Member B. The tops of the paleosol profiles are identified from decreasing carbonate content and

grain size from fine/very-fine sand to clay/silt, along with increasing red coloration (Fig. 2B). We sampled paleosol carbonates from Cerro Quirma, San Genaro and Ocoruro (Fig. 1 and 2), and the majority of our samples come from a depth of ≥ 1 m from the top. Tabular lacustrine limestone units mark the top of Member B and we sampled these from San Genaro, Kanamarka and Hector Tejada sections. We also sampled lacustrine/palustrine deposits from Checca, San Miguel and Accocunca North sections (Fig 2B).

The stratigraphic positions of the sampled sections are ascertained with respect to Member B top limestones. As the limestone beds are absent in the Cerro Quirma section, we infer that the uppermost Member B is missing here. However, the presence of the Atuhuasi tuff that marks the Member A top, at the base of Cerro Quirma affirms that the lowermost part of this section represents the lowermost Member B. The stratigraphic position of the Ocoruro section is harder to constrain in the absence of a continuous stratigraphy, and we infer that the well-developed paleosol carbonates likely represent the middle part of Member B, between Cerro Quirma and the San Genaro North sections.

Deposition of Member C took place in a fluvial-lacustrine system. The deposits are generally non calcareous, dominated by diatomaceous mudstone. We sampled the diatomaceous mudstones from the Cerro Pucara, Versailles, Los Angeles and Hector Tejada sections (Fig. 1 and 2). We sampled palustrine mudstones from the Huano Huano section of Member C that contains both fluvial lacustrine/palustrine deposits. All three members contain distal ash fall tuff, signifying syn-depositional eruptions of the volcanoes bordering the southern and western side of the Descanso-Yauri basin.

3.1. U/Pb zircon geochronology

We constrain the depositional age through three new U-Pb zircon age analyses from primary tuffs in members B and C (Fig. 3 and Table B.1). Previously published age (Boudesseul et al., 2000), 18.74 ± 0.19 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ biotite) from the Atunhuasi tuff at Member A top provides the oldest age constraint. Ages of 12.82 ± 1.08 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ biotite), 12.09 ± 0.18 Ma, 11.99 ± 0.27 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ Sanidine) (Cerpa and Meza, 2001) were also reported from the Yauri tuff that marks the middle of Member B.

We sampled the Yauri tuff (12MN54t) at Mammanoca (Fig. 1B). We note that the section top is eroded in Mammanoca, and its characteristics appear to reflect the middle of Member B, although it is present near the member B–C boundary. The ~ 2.5 m thick white, fine-grained and ash rich tuff yields a weighted mean U-Pb age of 12.3 ± 1.0 Ma (2σ) (Fig. 3A, Table B.1), supporting the previously reported ages.

We sampled a ~ 1.5 m thick, very fine-grained, ash rich tuff with pumice fragments from San Genaro (12SG19t), exposed directly below the limestone beds that mark the Member B top. The weighted mean age of 9.1 ± 0.7 Ma (2σ) (Fig. 3B, Table B.1) thus provides the youngest Member B age and we assign this age to the upper parts of the Hector Tejada and Kanamarka sections, where limestones are exposed.

Sample (12YR07t) from a ~ 4.5 m thick gray–white fine-grained ash to sand rich tuff in Versailles, exposed approximately 50 m below the top of Member C constrains the youngest depositional age in the Descanso–Yauri basin. The weighted mean age of 4.8 ± 1.1 Ma (2σ) (Fig. 3C, Table B.1) indicates that deposition continued in the basin well into the Pliocene.

The oldest and the youngest age constraints of Member C come from two detrital U-Pb analysis. The oldest Member C age comes from sample 13AC71t, a reworked tuff,

situated just above the unconformity in the Accocunca section. The youngest zircon population (Fig. 3D, Table B.1) yielded a weighted average age of 5.4 ± 1.0 Ma (2σ). Youngest age constraint of the basin comes from the sample 13CP97s, a fine-grained tuff collected near the top of the Cerro Pucara section. This sample yielded a weighted mean age of 3.9 ± 0.1 Ma (2σ) from the youngest zircon population (Fig. 3E, Table B.1), indicating that Member C had the shortest depositional age span of the three members.

3.2. Deformation History

Members A and B deposits are deformed, especially in the northern part of the basin (Fig. 1) where they are folded in anticlines and synclines with nearly vertical limbs (Fig. 2). In comparison, Member C sits in angular unconformity on Member B (Fig. 2) and is relatively flat lying (maximum dip of 5°). Study of the deformation, depositional systems, provenance history and U/Pb geochronology (Kar et al., 2015) of this basin suggests that the basin bounding Langui and Descanso thrusts (Fig. 1A) were active during deposition of members A and B. Activity of the thrusts are also evidenced by folding of members A and B, which show fold axes parallel to the strike of the thrusts. Thrust activity waned by 9.1 ± 0.7 Ma, and the deformation regime switched from compressional to extensional after 9.1 ± 0.7 Ma. Our inference is based on the observed offset of otherwise undeformed 9.1 ± 0.7 Ma limestones along normal faults (Fig. 1A). As the Member C deposits are not affected by the normal faults, we infer that these faults were active prior to 5.4 ± 1.0 Ma.

4. Paleoelevation Proxies

To evaluate the paleoelevation history of the Descanso–Yauri basin, we used four

different climate indicators: pollen taxa, temperature from carbonate clumped isotope analysis (reported here as $T\Delta_{47}$) and the isotopic composition of rainfall inferred from carbonate ($\delta^{18}O_c$) and leaf wax (δD_{wax}). Reconstructed vegetation assemblages from pollen were compared against the distribution of equivalent present vegetation to infer paleoelevations. Paleotemperature estimates from $T\Delta_{47}$ provide a complementary approach, assuming that temperature lapse rates and regional temperature variations can be approximated (Ghosh et al., 2006a; Garzzone et al., 2006; Ehler and Poulsen, 2009). Changes in the stable isotope composition of paleo-meteoric water provide a third mechanism for reconstructing paleoelevation because the $\delta^{18}O$ and δD values of precipitation decrease systematically with altitude (e.g., Gonfiantini et al., 2001). This is caused by the progressive condensation of moisture with increasing elevation (and decreasing temperature) and removal of condensate by rain-out. Provided the moisture source and vapor transport pathway are constrained, these effects can be modeled as a Rayleigh distillation process to estimate paleoelevation (Gonfiantini et al., 2001; Rowley and Garzzone, 2007).

In carbonates, the clumped isotope or the Δ_{47} values characterize the ratio of mass-47 CO_2 (primarily $^{13}C^{18}O^{16}O$ isotopologue) to mass-44 CO_2 ($^{12}C^{16}O^{16}O$), relative to the randomly distributed heavy isotopologue ratio of mass-47/mass-44 (Ghosh et al. 2006a). The tendency to form a higher abundance of doubly substituted $^{13}C-^{18}O$ increases with decreasing temperature, and therefore the empirical relationship between temperature and abundance of clumped isotopologues (Ghosh et al., 2006a) reveals the temperature of carbonate formation. The oxygen isotopic composition of sedimentary carbonates (i.e., $\delta^{18}O_c$) depends on water composition ($\delta^{18}O_{mw}$) and the

temperature of carbonate precipitation. We used the average $T_{\Delta_{47}}$ from each section to constrain the temperature of carbonate precipitation and applied the corresponding fractionation factors (Kim and O'Neil, 1997) to calculate $\delta^{18}\text{O}_{\text{mw}}$ ($\delta^{18}\text{O}_{\text{mwc}}$). Our sedimentary carbonate samples come from Member B (with one exception), as Member C deposits are generally non-calcareous.

We sampled paleosols and other fluvial, palustrine and lacustrine sedimentary deposits of members B and C for analysis of long (>21 carbon) chained *n*-alkanes. These long chain *n*-alkanes are produced in the leaf waxes of the higher terrestrial plants and can be preserved for tens of millions of years in almost all soils and sediment types (Sachse et al., 2012). As the leaves source hydrogen from soil water, which in turn sources water from precipitation, thus the leaf wax *n*-alkane δD ($\delta\text{D}_{\text{wax}}$) can be correlated to rainfall δD ($\delta\text{D}_{\text{mw}}$). However, soil–water evaporation and transpiration (Sachse et al., 2012) during incorporation of soil water in leaf water makes the leaf water δD more positive than meteoric water; while biosynthetic fractionation in the process of formation of the leaf wax causes the leaf wax *n*-alkane δD to be more negative than the leaf water δD . The overall enrichment factor ($\epsilon_{\text{mw/wax}}$) is used to express the apparent fractionation between meteoric water and the $\delta\text{D}_{\text{wax}}$. Thus, meteoric water δD can be reconstructed from leaf wax *n*-alkane δD ($\delta\text{D}_{\text{wax}}$) when $\epsilon_{\text{mw/wax}}$ is known. Calibration of the overall enrichment factor ($\epsilon_{\text{mw/wax}}$) from soil water to leaf wax, however, remains a challenge in reconstruction of paleoclimate. The $\epsilon_{\text{mw/wax}}$ can vary among species and climate settings (Polissar and Freeman, 2010) due to species-dependent variations in evapotranspiration rates from leaves prior to biosynthesis (Sachse et al., 2012, Kahmen et al., 2013). This makes it difficult to choose an appropriate enrichment factor when the vegetation type is

unknown (Polissar and Freeman, 2010; Tipple and Pagani, 2013). We take these factors into account in evaluating paleoelevation from δD_{wax} . In this study, we have the unique opportunity to use our pollen data to calculate δD_{mw} using the representative $\epsilon_{\text{mw/wax}}$ of the vegetation assemblage indicated by the pollen and comparing the δD_{mw} values derived from leaf wax with the $\delta^{18}\text{O}_{\text{mw}}$ values derived from paleosol carbonates. While, soil water evaporation could impart some enrichment effect in δD_{wax} and $\delta^{18}\text{O}$ values of pedogenic carbonates, the degree of evaporative enrichment is not as severe as the bias associated with evaporative enrichment of ^2H and ^{18}O in lacustrine and palustrine water, and provides a significant advantage of using δD_{wax} over lacustrine and palustrine carbonates.

5. Methods

We sampled pollen from organic rich mudstone and analyzed them at the Smithsonian Tropical Research Institute, Panama following standard procedures (Traverse 2007).

The paleosol carbonates were sampled at a depth of $>1\text{m}$ from the paleosol top. Both paleosol and lacustrine/palustrine carbonates were crushed, (or microdrilled in the cases where minor sparite was present), pretreated and analyzed for $\delta^{18}\text{O}_\text{c}$ at the University of Rochester. A subset of carbonate samples were analyzed for Δ_{47} at CalTech (after Eiler, 2007). Analytical errors for Δ_{47} were calculated from the standards that were analyzed over the same time period of sample analysis. For three analytical seasons, the 2σ error of standards was 0.019%, and we propagated this through the Ghosh et al., (2006a) equation to get the analytical uncertainty in $T\Delta_{47}$. As analyses started before the absolute reference frame (Dennis et al., 2011) was established, our samples are calibrated with reference to the Caltech intralab reference frame described by Huntington et al. (2009).

We sampled paleosol, sand matrix of conglomerate and lacustrine/palustrine mudstone for leaf wax. We collected these samples from beds with no modern vegetation and sampled at a >5 cm depth within the rock layers to avoid any potential modern day contamination from detritus on the outcrop surface. The leaf wax n-alkane extraction and analyses of the non-polar fraction were carried out at the University of Texas at Austin. The δD values reported for each sample is the average of the long odd chains (C27-33). The full details of the analytical methods are located in supplementary data section (Appendix C).

6. Diagenetic History

We examined analyzed carbonate samples for diagenesis as it can modify the original isotope composition, potentially resetting the $T\Delta_{47}$ and $\delta^{18}O_c$ signatures in carbonates (Garzzone et al., 2004; Huntington et al., 2015). Diagenetic resetting by mineral recrystallization results in higher $T\Delta_{47}$ and more negative $\delta^{18}O_c$. Recrystallization produces easily observable coarse crystals or spar. However, micro-scale recrystallization or solid state reordering can alter the $T\Delta_{47}$ without any significant change in micritic texture or $\delta^{18}O_c$ and $\delta^{13}C$ values (Huntington et al., 2015). To identify the diagenetic temperature, we analyzed Δ_{47} from a concretion (10PE55c) collected from the top of the Cerro Quirma section in Member B. As the analyzed $T\Delta_{47}$ of $36 \pm 3^\circ C$ and $\delta^{18}O_c$ and $\delta^{13}C$ (VPDB) values of -11.3‰ of -7.3‰ fall within the range shown by Cerro Quirma paleosols (Fig. D.1), we interpret that the concretion formed under a shallow burial condition. However, the two lowest paleosol samples from this section, 10PE40c and 10PE41c display temperatures (61°C to 41°C) higher than that observed at the Earth's surface (Fig. D.1.). The $\delta^{18}O_c$ and $\delta^{13}C$ values of these

two samples are found to be among the most negative in comparison to the other Member B paleosols (Fig. D.1), and thin section examination of 10PE40c ($T\Delta_{47}$ of 62°C) shows pervasive recrystallization (Fig. 4A). Based on these observations, we infer that the samples 10PE40c and 10PE41c, from the lowest part of Member B, experienced burial diagenesis. Assuming a standard geothermal gradient of $\sim 25^{\circ}\text{C}/\text{km}$ and a surface temperature of 20°C , these temperatures indicate that recrystallization took place at a depth of up to $\sim 1600\text{m}$. As the current combined thickness of members B and C is $\sim 1200\text{m}$, we infer that Member B was possibly thicker and experienced some erosion during the formation of the 9.1 ± 0.7 to 5.4 ± 1.0 Ma angular unconformity.

High temperatures ($\geq 40^{\circ}\text{C}$), correlated negative $\delta^{18}\text{O}_\text{c}$ (VPDB) and $\delta^{13}\text{C}$ (VPDB), and extensive recrystallization are not observed in the other samples from the Cerro Quirma section, and we infer that those paleosol carbonates did not undergo diagenetic alteration. Detailed petrographic examination of the limestone samples from the San Genaro, Kanamarka and Hector Tejada sections (Fig. 4B-F) show that they suffered some amount of diagenesis associated with the infilling of primary porosity in the rock, but it was not pervasive. We micro-sampled the unaltered primary micritic parts of these limestones with a dental drill, avoiding the recrystallized areas to determine the original $T\Delta_{47}$.

7. Results and Paleoelevation Calculation

7.1. Pollen Assemblage

Pollen analysis (Fig. 5, Table D.1) shows that the middle part of Member B is rich in organic matter and palynomorphs, dominated by *Podocarpus* (mean abundance = $32.8 \pm 17\%$, $N=7$, Table D.1). In contrast, Member C samples have lower overall pollen

content and *Podocarpus* proportion (mean abundance=9.6±12.9%, N=4, Table D.1; statistical significant difference – t-test, $p<0.03$). Grasses and herbs are more abundant in Member C (mean= 47.1±23%, N=4) than in Member B (mean=20.8±12.1%, N=7), although the difference is not significant (t-test, $p<0.1$).

There are six extant species of *Podocarpus* in Peru that are mostly found at mid elevations of the eastern slope of modern day Andes (Fig. D.2) and are absent from the western slope of the Andes in central Peru due to low precipitation (Kuentz et al., 2007). The genus *Podocarpus* inhabits areas with a mean elevation of 2.1±0.9 km, a mean annual temperature of 17.2±4.5°C, and an annual precipitation of 135±80 cm (Fig. D.3.). The modern high elevation (~4 km) Descanso–Yauri basin is characterized by low mean annual temperatures (~8–9°C), low annual precipitation (50–80 cm), and it is dominated by low-lying vegetation including grasses (*Calamagrostis*, *Festuca* and *Stipa*), shrubs (Poaceae *Aciachne*, and Caryophyllaceae *Pycnophyllum*), several Asteraceae (e.g. *Ambrosia*, *Senecio*) and Rosaceae (e.g. *Polylepis*) (Kuentz et al., 2007, Nunez 2011). The dominant presence of *Podocarpus* in Member B (Fig. 5 and Table D.1) indicates that the region was warmer (~10°C) and more humid (~50–100 cm/yr) than modern conditions, suggesting that Member B accumulated at a significantly lower elevation (~2.1±0.9 km). The pollen record of Member C (Fig. 5 and Table D.1), in contrast, is dominated by low-lying vegetation similar to modern high elevation vegetation, suggesting that Member C accumulated at an elevation and/or climate similar to modern conditions. The higher temperature and humidity in Member B, leading to a higher *Podocarpus* abundance, can be explained by a lower elevation of the Altiplano, or by the Altiplano being much warmer and more humid than modern conditions, allowing

Podocarpus to expand its range upward by 2 km. However, most of the global climate models published for the region do not support the latter scenario. In addition, for large trees like *Podocarpus* to inhabit 4,000 m habitats would require major physiological modifications that are unlikely. It also could be argued that *Podocarpus* pollen could be transported by wind from lower elevations into higher elevations. However, the significant difference in *Podocarpus* abundance between members B and C would not be expected by wind-transport alone if both members were deposited at the same elevation. Likewise, the dominance of grasses and herbs in Member C supports the inference that Member C paleoelevation was similar to modern elevation, where these plant types are dominant.

7.2. Stable Isotopes

7.2.1. $T\Delta_{47}$

The $T\Delta_{47}$ values from all Member B samples show a large range from 38 to 22°C (Fig. 6A, Table D.2). However, paleosol and lacustrine or palustrine carbonates can precipitate under different temperature conditions within the same climate setting, and it is important to consider their values separately. Studies of Holocene soils show that solar insolation can produce warming of surface soils in excess of ~10°C of ambient air temperatures (Quade et al., 2013), with greater soil warming in regions of less plant cover. Furthermore, soil carbonates tend to precipitate when soils are drying out (Breecker et al., 2009) and they likely form during the austral summer after rainfall events in the Altiplano (Garzione et al., 2014). Thus, the $T\Delta_{47}$ signature of paleosol nodules could potentially reflect a warm season bias. Similar warm season bias has also been observed in lacustrine carbonates (Hren and Sheldon, 2012). However, the magnitude of total warming in comparison to air temperature is subdued in comparison to soil as the effect of insolation is smaller

in lakes or marshes (i.e., palustrine settings). We infer that the analyzed TA_{47} values of the lacustrine and palustrine carbonates reflect the warmest average monthly air temperature (WAMT) when lakes are most productive. To account for additional solar ground warming in paleosol carbonates, we reconstructed the WAMT from these samples. We use the empirical relation (equation 4 of Quade et al., 2013) for $\geq 1m$ deep soils that relates TA_{47} of pedogenic carbonate to WAMT as almost all of our paleosol samples came from $\geq 1m$ depth. The calculations are done by a bootstrap Monte Carlo method, propagating both the analytical errors and errors in regression. We also calculated the mean annual air temperature (MAAT) from all the samples (excluding the diagenetically altered ones). For the lacustrine/palustrine carbonates, we employed the transfer function (equation 2 of Hren and Sheldon, 2012) between Mean Annual Lake Surface Temperature (MALST) and MAAT. For the paleosols, we used the empirical relation for $\geq 1m$ deep soils that relates TA_{47} of pedogenic carbonate to MAAT (equation 3, Quade et al., 2013). The reconstructed MAAT from all Member B carbonates have an average of $25 \pm 8^\circ C$ (2σ) (Fig. 6B, Table D.2). Even with the large scatter in the MAAT data, both the average and the coolest MAAT of $21 \pm 2^\circ C$ (2σ) (Table D.2) are substantially warmer than the modern day MAAT of $8-9^\circ C$, indicating a low paleoelevation during deposition of Member B.

7.2.2. Carbonate $\delta^{18}O_c$

We reconstruct the meteoric water isotopic value ($\delta^{18}O_{mwc}$) using the isotopic composition of carbonates ($\delta^{18}O_c$) (Fig. 6C, Table D.2) and TA_{47} . The average $\delta^{18}O_{mwc}$ (VSMOW) value from all Member B carbonates is $-9.0 \pm 4.2 \text{ ‰}$ (2σ) (Fig. 6D, Table D.2). However, the $\delta^{18}O_{mwc}$ (VSMOW) reconstructed from lacustrine/palustrine carbonates shows a larger scatter in comparison to paleosol carbonates. This scatter possibly arises from the influence of high elevation run-off in the most negative values and

evaporative enrichment in the most positive values. To minimize these uncertainties, we only use $\delta^{18}\text{O}_{\text{mwc}}$ (VSMOW) reconstructed from paleosol $\delta^{18}\text{O}_{\text{c}}$ (VPDB) with known carbonate formation temperature ($T_{\Delta_{47}}$) for paleoelevation calculation. The sampling depth of $\geq 1\text{m}$ and the high rainfall amount ($\sim 0.6\text{--}1.2\text{ m/yr}$) in the northern Altiplano minimizes the effect of evaporative enrichment in $\delta^{18}\text{O}_{\text{c}}$ (Quade et al., 2007). Reconstructed meteoric water from paleosols show (Fig. 6D, Table D.2) an average $\delta^{18}\text{O}_{\text{mwc}}$ (VSMOW) of $-8.3 \pm 2.0\text{‰}$ (2σ), ~ 6 to 8‰ more positive than modern high elevation surface waters, which average -15 to -16‰ (Gonfiantini et al., 2001; Bershaw et al., 2010; Insel et al., 2012).

7.2.3. Leaf Wax *n*-alkane δD

δD_{wax} provides a crucial estimate of paleoelevation changes in the Descanso–Yauri basin because it is the only geochemical proxy that spans both members B and C. To mitigate concerns that the lacustrine/palustrine samples (San Miguel and Checca) of Member B might have a different source from the soil samples (Cerro Quirma), we compared the means of δD_{wax} (via an independent samples t-test), and did not find a statistically significant difference ($p\text{-value} = 0.63$). Additionally, the Carbon Preference Index (CPI) of our samples range from 2.9 to 10.6 (Table D.3), suggesting that they likely did not undergo degradation (Eglinton and Hamilton, 1967).

Measured δD_{wax} values show a large shift from $-143.4 \pm 12.8\text{‰}$ (2σ) in Member B to $-209.2 \pm 21.1\text{‰}$ (2σ) in Member C (Fig. 7A, Table D.4), indicating a major change in precipitation δD . Within Member B, the δD_{wax} values show a shift to more negative values from the lowermost ($-144.4 \pm 1.3\text{‰}$, 2σ) part of the Member to the lower–middle part (-168.6‰) and gradual shift back to more positive values from the middle (-156.4‰) to

upper (-142.7‰) part. This shift possibly correlates with climate variation from dry to wet and then progressively drier condition.

To interpret paleoelevation from the isotopic signature, we considered associated changes in vegetation and aridity, as indicated by pollen assemblage. Increasing aridity leads to higher fractionation between precipitation and leaf water, resulting in enriched δD in the latter (Kahmen et al., 2013), while vegetation change to more drought tolerant plants can result in higher biosynthetic fractionation. A recent study (Gao et al., 2014), however, suggests that these two opposite changes offset each other and result in an almost uniform $\epsilon_{mw/wax}$ from humid subtropical to arid lowland conditions. Thus, we infer that use of vegetation-specific $\epsilon_{mw/wax}$ reduces uncertainty associated with vegetation change and provides a more conservative estimate of surface uplift. The vegetation assemblage of members B and C are broadly equivalent to modern North American montane gymnosperms and Páramo vegetation, and we use the published $\epsilon_{mw/wax}$ values associated with these two vegetation types (i.e., montane $\epsilon_{mw/wax} = -98 \pm 6‰$ and Páramo $\epsilon_{mw/wax} = -120 \pm 9‰$; Polissar and Freeman, 2010) to reconstruct precipitation δD (δD_{mwwax}). We then used the local meteoric water line for the Andes (Gonfiantini et al. (2001) (equation 5) (Fig. 7B and 7C) to convert δD_{mwwax} to $\delta^{18}O$ of precipitation ($\delta^{18}O_{mwwax}$).

The average reconstructed δD_{mwwax} (and $\delta^{18}O_{mwwax}$) values of members B and C (Fig. 8) are $-50.3 \pm 14.2‰$ (and $8.6 \pm 1.8‰$) (2σ) and $-101.4 \pm 24.0‰$ (and $-14.6 \pm 3.0‰$) (2σ), similar to modern foreland and high elevation surface water δD (and $\delta^{18}O_{mw}$) values (Gonfiantini et al., 2001; Bershaw et al., 2010; Insel et al., 2012; Saylor and Horton, 2014). Furthermore, the reconstructed $\delta^{18}O_{mwwax}$ value for Member B is also very

similar to the $\delta^{18}\text{O}_{\text{mw}}$ reconstructed from pedogenic carbonates ($-8.3 \pm 2.0\text{‰}$, 2σ). While we acknowledge that uncertainty exists over the timescale of production and integration of leaf wax (Tippie and Pagani, 2013) and the local d-excess relationship (Fiorella et al., 2015), the conformity of our leaf-wax estimates with the pedogenic carbonates suggests that they record similar meteoric water signals.

7.3. Paleoelevation Calculation from Stable Isotope Data

7.3.1. Paleotemperature

To estimate the paleoelevation of Member B from the $\text{T}\Delta_{47}$, we first consider the possible sources of additional temperature change not associated with adiabatic cooling of an ascending air mass. The depositional age of $\sim 18\text{--}9$ Ma spans the $17\text{--}15$ Ma middle Miocene climate optimum (Zachos et al., 2001). This could contribute $4\text{--}5^\circ\text{C}$ of additional warming in the lower Member B average MAAT of $26 \pm 6^\circ\text{C}$ (2σ). However, it cannot account for the $\sim 15^\circ\text{C}$ difference between modern MAAT and the average MAAT of $23 \pm 4^\circ\text{C}$ (2σ) from the uppermost Member B limestones of Hector Tejada, Kanamarka and San Genaro sections. Higher $p\text{CO}_2$ level in the late Miocene, compared to Pliocene-Pleistocene, could account for 2.8°C warming relative to modern temperature (Poulsen and Jeffery, 2011), just 20% of the entire shift in temperature from upper Member B to present.

The modern day air temperature lapse rate of $-5^\circ\text{C}/\text{km}$ in the central Andes (Gonfiantini et al., 2001) suggests a paleoelevation of ~ 1 km based on the $\sim 15^\circ\text{C}$ warmer temperatures observed in the top of Member B. However, this lapse rate does not consider additional warming under a lower Andes scenario (Ehlers and Poulsen, 2009). We defined a new temperature–elevation regression accounting for this effect, to avoid under-prediction of paleoelevation. We used average December-January-February (DJF) temperatures from

72 weather stations (weatherbase.com, 2016) situated in the northern Central Andean plateau (10 to 16°S), at elevations of 4497m to 137m. We added temperature correction factors (+0.6°C at 50%, +3.4°C at 25% and +6.6°C at 0% of modern elevation; i.e., 2km, 1km, 0km respectively) to data from all the stations situated at 2 km (i.e., 50% of modern height) or lower based on Ehlers and Poulsen, (2009). We used linear interpolation to find the required correction at intermediate elevations. Since additional temperature change due to uplift is insignificant above 2 km elevations, no correction was applied. Our uplift corrected DJF air temperature–elevation regression of $y = -0.15x + 5.18$ (Fig. 8A) yields a paleoelevation of 0.9 ± 0.8 km (2σ) through a bootstrap Monte Carlo method from the average TA_{47} of $27.9 \pm 8.8^\circ\text{C}$ (2σ) in 9.1 ± 0.7 Ma limestones. Uncertainties in this method stem from the absence of a baseline temperature by which to compare Member B paleotemperature (e.g., Garzione et al., 2014) and potentially smaller temperature differences at the latitude of the study area compared to the modeled average temperature difference (Ehlers and Poulsen, 2009). However, this temperature correction ensures that the paleoelevation estimate for lower elevation sites is biased toward maximum estimates and therefore results in a conservative (i.e., minimum) estimate of the magnitude of surface uplift, more so because the warming effect is muted north of 20°S, in our study area (Ehlers and Poulsen, 2009). While this paleoelevation estimate is lower than that estimated from pollen analysis, both estimates fall within uncertainty of each other and indicate low paleoelevation during the deposition of Member B.

The TA_{47} of 26°C from the Member C carbonate sample is difficult to understand in the context of other proxy data from Member C. Paleoelevation calculations using the previously described process yield an elevation of 1.2 ± 0.4 km (2σ), similar to Member B

and notably different from modern elevation conditions suggested by the pollen assemblage. While this calcareous lacustrine mudstone does not show evidence of diagenetic alteration, the temperature is anomalously warm compared to both pollen and fossil assemblages collected from the same location. Sparse small fossil leaves with serrated edges (Fig. D.4) found in this section are tentatively identified as *Polylepis* leaves, from the Rosaceae family. These are commonly found in the modern high Andes (Gregory-Wodzicki, 1998) and are characteristic of cool mean annual air temperature (~10°C). Furthermore, leaf wax data (discussed in 7.3.2) also indicate cooler conditions during deposition of Member C. Given the overall lack of carbonate in Member C lacustrine deposits, we suspect that the Cerro Pucara carbonate sample reflects carbonate cement precipitated during a lake drying event. Under these conditions the paleotemperature would reflect summer temperature that was subjected to solar ground warming in the absence of significant plant cover to produced an additional ~10 to 15°C warming (Passey et al., 2010; Quade et al., 2013) above the modern (or similar) warm season average temperature of 9–10°C.

7.3.2. *Meteoric Water*

We computed paleoelevation from reconstructed $\delta^{18}\text{O}_{\text{mw}}$ by the variable lapse rate model described by Insel et al. (2012). Initiation of low level jet (LLJ) resulting in increased precipitation was proposed along the eastern flank of the plateau (Poulsen et al., 2010; Insel et al., 2012) at threshold (~50 to 75% of modern) elevation condition. An amplification of the LLJ was proposed at an elevation between 75 to 100%. Therefore, use of the variable lapse rate model considers the influence of those climate changes in the paleoelevation calculation. We also calculated paleoelevation from the Rayleigh

distillation model (Rowley and Garzione, 2007) of an ascending vapor mass. We ran this model with 81% relative humidity, a -6.0‰ 0 m elevation starting value (Gonfiantini et al., 2001) and three temperatures conditions (calculated from weather stations at <200 m elevation, between 14°S–15°S; weatherbase.com) to encompass the full range of possible starting conditions before vapor mass ascent: 1. WAMT modified for additional warming in lower Andes scenario (33°C), 2. modern WAMT (28°C) and 3. modern MAAT (25°C). We use the WAMT parameters to calculate paleoelevation from $\delta^{18}\text{O}_{\text{mwc}}$, as these carbonates likely formed during summer conditions. We use the MAAT parameter to calculate elevation from $\delta^{18}\text{O}_{\text{mwwax}}$, as the time of leaf wax formation is less constrained. We note that consideration of the warm extreme climate change scenario (scenario 1) ensures a conservative paleoelevation estimate and gives us a lower bound on the total magnitude of surface uplift.

The average $\delta^{18}\text{O}_{\text{mwc}}$ of $-8.3 \pm 2.0\text{‰}$ (2σ) of Member B defines an elevation range of 1.7 ± 1.1 km (Fig. 8B) with the Insel et al. (2012) model and elevations of 2.0 ± 1.6 km (2σ) and 1.7 ± 1.4 km (2σ) with Rayleigh distillation models 1 and 2. The average $\delta^{18}\text{O}_{\text{mwwax}}$ of $-8.6 \pm 1.8\text{‰}$ (2σ) and $-14.6 \pm 3.0\text{‰}$ in members B and C yield paleoelevations of 1.7 ± 1.0 km and 4.2 ± 0.9 km (2σ) using model 3. While this model assumes modern climate conditions, we note that the reconstructed Member B paleoelevation falls within the range (0.9 ± 0.8 km from $T\Delta_{47}$ to 2.1 ± 0.9 km from pollen) described from other climate proxies. We also note that the 4 different approaches for calculating paleoelevations from $\delta^{18}\text{O}_{\text{mw}}$ reconstructed from paleosol carbonates and leaf waxes yield a narrow range of average Member B paleoelevation estimates of 1.7 km to 2.0 km. The -6‰ change in $\delta^{18}\text{O}_{\text{mwwax}}$ from members B to C is also equivalent to at least ~2km

surface uplift, if we apply the $\sim 3\text{‰}/\text{km}$ lapse rate described at high elevation (Insel et al., 2012). Furthermore, modern *n*-alkane (C_{29}) values, $-206 \pm 11\text{‰}$ and $-227 \pm 14\text{‰}$ from high elevation (4600 and 4300 m) lakes in Bolivia (Polissar and Freeman, 2010) are comparable with average $-206 \pm 14\text{‰}$ (2σ) from Member C. Based on this isotopic evidence, we infer that the Descanso-Yauri basin reached its current elevation by 5.4 ± 1.0 Ma, before the deposition of Member C.

8. Discussion

All 4 proxies from Member B (pollen, $\delta^{18}\text{O}_{\text{mwax}}$, $\delta^{18}\text{O}_{\text{mwc}}$, and TL_{47}) result in paleoelevation estimates that range between 0.9 ± 0.8 km and 2.1 ± 0.9 km, while two different proxies (pollen and $\delta^{18}\text{O}_{\text{mwax}}$) for Member C indicate paleoelevations that are the same as modern elevation. The multiproxy based paleoclimate history from the northernmost Altiplano plateau points to at least 2 ± 1 km of surface uplift between 9.1 ± 0.7 and 5.4 ± 1.0 Ma that established the modern elevation of the Altiplano by 5.4 ± 1.0 Ma. This result generally agrees with the previous genetic divergence based estimate of 2.0–2.5 km elevation by middle late Miocene (Picard et al., 2008). Regionally the timing is coeval or slightly younger than the 10–6 Ma surface uplift event (Fig. 9) described from the central plateau (Garzione et al., 2008). The timing of the regional north–central Altiplano surface uplift also corresponds well with the reported increased incision from 9–5 Ma in the Western Cordillera (Schildgen et al., 2009).

Comparison with the 16–9 Ma uplift pulse in the southern Altiplano (Garzione et al., 2014) shows that surface uplift in the plateau postdated surface uplift in south. Paleoelevation studies from the plateau flanks (Leier et al., 2013; Saylor and Horton, 2014) indicate that a rapid pulse of surface uplift established moderate paleoelevations in

the Eastern Cordillera of ~2.5 km in the central part of the plateau between 24 and 17 Ma) and modern elevations in the Western (by 16 Ma) Cordilleras, and uplift was rapid, similar to rates in the Altiplano (Fig. 9). Such rapid and diachronous pulses of surface uplift indicate that the geodynamic mechanisms responsible for elevating the plateau to its modern condition acted in a local scale in the Central Andes.

To evaluate the role of the local crustal shortening we examined the deformation history of the Descanso–Yauri basin. Basin bounding thrusts were active prior to 9.1 ± 0.7 Ma (Kar et al., 2015), evidenced by deformed deposits of the members A and B. The deformation regime also changed from compressional to extensional post 9 ± 0.7 Ma, forming normal faults that offset the limestones capping Member B. The surface uplift pulse was thus decoupled from deformation and postdated the local crustal shortening and was concurrent with regional extension. Major regional crustal shortening took place even earlier, at 45 to 20 Ma in the central Altiplano and Eastern Cordillera (McQuarrie et al., 2005), and at ~50 Ma in the Peruvian Altiplano (Chapman et al., 2015).

The estimated crustal shortening of 115 km (Chapman et al., 2015) to 123 km (Gotberg et al., 2010) in the northern Altiplano is much lower than the ~300 km crustal shortening estimated from the Central Altiplano. It is noteworthy that the shortening estimates from the northernmost Altiplano fall short of the 240 to 300 km of crustal shortening that is required to account for the 60–70 km thick modern crust (eg. Gotberg et al., 2010). The gap between the required and estimated crustal budget could be explained by additional crustal shortening and thickening hidden in the Western Cordilleran magmatic arc that was overprinted by magmatism or by addition

of crustal material from other parts of the Central Andean plateau via crustal flow. Recent crustal volume modeling from the Altiplano and Eastern Cordillera (Eichelberger et al., 2015) proposed that excess crustal volume might have been sourced from the lower crust in the Subandean zone or by lower crustal flow to the northern portion of the plateau.

The average rate of surface uplift in the northernmost plateau also points to a mechanism that produces rapid elevation change. The calculated rate of 0.5–0.8 km/Myr is faster than the highest rate of surface uplift (0.25 km/Myr) predicted by upper crustal shortening from the Central Altiplano (McQuarrie et al., 2005). We infer that processes such as lower crustal flow (Husson and Sempere, 2003) or lower lithosphere removal (Garzione et al., 2006; Leier et al., 2013; Garzione et al., 2014) have played an important role in plateau formation as they can produce such rapid surface uplift. The switch from compressional to extensional deformation in the Descanso-Yauri basin between 9.1 ± 0.7 and 5.4 ± 1.0 Ma, during the same time period as surface uplift, is also consistent with the removal of a lithospheric root (Göğüş and Pysklywec, 2008). Further evidence of lower lithosphere removal in the Altiplano includes: the predominantly felsic nature of the 55–70 km thick crust, indicating the absence of mafic lower lithosphere (Beck and Zandt, 2002); high $^3\text{He}/^4\text{He}$ ratios of hydrothermal fluids representative of mantle asthenosphere derived magmas (Hoke et al., 1994); and the influx of primitive magmas in Miocene–Pliocene time (Chapman et al., 2015). More direct evidence of crustal removal comes from the recent observation of a vertically elongate fast anomaly adjacent to lower velocity anomalies, interpreted as a blocks (or drips) of delaminating (or convectively removed) lithosphere (Scire et al.,

2015).

9. Conclusion

The multiproxy based paleoclimate reconstructions from the Descanso–Yauri basin suggest that the northernmost plateau remained at 0.9 ± 0.8 km (2σ) to 2.1 ± 0.9 km between ~ 18 and 9 Ma. This period was characterized by temperatures that were $\sim 15^\circ\text{C}$ warmer than modern ($23 \pm 4^\circ\text{C}$, 2σ) (Table 1), abundance of *Podocarpus*, a low montane elevation vegetation, and more positive $\delta^{18}\text{O}_{\text{mw}}$ ($-8.3 \pm 2.0\text{‰}$ to $-8.6 \pm 1.8\text{‰}$). A 2–3 km surface uplift pulse after 9.1 ± 0.7 Ma established the modern elevation by 5.4 ± 1.0 Ma. Based on this new 9.1–5.4 Ma surface uplift constraint from the northernmost plateau, combined with the observations that surface uplift occurred in the south between 16–9 Ma (Garzione et al., 2014) and in the north–center between 10–6 Ma (Ghosh et al., 2006b; Garzione et al., 2006) we infer that a regional geodynamic phenomenon swept or stepped from south to north (Fig. 9) in the Altiplano. As this surface uplift was decoupled from the local crustal shortening, with a rate higher than that has been suggested from crustal shortening, we suggest that lower lithospheric removal initiated in the southern Altiplano and drove the northward flow of weak middle-lower crust (e.g., Ward et al., 2013), resulting in the lower lithosphere removal and progressive surface uplift of the central and northern Altiplano.

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837 **Figure Captions**

838 Fig. 1.A. Topographic map of the Central Andes showing the Altiplano plateau, flanked
839 to the West by the Western Cordillera magmatic arc and to the east by the Eastern
840 Cordillera fold-thrust belt. Locations of the previous paleoclimate studies in the Altiplano

plateau and the plateau flank are shown. Timing of surface uplift/climate change reported by these previous studies are shown in legend. If only the low elevation estimate is known, then it is shown in italics. Location of this study (Descanso–Yauri basin) at the northernmost plateau in southern Peru is shown with an outline of basin map. **B.** Geologic map of the Descanso–Yauri basin showing members A, B and C and sampling locations and type of samples collected in each location. Member A was not sampled, as it is generally non-calcareous and organic-poor. The map is modified from Cerpa and Meza (2001) based on field observation. Note that all the Cenozoic volcanic sections are mapped as one unit, but they span a wide age range.

Fig. 2. **A.** Generalized lithologic column of members A, B and C. Inferred stratigraphic position of the sampling localities and sample types are shown with respect to the composite lithologic column. Member A is the coarsest member, dominated by sandstone and conglomerate. Member B is dominated by mudstone and is the only member with paleosol carbonates. Member C is dominated by diatomite and mudstone with some sandstone. **B.** The Member A picture shows vertical beds of clast supported conglomerate and sandstone. Member B image shows a paleosol from the Cerro Quirma section. The arrow points to the top of the paleosol, with B_k horizon present at the base of the hammer. It is ~40 cm thick and contains carbonate nodules ~0.5 to 1 cm in diameter. The angular unconformity between Member B and Member C is observed at Accocunca (refer to Fig. 1B) and is marked by a red dashed line on the picture. The black dashed lines marking Member B beds show truncation against the unconformity surface. The green dashed lines above the unconformity surface trace Member C beds, parallel to the unconformity

surface. The Member C picture shows the typical Member C deposit of horizontally laminated ash rich mudstones.

Fig. 3. Weighted mean age distribution plots of 12MN54t, 12SG19t and 12YR07t and histogram and probability density function of the analyzed zircon grain U-Pb ages of 13AC71t and 13CP97s. In the weighted mean age distribution plots red boxes mark the zircon analyses that were included in the age, whereas the blue boxes show results that are not included. Box heights represent 2σ uncertainty. **A.** 12MN54t: The weighted mean U-Pb age of 12.3 ± 1.0 Ma (2σ) comes from the youngest coherent cluster of 9 out of 12 analyzed zircon grains. This weighted mean age excludes one crystal with a 4.7 ± 0.5 Ma age having a measured U concentration much higher than the zircon crystals used to calculate the weighted mean age (~ 1500 ppm) and two significantly older crystals that yielded a 25.8 ± 1.7 Ma and a 248.2 ± 3.5 Ma age. **B.** 12SG19t: The weighted mean age of 9.1 ± 0.7 Ma (2σ) is defined by the youngest coherent cluster of 12 out of 20 analyzed zircon grains. The older ages ranging from 12.7 Ma to 26.5 Ma are interpreted as inherited crystals and were therefore excluded from the weighted mean age. One additional data point (marked by hatched lines) was excluded, from the TuffZirc Age based on Ken Ludwig's algorithm in Isoplot (Ludwig, 2003). **C.** 12YR07t: The weighted mean age of 4.8 ± 1.1 Ma (2σ) is defined from the coherent age cluster of 18 of the 18 analyzed zircons. **D.** 13AC71t and **E.** 13CP97s: Results over the 0–50 Ma timespan are shown as the focus is on the youngest zircon ages (n represents number of analyzed grains), bin span of each histogram is 2 Myr.

Fig. 4. Photomicrographs of examined thin sections, highlighting the diagenetic features observed in the most altered samples and general composition of the same samples (note the different magnification indicated by scale). **A.** The Cerro Quirma concretion sample (10PE40c) shows large recrystallized zones. **B.** The San Genaro limestone (12SG22c) shows presence of vugs with calcite recrystallization around the rim. **C.** The overall composition (70 to 80%) of the same sample (12SG22c) is micritic. **D.** The Kanamarka limestone (12KM57c) shows veins with recrystallization. **E.** The overall composition of the same sample (12KM57c) shows the presence of vugs and pockets of recrystallization but the dominant constituent (~60%) is unaltered micrite. **F.** The Hector Tejada limestone (12HT134c) shows some recrystallization, but the texture is dominantly micritic (~60%).

Fig. 5. Palynomorph abundances of eleven samples, seven from Member B and four from Member C, of the main palynomorph groups found in this study, including Podocarpus, other arboreal taxa, shrubs and herbs. Plot illustrates the raw counts rather than percentages. There is a significantly decrease in the proportion of Podocarpus pollen from member B to C samples. Samples belonging to a Member are not in stratigraphic order (overall members B and C are shown in order with Member C overlying member B) and represent replicates of each member. Total count refers to the total number of palynomorphs counted per sample.

Fig. 6. Isotopic data from carbonate samples plotted against the inferred stratigraphic height. Error bars show 1 σ error. **A.** Plot of $T\Delta_{47}$ (°C) temperatures. **B.** Plot of MAAT (°C) inferred from $T\Delta_{47}$ (°C). MAAT is not shown for diagenetically altered samples as their

$T_{\Delta_{47}}$ (°C) do not represent carbonate formation temperature. **C.** Plot of $\delta^{18}\text{O}_c$ (VPDB) from paleosol, palustrine and lacustrine carbonates. **D.** Plot of $\delta^{18}\text{O}_{\text{mwc}}$ (VSMOW) values reconstructed from $\delta^{18}\text{O}_c$ (VPDB).

Fig. 7. Isotopic data from leaf wax *n*-alkanes. The error bars show 1 σ error. **A.** Plot of the analyzed $\delta\text{D}_{\text{wax}}$ (VSMOW) values. Note the $\sim$ -66‰ negative shift between the average $\delta\text{D}_{\text{wax}}$ (VSMOW) of members B and C. **B.** Plot of the $\delta\text{D}_{\text{mw-wax}}$ reconstructed from $\delta\text{D}_{\text{wax}}$ (VSMOW). Use of different enrichment factors $\epsilon_{\text{wax/p}}$ (-120‰ for Member C and -98‰ for Member B) decreases the amount of negative shift from Member B to Member C, observed in the $\delta\text{D}_{\text{wax}}$ (VSMOW) values, but still shows a -51‰ change from average Member B to Member C. **C.** Plot of $\delta^{18}\text{O}_{\text{mw-wax}}$ (VSMOW) reconstructed from $\delta\text{D}_{\text{mw-wax}}$ (VSMOW) by the local meteoric water relation described by Gonfiantini et al. (2001).

Fig. 8. **A.** Elevation and WAMT relationship defined from weather station data (www.weatherbase.com), corrected for extra warming below 2 km. **B.** Calculation of the paleoelevation range of Members B by the variable lapse rate isotopic model (Insel et al., 2012). The stars show average $\delta^{18}\text{O}_{\text{mw}}$ (VSMOW) values at 25%, 50%, 75% and 100% of the modern elevation of 4 km. The black square marks the average Member B $\delta^{18}\text{O}_{\text{mwc}}$ value of -8.3 ± 2.0 ‰ (VSMOW). The vertical lines show the 2 σ uncertainty of the isotopic values and the reconstructed paleoelevation range of Member B shown at the bottom includes this uncertainty.

Fig. 9. Compilation of paleoelevation reconstructions along strike of the Central Andean Plateau (modified from Garzione et al., 2014), from south to north. The symbols show type of paleoclimate proxy used and the numbers refer to the corresponding reference. The reconstructed surface uplift in the Altiplano Plateau is referred by A.P. and by E.C and W.C in the Eastern and the Western Cordillera. The reconstructions show that both the Altiplano plateau and the plateau flank experienced rapid surface uplift in Miocene–Pliocene, but they are diachronous in nature. In the Altiplano proper, the surface uplift pulse propagated from south to north along the strike.

Table 1. Summary of paleoelevation estimates from members B and C, based on $T\Delta_{47}$ and $\delta^{18}O_{mw}$ (both $\delta^{18}O_{mwc}$ and $\delta^{18}O_{mwwax}$). The table shows average $T\Delta_{47}$, $\delta^{18}O_c$ and reconstructed $\delta^{18}O_{mwc}$, and $\delta^{18}O_{mwwax}$ from each sampled section. The sections are grouped according to the Member they belong to. The average isotopic values shown for each Member correspond to the final average isotopic values that we use for paleoelevation calculation.

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Table 1

Table 1

	Stable isotope Results*							Member C	
	Carbonate				Leaf Wax				
	TΔ ₄₇	Temp (°C)	δ ¹⁸ O _c	δ ¹⁸ O _{mwc}	δD _{wax}	δD _{mwwax}	δ ¹⁸ O _{mwwax}	Paleoelevation (in km)#	
	Extracted CO2	Ghosh2006	(vpdb)	(vsmow)	(vsmow)	(vsmow)	(vsmow)		
Member C					-209.2 ± 21.1	-101.4 ± 24.0	-14.6 ± 3	Rayleigh (3)§	δ ¹⁸ O _{mwwax}
Cerro Pucara	0.608 ± 0.01	26 ± 4	-11.8 ± 8.2	-12.6 ± 2.0	-204.0 ± 47.4	-95.4 ± 53.8	-13.8 ± 3.4	4.2 ± 0.9	
Los Angeles					-181.4 ± 3.3	-69.8 ± 7.6	-10.6 ± 1.0		δ ¹⁸ O _{mwc}
Versailles					-209.2 ± 1.5	-103.1 ± 1.7	-14.5 ± 1.2		
Hector Tejada					-206.2 ± 20.5	-98.0 ± 23.3	-14.1 ± 3.0		
Huano Huano					-213.4 ± 16.1	-106.1 ± 18.3	-15.1 ± 2.8		WAMT (°C)

	Stable Istope Results*							Member B	
	Carbonate				Leaf Wax				
	TΔ ₄₇	Temp (°C)	δ ¹⁸ O _c	δ ¹⁸ O _{mwc}	δD _{wax}	δD _{mwwax}	δ ¹⁸ O _{mwwax}	Paleoelevation (in km)#	
	Extracted CO2	Ghosh2006	(vpdb)	(vsmow)	(vsmow)	(vsmow)	(vsmow)		
Member B		28±9°C ¶		-8.3 ± 2.0	-143.4 ± 12.8	-50.3 ± 14.2	-8.6 ± 1.8	Rayleigh (3)§	δ ¹⁸ O _{mwwax}
Kanamarka	0.615 ± 0.014	32 ± 4	-14.9 ± 2.4	-11.1 ± 2.5				1.7 ± 1.0	
Hector Tejada	0.651 ± 0.011	24 ± 1	-15.1 ± 4.2	-12.9 ± 4.3				Insel(2012)	
San Genaro	0.614 ± 0.014	32 ± 6	-12.1 ± 5.8	-8.3 ± 5.8				1.7 ± 1.1	
San Genaro N†	0.602 ± 0.01	35 ± 5	-12.3 ± 2.8	-8.0 ± 3.0				Rayleigh (2)§	δ ¹⁸ O _{mwc}
Ocoruro			-14.9	-10.5				1.7 ± 1.4	
Checca	0.653 ± 0.01	24 ± 4	-7.5	-5.4	-142.7 ± 13.0	-49.6 ± 14.8	-8.1± 3.0	Rayleigh (1)§	
Accocunca N	0.635 ± 0.014	27 ± 5	-12.7 ± 3.2	-9.9 ± 3.1				2.0 ± 1.6	
San Miguel	0.661 ± 0.01	22 ± 4	-10.9 ± 1.5	-9.2 ± 1.5	-156.4 ± 5.2	-64.7 ± 6.0	-10.0± 0.5	Lapse rate#	
Cerro Quirmat†	0.608 ± 0.007	33 ± 7	-12.3 ± 1.0	-8.4 ± 1.0	-145.3 ± 22.0	-52.4 ± 24.4	-8.4 ± 3.0	0.9 ± 0.8	WAMT (°C)

* Error reported for each Member is the standard deviation (2σ) of the section averages and for each section, it is the standrd deviation (2σ) of sample values.

¶ The average from 9 Ma limestones (Kanamarka, Hector Tejada and San Genaro)

† The average value of sections consist of only paleosol δ¹⁸O_{mwc} as only those are used for paleoelevation calculation.

Errors in paleoelevation only reflect the scatter in the data and analytical uncertainties. They do not include uncertainty in the temperature lapse rate, which is unconstrained.

§ The numbers in parenthesis refers to the 3 different scenarios used to run the Rayleigh distillation models. Rayleigh (1) corresponds to WAMT modified for additional warming in lower Andes scenario (33°C), Rayleigh (2) corresponds to modern WAMT (28°C) and Rayleigh (3) corresponds to modern MAAT (25°C).