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Influence of Andean uplift on climate and paleoaltimetry estimates

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ABSTRACT

Recent elevation reconstructions of the Andean Plateau suggest a rapid 2.5 ± 1.0 km rise of the central Andes between ~10 and 6 Ma. This rapid rise has been attributed to a catastrophic removal of a dense lithospheric mantle root beneath the Andes. However, these findings are based on the assumption that climate did not change during deposition of paleoaltimetry proxies. Here we evaluate South American climate change due to Andean uplift and its influence on interpretations of plateau elevation from climate-sensitive paleoaltimetry data. A series of experiments are presented using the RegCM3 regional general circulation model (RCM) to characterize changes in Andean precipitation amount, surface temperature, and wind direction (vapor source) as a function of changing plateau elevation. Results indicate that South American and Andean climate changed significantly in response to plateau growth. More specifically, rising of the plateau results in up to a 900 mm increase in rainy season (December–January–February) precipitation over the plateau. Plateau uplift also results in a decrease in non-adiabatic surface temperature of up to 6.5 °C (in addition to adiabatic cooling directly related to elevation change through the lapse rate). Finally, the prevailing wind direction and the vapor source for precipitation switches from the South Pacific Ocean to the equatorial Atlantic as plateau elevation increases above ½–¾ of its present-day elevation. Taken together, these changes in paleoclimate would have substantially depleted the oxygen isotopic concentration of paleoprecipitation through the Cenozoic. Unless this climatic effect is taken into consideration, paleoaltimetry reconstructions based on stable isotope methods may overestimate the rapid rise of the Andes by up to several kilometers. We conclude that some or all of the apparent rapid rise of the Andean Plateau from paleoaltimetry data could be an artifact of large changes in paleoclimate.

1. Introduction

The elevation history of mountains provides insight into orogen erosion and sedimentation histories, orographic and regional climate change, and geodynamic models for orogen formation. In particular, the elevation history of large orogenic plateaus such as the Tibetan and Andean (Fig. 1A) Plateaus are essential for constraining geodynamic models of crustal and lithospheric mantle mass accumulation and removal during plateau formation. For example, contending models for Andean plateau formation call upon processes such as delamination of a dense root, crustal shortening and thickening, under-plating, ablative subduction, lower crustal flow, and coupling between lithospheric and atmospheric processes through erosion (Isacks, 1988; Wdowinski and Bock, 1994; Allmendinger and Gubbels, 1996; Allmendinger et al., 1997; McQuarrie, 2002; Lamb and Davis, 2003; Sobel et al., 2003; Sobolev and Babeyko, 2005; Willett and Pope, 2006; Molnar and Garzione, 2007; see also Oncken et al., 2006, and references therein). To distinguish between these dynamic models, observational constraints on the evolution of the Andean Plateau are needed. These constraints can come from quantifying: (1) the kinematic (deformation) and erosional history associated with the plateau and its marginal thrust belts, (2) the present-day crustal and lithospheric structure, and (3) the paleoelevation history of the plateau.

Of these, our understanding of Andean paleoelevation is arguably the most uncertain of existing geologic constraints, and yet offers important insights into plateau formation. Several methods have been used in the Andes and elsewhere to infer paleoelevation such as foliar physiognomy (Gregory-Wodzicki et al., 1998; Forest et al., 1999; Gregory-Wodzicki, 2000; Spicer et al., 2003), stomatal density of fossil leaves (McElwain, 2004), stable isotopes in authigenic minerals such as soil carbonate (Chamberlain and Poage, 2000), and vesicularity of basaltic flows (Sahagian and Maus, 1994). Of these methods, stable isotope paleoaltimetry is considered the most promising method due to the strong correlation between air temperature and precipitation $\delta^{18}O$ (Dansgard, 1964) and the perception that it is a direct method for inferring paleoelevation in soil carbonates (Quade et al., 2007). This method has been used to provide detailed records of paleo-altimetry for the North American Western Cordillera, Southern Alps of New Zealand, and Tibetan and Andean Plateaus (e.g., Chamberlain et al., 1999; Rowley et al., 2001).

Stable isotope paleoaltimetry techniques use modern climate to infer elevation change from data collected in the stratigraphic record. As acknowledged by many of its practitioners (Cerling and Quade, 1993; Chamberlain and Poage, 2000; Poage and Chamberlain, 2001;

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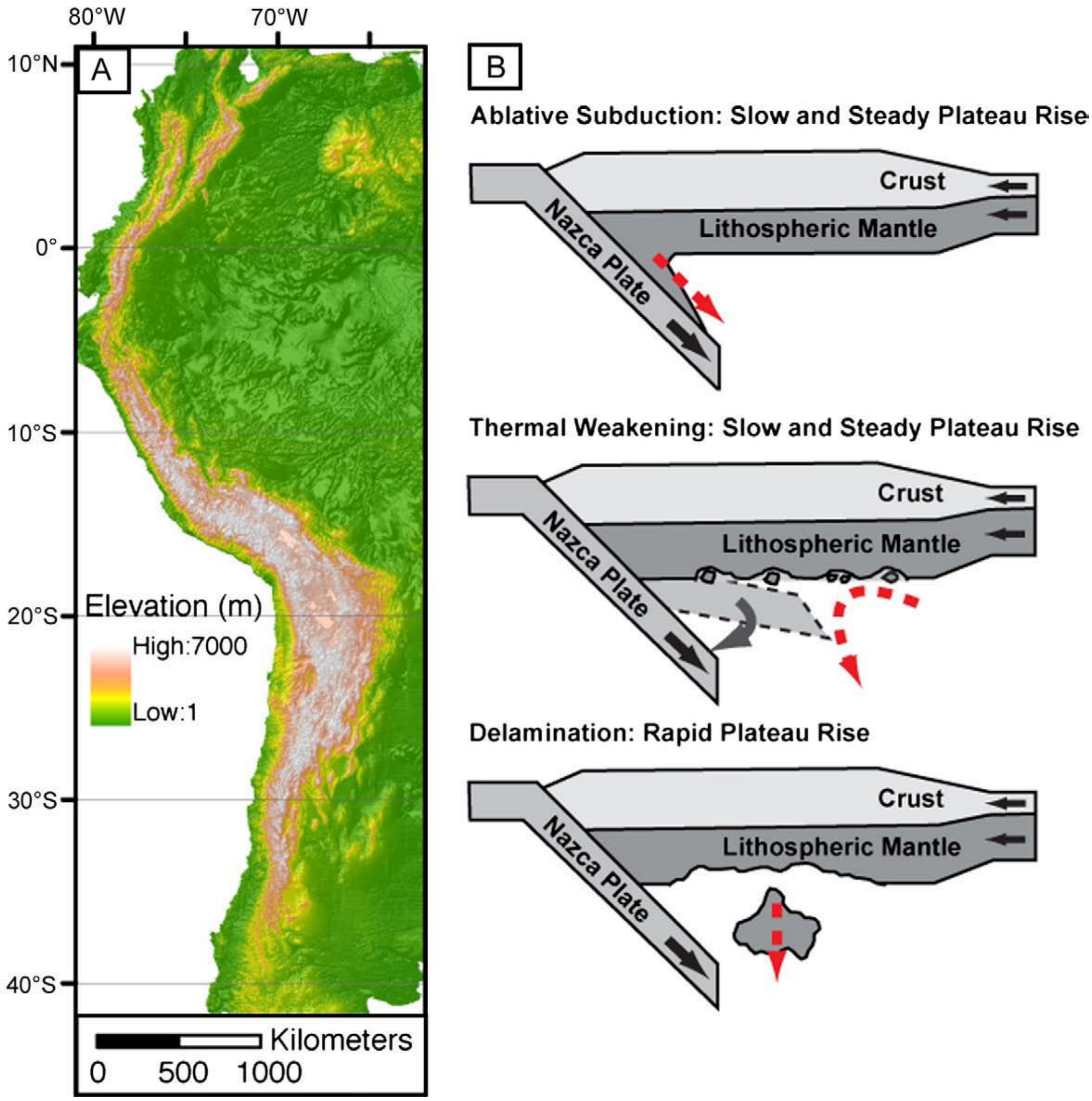

Fig. 1. (A) Modern topography of western South America. (B) Contending geodynamic models for plateau surface uplift. Light gray represents over thickened low-density crust. Red dashed lines indicate regions where higher-density lithospheric mantle material (dark gray) is removed, thereby resulting in surface uplift. Each of these models is described in detail in Section 2.

Blisniuk and Stern, 2005), this approach assumes that at the time of paleoaltimetry proxy deposition the past climate was analogous to the modern, despite dramatic changes in orography and regional climate that take place when orogenic plateaus form (Lenters and Cook, 1997; Ruddiman et al., 1997). In this study, we challenge this assumption and demonstrate that climate changes related to the rise of the Andean Plateau could significantly influence paleoaltimetry interpretations. Our approach is to use regional general circulation models (RCM) to quantify changes in South American climate solely as a function of variations in plateau elevation. As we show, dramatic changes in climate result from the rise of the Andes and warrant a re-evaluation of paleoelevation reconstructions using an isotope-tracking RCM.

2. Background

2.1. Geodynamics of Andean surface uplift

The Andes Mountains (Fig. 1A) were built primarily during the Cenozoic due to convergence between the Nazca and South American Plates. The Andean Plateau formed in the central portion of the orogen and is defined (Isacks, 1988) as the broad area of moderate relief and internal drainage with elevations >3 km. Over several decades significant progress has been made in reconstructing the deformation history of the Andean Plateau (e.g., Allmendinger et al., 1997 and references therein; Oncken et al., 2006). In the central Andes of Bolivia results from balanced cross sections, basin analysis, and geo- and thermochronology suggest (a) significant deformation since the late Eocene (~40 Ma) (Hartley et al., 2006; Sempere et al., 2006; Ege et al., 2007), and (b) development of the modern east–west width of the Andean Plateau, but unknown elevation, by 15–25 Ma (Horton et al., 2002; Barnes et al., 2006, 2008). The inferred elevation history of the plateau has until recently been less well constrained (e.g., Gregory-Wodzicki, 2000; Garzione et al., 2008; Picard et al., 2008) and depends on several factors, the most important being the assumed temporal invariance of regional paleoclimate and the evolution of the thickness and density of the crust and lithospheric mantle.

Several end-member geodynamic models have been proposed for the central Andes (Fig. 1B), each with specific implications for the past surface uplift history of the Andes. All models are similar in that they require thickening of low-density crustal material and removal of higher density lithospheric mantle such that isostatic compensation increases the plateau elevation. They differ in their proposed mechanism for, and timing of, lithospheric mantle removal. For example, the ablative subduction model (Tao and O'Connell, 1992; Pope and Willett, 1998) (Fig. 1B) suggests lithospheric mantle removal from subduction–erosion by the Nazca Plate. In this model, the progressive removal of lithospheric mantle results in a slow and steady plateau rise. Thermal weakening models (Fig. 1B) also suggest a slow and steady rise of the plateau (e.g., Isacks, 1988; James and Sacks, 1999). In this model, initial flat-slab subduction is hypothesized for the central Andes facilitating dehydration of the slab during subduction, which hydrates the overlying lithospheric mantle and weakens it. After slab dip increases to its modern geometry mantle wedge corner flow mechanically removes the hydrated lithospheric mantle. Alternatively, delamination of eclogized lower crust (Fig. 1B) results in the rapid removal of lithospheric mantle from below the central Andes (Garzione et al., 2006; Ghosh et al., 2006; Sempere et al., 2006; Garzione et al., 2007; Molnar and Garzione, 2007; Quade et al., 2007) and elsewhere (Bird, 1979). This delamination model predicts a rapid surface uplift of the Bolivian Altiplano plateau over several million years following mantle root removal (Garzione et al., 2006; Ghosh et al., 2006; Molnar and Garzione, 2007; Garzione et al., 2008).

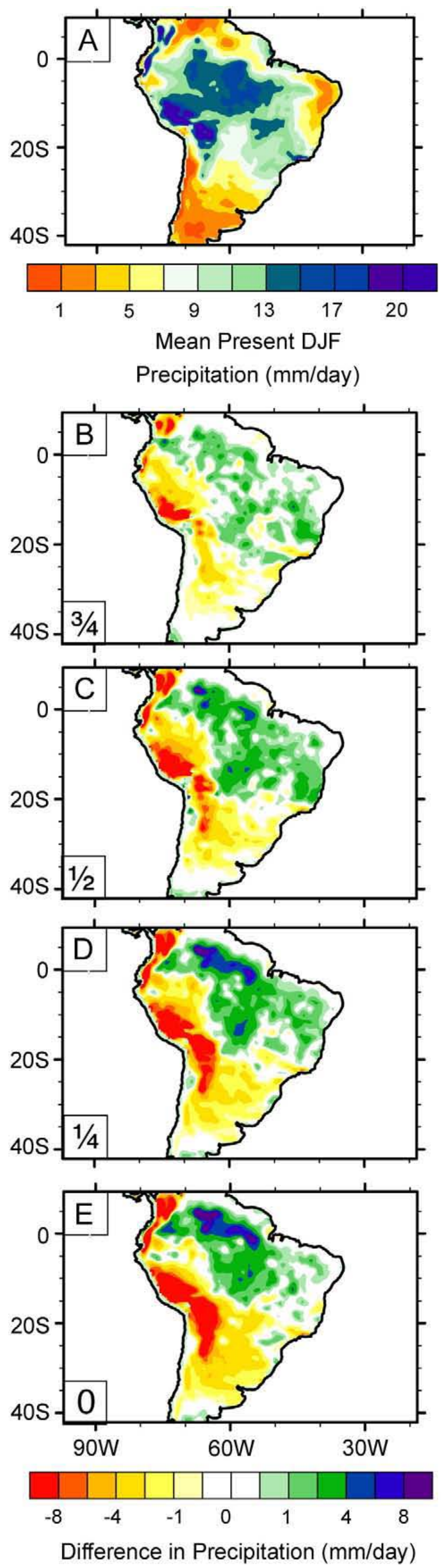

Fig. 2. Simulated austral summer continental precipitation rate (mm/day). (A) Mean summer precipitation rate from a modern simulation of South America. (B–E) Summer precipitation rate differences between the ¾ (B), ½ (C), ¼ (D), and no-Andes (E) simulations, and the modern simulation. All data represent 5-year December–January–February (DJF) averages of RegCM simulations. Note that over the central Andean Plateau precipitation rates increase substantially as the plateau is elevated. As explained in Section 5.3, greater precipitation rates would enhance the "amount effect", causing a depletion of precipitation $\delta^{18}O$. Unless this effect is considered in the interpretation of paleoaltimetry proxy data, it would appear that Andean Plateau uplift was more rapid.

2.2. Paleoaltimetry data and Andean Plateau surface uplift

Support for the lithospheric delamination model under the Bolivian Altiplano comes from several different, climate sensitive, paleoaltimetry techniques. These techniques include paleobotany foliar physiognomy (Gregory-Wodzicki et al., 1998; Gregory-Wodzicki, 2000), a 3–4‰ decrease in soil carbonate $\delta^{18}O$ (Garzione et al., 2006, with corrected values in Garzione et al., 2007; Quade et al., 2007; see also Hartley et al., 2006), and clumped (Δ_{47}) isotope paleothermometry (Eiler et al., 2006; Ghosh et al., 2006; Sempere et al., 2006). When combined, these techniques have been interpreted to suggest a 2.5 ± 1.0 km surface uplift of the Bolivian Altiplano between ~10 and 6 Ma (Garzione et al., 2006; Ghosh et al., 2006; Quade et al., 2007; Garzione et al., 2008). For each of these techniques, the interpreted paleoelevation was determined by utilizing either modern $\delta^{18}O$–elevation relationships in precipitation, or modern mean-annual temperatures and atmospheric adiabatic lapse rates (paleobotany and Δ_{47} techniques).

Soil carbonate, paleobotany and the Δ_{47} paleoaltimetry techniques use modern atmospheric adiabatic lapse rates and mean-annual temperatures in the foreland to reconstruct the paleoelevation of sample deposition. These techniques assume that surface temperature only changes as a function of lapse rate that may or may not vary in response to plateau uplift. These caveats to interpreting paleoelevations are often generally acknowledged in paleoaltimetry studies. However, to our knowledge, a detailed, quantitative assessment of these effects and their implications for the Andean Plateau paleoelevation history has not yet been undertaken.

2.3. South American climatology and precipitation $\delta^{18}O$

Precipitation over the present-day Andean Plateau is mainly associated with deep, moist convection, i.e., the buoyant ascent of near-surface air to the base of the troposphere, as air is dynamically lifted up the eastern flanks of the Andes (Lenters and Cook, 1995). Moist convection is fueled by latent heat release during condensation. In the Andes, the source of latent heat and precipitation is predominantly the Amazon region (Fuenzalida and Rutllant, 1987; Vuille et al., 1998). This observation is supported by the low precipitation $\delta^{18}O$ on the Andean

Plateau, which is related to progressive fractionation as air masses move across the Amazon, ascend the Andes, and rainout in convective storms (Aravena et al., 1999). The flow of moist, cool air from the Pacific Ocean is limited by topography and a persistent inversion arising from large-scale subtropical subsidence over the southeast Pacific (Garreaud et al., 2003).

Andean precipitation is highly seasonal with 70% occurring during austral summer (e.g. Fig. 2A, Appendix Fig. A1). The seasonality of precipitation is attributed to changes in the zonal winds, rather than direct forcing through seasonal insolation (Garreaud, 1999; Lenters and Cook, 1999; Garreaud et al., 2003). During winter and early spring, the subtropical jet stream is located in its most northerly position, causing westerly flow aloft over the Andean Plateau. This eastward flow hinders the flow of moist air across the eastern slope of the Andes and thus impedes moist convection. During summer, when the subtropical jet stream has shifted southward, flow aloft is predominantly easterly and promotes the transport of moist air across the continent. This moist air, in conjunction with dynamic forcing, gives rise to convection and rainfall over the plateau (Garreaud, 1999; Garreaud et al., 2003). The influence of large-scale circulation on Andean precipitation is further supported by interannual variability over the plateau. For example, variations in Andean summer precipitation have been linked to large-scale forcing due to changes in tropical Pacific climate (the El Nino Southern Oscillation phenomenon) (Lenters and Cook, 1999; Garreaud et al., 2003; Vuille et al., 2003b).

The stable isotopic concentration of meteoric water (rain or snow) is determined by the isotopic concentration of the source vapor, isotopic dilution/enrichment through mixing with other air masses, and isotopic fractionation through mass-, and symmetry dependent processes that act upon the vapor (see Gat, 1996 for a review). At high and mid latitudes, a strong positive correlation exists between $\delta^{18}O$ and temperature. This temperature–$\delta^{18}O$ relationship is the basis for carbonate $\delta^{18}O$ paleo-altimetry and thermometry, and reflects the tendency for air masses to become saturated and form precipitation as air temperature falls following the Clausius–Clapeyron law. The loss of water from an air mass through precipitation causes the remaining vapor (and the resulting condensate) to become depleted with respect to the heavier isotope. The progressive isotopic depletion of an air mass through rainout is often considered to follow a Rayleigh law. In the tropics the isotopic composition of precipitation depends on the integrated precipitation amount along the entire air-mass pathway (from the moisture source to the site of precipitation). The fractionation of vapor is more complicated than a simple distillation process by progressive rainout, and includes non-equilibrium precipitation and re-evaporation of vapor within an air column and from surface waters, and amount effects (Gat, 1996). Amount effects are important in tropical regions such as the central Andes where precipitation $\delta^{18}O$ is controlled by the rainout amount (hence the name) and convective instabilities rather than condensation temperature. In these regions, the relationship between temperature and $\delta^{18}O$ can be negative and the fractionation history deviates from that predicted by Rayleigh distillation over seasonal timescales (Rosanski and Araguas-Araguas, 1995; Aravena et al., 1999; Gonfiantini et al., 2001; Blisniuk and Stern, 2005; Vimeux et al., 2005).

It is evident from the above that the Andean Plateau is a first order control on the South American climatology and precipitation $\delta^{18}O$. Here, we consider how much South American climate differed in the past when the plateau grew over time.

3. Methods

3.1. Global and regional general circulation modeling

The objective of this study is to evaluate climate change related solely to Andean uplift. Our approach to investigating past changes in climate over the Andean Plateau and South America involves the use of regional general circulation models (RCMs) of Earth's climate. RCMs predict the three-dimensional climate based on the primitive equations for atmospheric circulation and on representations (parameterizations) of physical processes (e.g. precipitation, clouds, convection) that occur on sub-grid scales. Due to generally coarse spatial resolution, global circulation models (GCM) can have severe deficiencies in regions with steep topography. For this reason, we use RegCM3, a grid-point limited-domain RCM. RegCM3 has been tested and shown to successfully simulate regional climate in a variety of domains throughout the world (c.f., Giorgi et al., 2006). RegCM3 has a hydrostatic dynamical core (similar to the NCAR/PSU MM5; (Grell et al., 1994)) and a full radiation package (CCM3, Kiehl et al., 1996). In our implementation of RegCM3, we employ the Emmanuel convection scheme (Emmanuel, 1991). The physical domain covers much of South America from ~46 °S to 13 °N and 100 °W to 18 °W (e.g., Fig. 2). We use a 60-km horizontal grid spacing (for a total domain of 160×120), 18 vertical levels, and a timestep of 100 s.

In this study, we have conducted a series of experiments with systematic changes in the elevation of the Andes. The control case represents a modern simulation with present-day Andean topography. Three sensitivity experiments were completed in which the Andean topographies were reduced to ¾, ½, and ¼ of their present-day elevations. A fourth sensitivity experiment (referred to as no-Andes) was completed in which the Andes are flat with an elevation of approximately 200 m. All other boundary conditions, including trace gas concentrations (i.e. 355 ppmv CO_2, 1714 ppbv CH_4, 311 ppbv NO_2), land-surface characteristics (modern vegetation), sea-surface temperatures (from the Atmospheric Modeling Intercomparison Project), and solar and orbital parameters, were specified to represent modern conditions and remain constant between the simulations. Thus, the only variable in the experiments is Andean topography, allowing isolation of the effect of the rise of the Andes on South American climate.

Regional climate models require climatological information along the boundaries of their domain. This information can be supplied by (1) observational or reanalysis data or (2) global GCMs, in which case the regional model is "nested" within the global GCM. We have used both techniques. In the first case, we used gridded atmospheric variables (zonal and meridional wind, specific humidity, surface and atmospheric temperature, surface pressure) for years 1991–2000 from the NCEP/NCAR global reanalysis product (Kistler et al., 2001) to force RegCM3. The advantage to this method is that boundary conditions are based on assimilated observed data, and therefore do not include biases that might result from using a global GCM. In the second case, we ran a series of simulations with differing Andean elevations using the GENESIS 2.3 GCM, and then used predicted atmospheric variables as boundary conditions to RegCM3. The advantage to this method is that the atmospheric boundary conditions include the effects of changing Andean elevations. In this report, we focus on and present experimental results from simulations using the first method. Our decision to focus on these results is simple; the modern (control) simulation using this method closely matches modern observations, particularly the magnitude and distribution of precipitation. Importantly, the sensitivity to changing Andean topography was very similar in both techniques, indicating that the atmospheric boundary conditions are not the principal control on the regional climate over the Andes.

Each sensitivity experiment was run for 10 years. Because sea-surface temperatures were specified, model equilibrium is attained in only a couple of years. All model results reported below have been averaged over the final 5 years of the simulations.

3.2. Model limitations

In this study, our goal is to quantify climate change related to Andean uplift and assess its significance to climate-sensitive paleoaltimetry proxies. To this end, we have designed RCM sensitivity experiments that provide for the most straightforward assessment of the relationship between Andean height and climate. These experiments are clearly idealized; the uplift of the Andes was not uniform, but likely varied across the range (e.g., Allmendinger et al., 1997; Barnes et al., 2008; McQuarrie et al., 2008). Moreover, during the Cenozoic uplift of the Andes, additional tectonic and climatic factors were evolving that may have influenced regional and global climate including minor continental drift of South America, the glaciation of Antarctica, changes in large-scale ocean circulation, and the decline of atmospheric pCO_2. These factors will need to be addressed in future studies. However, the large changes in plateau elevation investigated here are very likely to dominate regional continental climate change over South America.

A significant contribution of this work is the estimation of the influence of climate change over South America on the oxygen isotopic concentrations of precipitation (and carbonates). Because RegCM3 does not currently include oxygen isotopic tracer capabilities, we use the best available modern relationships between climate and $\delta^{18}O$ to evaluate this influence. It is possible that Andean climate–$\delta^{18}O$ relationships may have differed from those cited here (see below) under different climatic conditions. Though these variations may alter the details of this work, they are unlikely to change our conclusion that paleoclimate considerations can substantially influence paleoaltimetry interpretations.

4. Results

In the following, we present the predicted climate changes due to lowering of the Andes. Our analysis here is limited to differences in South American precipitation, surface temperature, and near-surface (810 mb) winds for past elevations of the Andes. A more detailed analysis of the dynamical and physical atmospheric changes associated with variations in plateau height is beyond the scope of this paper, and will be presented in a future publication currently under preparation. Results are shown for the Southern Hemisphere summer (December, January, February; DJF), the rainy season when 70% of the moisture falls on the Andes and soil carbonates used as paleoaltimetry proxies acquire their isotopic signature (Quade et al., 2007). Differences highlighted for the rainy season precipitation and surface temperature (Figs. 2 and 3) were calculated by subtracting the modern prediction (Figs. 2A, 3A) from the predicted value for the experiments with reduced plateau heights.

4.1. Changes in paleo precipitation magnitude

Modern predicted and observed (Appendix Fig. A1c) (Vose et al., 1992) central South American precipitation is highly seasonal. The rainy season occurs during the Southern Hemisphere summer (DJF) and is associated with the southern migration of the Intertropical Convergence Zone (ITCZ) over the region. Daily predicted (Fig. 2A) and observed (Appendix Fig. A1c) precipitation rates during the rainy season are 9–20 mm/day between 0° and ~20° south latitude. During winter the ITCZ migrates northward, as does the locus of precipitation.

The lowering of plateau elevations results in large changes to low-latitude precipitation (Fig. 2B–E). Precipitation rates over the central plateau systematically decrease as elevations are reduced from their modern height (Table 1). For example, predicted precipitation in the central Andean Plateau ranges between ~0.5 and 8 mm/day less when the Andes are at ¾ of their modern elevation (Fig. 2B). Further decreases in elevations to ½, ¼, and no-Andes (Fig. 2C–E) lead to (a) larger reductions in precipitation, and (a) an expansion of the area of reduced precipitation on the plateau. Not all areas of the Andes are predicted to be drier with lower plateau paleoelevations. In fact, the northern Andes are substantially wetter. This pattern of increased precipitation over the northern Andes and decreased precipitation over the central Andes indicates that the modern high-elevation Andean plateau serves as an important platform for atmospheric convection that effectively promotes vapor convergence and seasonal rainfall. Finally, isolated areas on the steep western flank of the Andes (not the plateau) are also predicted to be 0.5–3 mm/day wetter than modern (green areas, Fig. 2B–E).

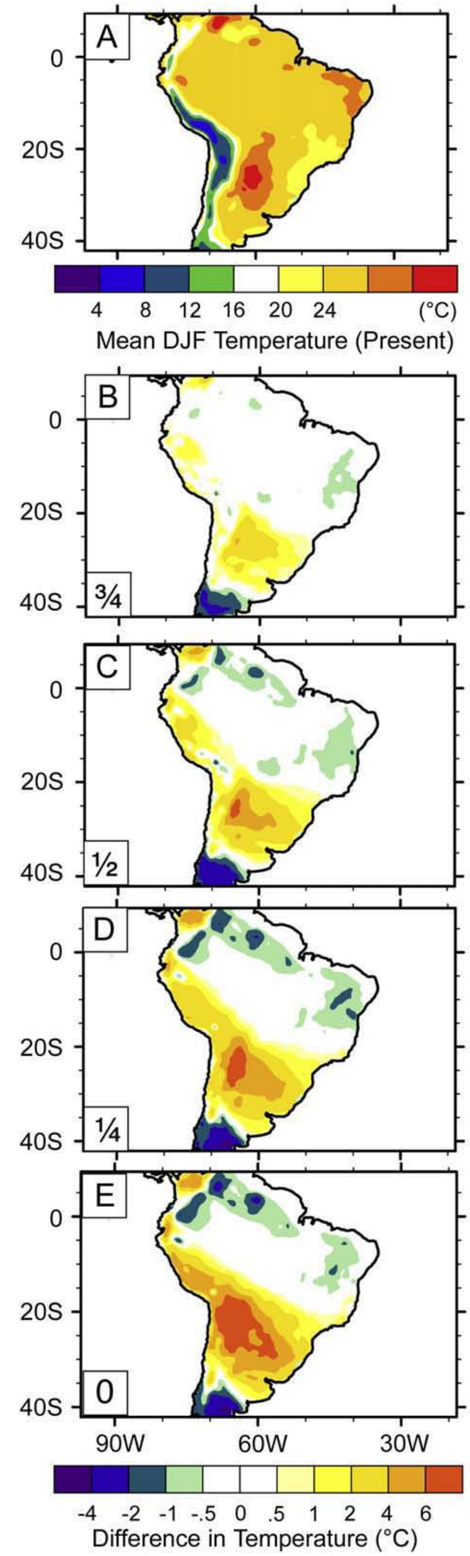

Fig. 3. Simulated austral summer continental surface temperature (°C). (A) Mean summer temperature for a modern simulation of South America. (B–E) Non-adiabatic summer temperature difference between the ¾ (B), ½ (C), ¼ (D), and no-Andes (E) simulations, and the modern simulation. All data represent 5-year December–January–February (DJF) averages of RegCM simulations. The non-adiabatic temperature differences in B–E were calculated assuming a lapse rate of 5 °C km^{-1}. In the model, the rise of the Andean Plateau leads to substantial non-adiabatic cooling. As explained in Section 5.1, this cooling would deplete precipitation $\delta^{18}O$. Unless this cooling effect is considered in the interpretation of paleoaltimetry proxy data, it would appear that Andean Plateau uplift was more rapid.

Table 1
Climate influence of Andean Plateau surface uplift; magnitude of climate response; and estimated influence on precipitation $\delta^{18}O$.

	Climate effect of uplift	Magnitude of response[a]	Change in $\delta^{18}O$/elevation[b]
Temperature	Non-adiabatic decrease in temperature		Decrease in $\delta^{18}O$/increase in paleo-elevation
¾ Andes		+0.0 °C	+0.0‰/0 km
½ Andes		+0.6 °C	+0.7‰/0.3 km
¼ Andes		+3.4 °C	+3.8‰/1.6 km
No Andes		+6.6 °C	+7.3‰/3.0 km
Precipitation	Increase in precipitation rate and occurrence of high intensity events		Decrease in $\delta^{18}O$/increase in paleo-elevation
¾ Andes		−1.6 mm day^{-1}	+1.8‰/0.8 km
½ Andes		−4.0 mm day^{-1}	+4.2‰/1.8 km
¼ Andes		−7.8 mm day^{-1}	+8.8‰/3.6 km
No Andes		−10.0 mm day^{-1}	+11.3‰/4.7 km
Wind direction and vapor source	Shift in low-level winds to predominantly easterly and vapor source from Atlantic		Decrease in $\delta^{18}O$/increase in paleo-elevation
¾ Andes		Westerly from Pacific	>0.0‰/>0 km
½ Andes		Westerly from Pacific	>0.0‰/>0 km
¼ Andes		Westerly from Pacific	>0.0‰/>0 km
No Andes		Westerly from Pacific	>0.0‰/>0 km

[a] Data are averaged over central Andes Plateau (~17–22 °S, 69–66 °W) for December–January–February. Simulated modern summer surface temperature is 9.7 °C. Simulated modern summer precipitation rate is 11.0 mm day^{-1}. Simulated modern low-level winds are predominantly from the east.
[b] Explanation of methods used for estimating the $\delta^{18}O$ and elevation change is in Section 5.

4.2. Changes in paleo surface-air temperature

Modern observed and predicted rainy season surface-air temperatures range from ~28 °C in the Andean foreland to ~10 °C on the plateau (Fig. 3A). The surface-air temperature distribution is largely a function of elevation due to the adiabatic lapse rate, which ranges between −4.5 and −5.5 °C km^{-1} in the tropics (e.g. Gonfiantini et al., 2001). Lapse rates can vary by region due to differences in tropospheric conditions mainly related to local atmospheric humidity.

A decrease in plateau elevation can influence surface temperature due to both (a) adiabatic warming directly linked to elevation lowering, and (b) non-adiabatic warming or cooling associated with climate change (due to changes in atmospheric circulation, radiative and/or surface heating). To estimate the temperature change attributed to non-adiabatic warming (Fig. 3B–E), we first estimate adiabatic warming by applying an adiabatic lapse rate of 5.0 °C km^{-1} (Gonfiantini et al., 2001), and then subtract the adiabatic warming from the total temperature change. For example, in the case with a flat Andes, the absolute temperature change central Andean Plateau is ~24 °C; we estimate a regional surface temperature increase of ~6.5 °C after subtracting the temperature increase (~17.5 °C for 3.5 km elevation change at 5 °C km^{-1}) due to adiabatic warming.

In our simulations, lowering the Andean Plateau results in regional non-adiabatic warming (Table 1). For example, when the Andean Plateau was ¾ to ½ of its present elevation, regional climate change caused central Andean surface temperatures to increase by 0.0 to 0.6 °C (Fig. 3B, C) relative to modern temperatures, respectively. Further lowering of the plateau in the ¼ (Fig. 3D) and no-Andes cases (Fig. 3E) leads to additional warming of 3.3–6.5 °C. The regional extent of warming expands as the plateau elevations decrease (compare Fig. 3B and E). The warming of central Andean surface temperatures is mainly associated with the northward shift in the convective region. In the absence of atmospheric convection, latent and sensible heat transfers to the atmosphere are reduced, leading to higher surface temperatures.

4.3. Change in prevailing wind direction and vapor source

Predicted and observed rainy season lower tropospheric winds (810 mb) over the central Andean Plateau flow westward across the Amazon basin. The winds are deflected southward along the eastern flank of the Andes before moving southeast to form the south Atlantic convergence zone (Fig. 4A). As a result of this wind pattern, the

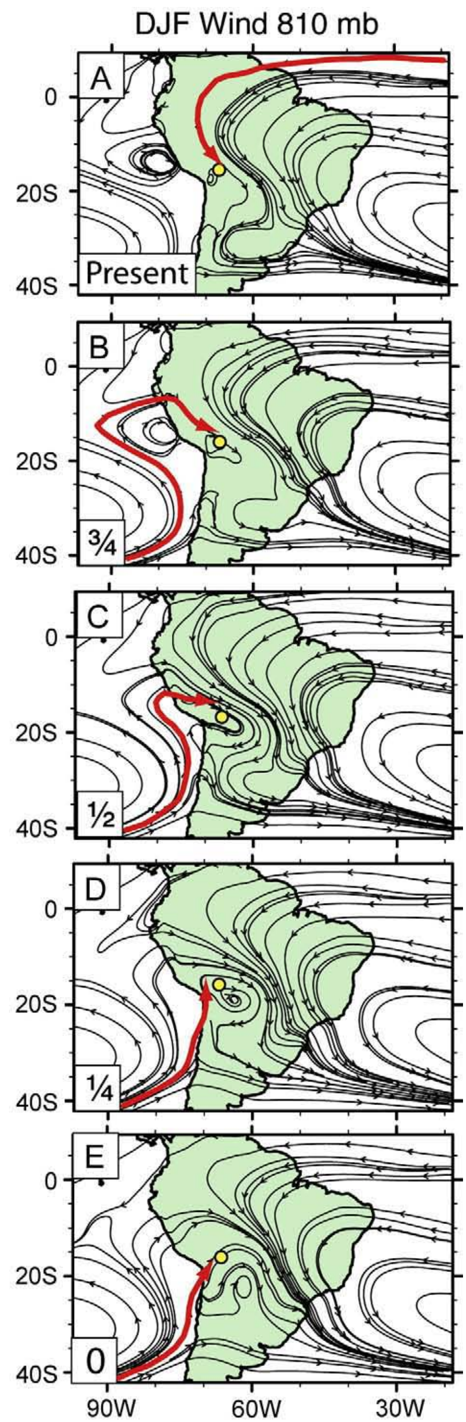

Fig. 4. Simulated austral summer 810-mb wind streamlines for modern (A), ¾ (B), ½ (C), ¼ (D), and no-Andes (E) cases. All data represent 5-year December–January–February (DJF) averages of RegCM simulations. Flow onto the central Andean Plateau (yellow filled circle) is denoted by a thick red line. The summer wind direction changes from predominantly westerly (B–E) to northeasterly (A) as plateau elevations are increased. The shift in wind direction would influence the source and fractionation history of vapor precipitated over the plateau.

equatorial Atlantic Ocean is the dominant vapor source for much of the modern central Andean precipitation, including rainfall on the plateau.

Prevailing wind directions are very sensitive to the height of the plateau. A marked change in wind directions occurs at lower-than-modern plateau elevations (compare Fig. 4A with Fig. 4B–E). In the ¾, ½, ¼ (Fig. 4B, C) and no-Andes experiments prevailing winds are from the west to northwest (Fig. 4B, C) or southwest (Fig. 4D, E) rather than the east. These changes in wind direction result in (a) a Pacific Ocean vapor source and (b) a shorter transport distance between the vapor source and plateau. Both changes can influence the isotopic composition and magnitude of precipitation over the plateau.

5. Discussion

Our climate model results predict that the uplift of the Andean plateau would have led to the following changes in central Andean climatology: (1) an increase in precipitation, (2) a decrease in non-adiabatic surface temperature (in addition to decreases in adiabatic surface temperature), and (3) a reversal of low-level winds from predominantly westerly to easterly. Soil carbonate and paleobotany data used to calculate the elevation history of the Andean Plateau do not consider these changes. Here we evaluate the influence these climatic changes would have on plateau elevation reconstructions that use a 3–4‰ depletion of δ^{18}O, and a ~17 °C decrease in temperatures as the basis for a rapid rise of the plateau (Quade et al., 2007). For simplicity we discuss the influence of each variable (temperature, precipitation rate, vapor source) separately, while recognizing that in reality these are not independent of each other.

5.1. Influence of paleo-surface temperature changes on paleoaltimetry data

Surface temperature changes associated with changes in plateau elevation influence precipitation δ^{18}O. Isotopic fractionation of vapor increases as temperature decreases (the "temperature" effect) leading to more negative δ^{18}O values in rainwater (e.g., Dansgard, 1964). In the tropics, sampling stations in the Amazon Basin indicate a δ^{18}O–temperature dependence of −1.13 to −1.64‰/°C (Rosanski and Araguas-Araguas, 1995). Our model simulations indicate that uplift of the plateau would lead to a 0–6.5 °C decrease (Table 1) in regional surface temperature (in addition to any adiabatic cooling) resulting in ^{18}O depletion and more negative precipitation δ^{18}O (Fig. 3). Taking a −1.13‰/°C dependence of δ^{18}O on surface temperature times a −6.5 °C predicted change in temperature leads to a reduction of modern precipitation (and thus carbonate) δ^{18}O of 7.3‰ (Table 1). We emphasize that this 7.3‰ shift due to regional climate change is in addition to any isotopic changes resulting from adiabatic cooling of an air mass lifting over the Andes. Unless this regional temperature effect is considered, stable isotope paleoaltimetry estimates likely overestimate the rapid rise of the plateau. For example, the modern observed precipitation δ^{18}O–elevation relationship on the eastern flank of the central Andean Plateau is −2.39‰ km^{-1} (Gonfiantini et al., 2001). Using this δ^{18}O–elevation relationship and our estimated 7.3‰ temperature effect for the no-Andes scenario suggests a ~3 km overestimation of the rapid rise of the plateau from paleoaltimetry estimates.

The decrease in surface temperatures associated with regional climate change during uplift also influences elevation reconstructions using paleobotany and clumped isotope techniques (Gregory-Wodzicki et al., 1998; Gregory-Wodzicki, 2000; Ghosh et al., 2006). These techniques rely on modern temperature-elevation relationships to determine paleoelevations at the time of soil formation. However, modern predicted temperatures are 0–6.5 °C cooler (after corrections for lapse rate and elevation changes) than when the Andes were lower (Fig. 3B–E). Using a lapse rate of 5.0 °C km^{-1} (Gonfiantini et al., 2001), our predicted changes in surface temperatures imply apparent paleoelevations from paleobotany and clumped isotope techniques overestimate a rapid rise of the Andes by 0.0 to 1.3 km.

5.2. Influence of changing vapor source on stable isotope paleoaltimetry data

The δ^{18}O composition of precipitation is also sensitive to the vapor source and transport distance from the source. Results shown in Fig. 4B–D demonstrate a large change in rainy season wind directions associated with plateau uplift. Most notably: (a) the vapor source shifts from the southwestern Pacific to the equatorial Atlantic as the plateau rises to above ½ to ¾ of its modern elevation, and (b) the change in vapor source results in an increase in transport distance. The δ^{18}O of water vapor from the southwest Pacific is ~5–10‰ enriched in ^{18}O relative to the equatorial Atlantic (Vuille et al., 2003a). Due to atmospheric mixing of vapor between air masses, the change in the δ^{18}O of Andean precipitation would likely be somewhat less. In addition, the change in prevailing wind direction with plateau uplift would increase the transport distance from the vapor source to the plateau, increasing isotopic fractionation of vapor and decreasing precipitation δ^{18}O. The magnitude of this effect is difficult to constrain without isotopic tracer capabilities in our regional climate model. Nonetheless, if these effects are not considered, interpretations based on stable isotope paleoaltimetry will likely overestimate the magnitude of surface uplift of the Andes.

5.3. Influence of paleoprecipitation changes on stable isotope paleoaltimetry

The amount of precipitation that falls in individual storms has a large effect on the δ^{18}O composition of rainwater. High rainfall rates enhance isotopic depletion of rainfall (the "amount effect"). Simulated modern rainy season (DJF) precipitation magnitudes on the central Andean Plateau are ~4 mm/day, or 360 mm/season, greater in the control case than precipitation rates in the ½ Andes experiment (Fig. 2C, Table 1). Importantly, the rainfall intensity also differs between experiments with more numerous large rainfall events in the control case than in the experiments with reduced Andean elevations (Fig. 5). The increase in modern precipitation amounts over the plateau likely led to an increase in the ^{18}O depletion of rain and more negative precipitation and soil carbonate δ^{18}O. Unless this amount effect is considered, apparent paleoelevations inferred from the δ^{18}O of paleosol carbonates will overestimate the magnitude of surface uplift.

Modern estimates of the precipitation–δ^{18}O estimates in the Andes indicate an amount effect of −0.038‰ mm^{-1} precipitation, month^{-1} (calculated from Vimeux et al., 2005). We calculate the amount effect for the central Andes using the previous value multiplied by the change in the

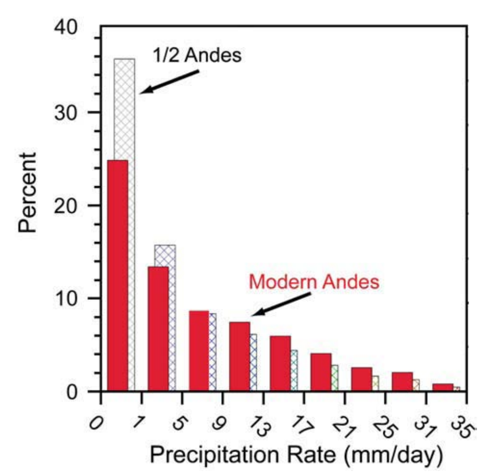

Fig. 5. Histogram of austral summer (December–January–February) precipitation intensity for modern (color bars) and ½ Andes (cross-hatch pattern) simulations. Five-years of daily precipitation were binned into precipitation intensity (rate) intervals. The percent (*y*-axis) represents the relative number of precipitation events in each intensity (rate) interval. Note that the ½ Andes simulation has a larger number of low-intensity rainfall events compared to the modern simulation. In the modern simulation, more rain falls in high-intensity events. This difference in the intensity of precipitation during storms influences the fractionation of water isotopes, as discussed in Section 5.3, causing modern precipitation have lower δ^{18}O on average.

average rainy season monthly precipitation on Andean Plateau (Table 1). For the ¾, ½, ¼, and no-Andes simulations (Fig. 2B–E) we predict a 48, 120, 234, and 300 mm month^{-1} decrease in precipitation over the plateau, respectively. These changes in precipitation correspond to δ^{18}O depletions of 1.8, 4.2, 8.8, and 11.3‰. Using an isotopic lapse rate of −2.39‰ km^{-1} (Gonfiantini et al., 2001), this amount effect might lead to overestimates of plateau uplift from soil carbonate paleoaltimetry on the order of ~0.8 to 4.7 km.

5.4. Paleoclimate implications for the plateau elevation history

As described above and summarized in Table 1, the Cenozoic uplift of the Andean Plateau influenced regional climate in dramatic ways. When taken together, these paleoclimate changes result in a substantial depletion of precipitation ^{18}O. The magnitude and timing of the ^{18}O depletion depends on the uplift history of the plateau, but over the Cenozoic was likely >1.8–18.6‰ (the minimum and maximum of possible effects summed from Table 1). The cumulative effect of these changes is comparable to, or greater than, the measured 3–4‰ depletion of ancient carbonate δ^{18}O used to interpret a rapid plateau rise (Garzione et al., 2006, 2007; Molnar and Garzione, 2007; Garzione et al., 2008) and draws into question the interpretations of lithospheric mantle delamination (Fig. 1B).

Much of the apparent rapid rise of the Andes could be associated with climate change. To date, soil carbonate δ^{18}O has been interpreted solely as a function of paleo-elevation without consideration of climate effects. Using a modern precipitation δ^{18}O–elevation relationship of −2.39‰ km^{-1} for the Bolivian Andes (Gonfiantini et al., 2001; see also Vimeux et al., 2005), a depletion of >1.8–18.6‰ due to climate change associated with uplift between no- and 3/4 Andes is equivalent to an apparent increase in paleo-elevation of 0.8 to 7.7 km. The upper value of this estimate was calculated for the no-Andes scenario and is greater than the modern elevation of the central Andean Plateau (~3.7 km). If isotope data were available from the early phases (circa 40 Ma) of Andean mountain building and were interpreted using modern δ^{18}O–elevation relationship then we predict they would suggest apparent paleoelevation increases that are greater than the current elevation of the plateau. Unfortunately, the oldest paleoaltimetry data currently published from the plateau are ~27 Ma (Garzione et al., 2008).

Furthermore, our results also suggest that temperature-sensitive paleoaltimetry proxies such as paleobotany and clumped isotope techniques will produce apparent elevations that are 0 to 1.3 km lower than the true elevation at the time of sample formation. The key point demonstrated here is that paleoclimate changes associated with plateau uplift are significant, and the effect of climate change on paleoaltimetry proxies is as large, or larger than, the signal of elevation change interpreted for the Andes. Thus, the interpretation of paleoelevations by comparison to modern climate is fraught with uncertainty, and should not be decoupled from consideration of regional climate change associated with orogen growth.

We summarize the impact of climate change on reconstructions of Andean paleoelevation. The reconstruction shown in Fig. 6 is based on existing paleobotany, δ^{18}O, and clumped isotope techniques (Ghosh et al., 2006; see also Garzione et al., 2008). In this figure, we also show the 1σ, and previously unplotted, 2σ uncertainties affiliated with each study (including reported standard deviations and standard error). Previous studies proposed a rapid 2.5 ± 1.0 km increase in plateau elevation between ~10 and 6 Ma. Prior to ~10 Ma, the plateau elevation is inferred to be low (<~2 km). However, this interpretation does not account for (a) the 2σ uncertainties associated with the data, and (b) climate change associated with plateau rise. Results from this study demonstrate that paleoelevations determined from the δ^{18}O of carbonate minerals recovered in soils >~10 Ma need to be corrected by >0.8 to 7.7 km to show the true elevation at the time of sample preservation. After applying this correction, previous interpretations of a rapid rise of the plateau are no longer required at the 1σ, or 2σ levels of uncertainty. The implication is that the plateau could have been as high as 2–3 km by ~20 Ma when the modern E–W width of the plateau was established (Horton et al., 2002; McQuarrie, 2002; Barnes et al., 2006, 2008), and risen at a slow and steady rate since then.

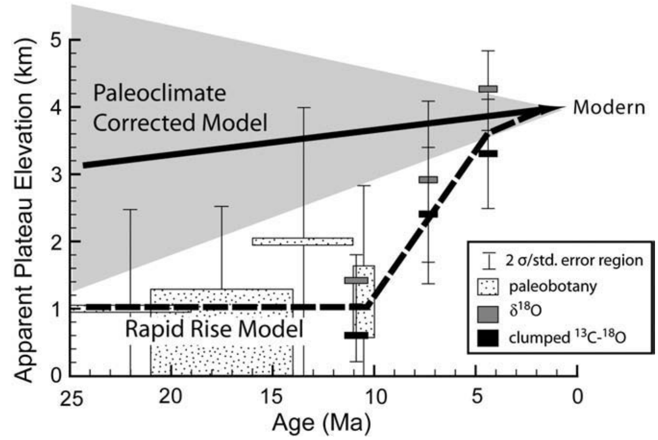

Fig. 6. Paleoclimate influence on apparent paleoelevation reconstructions from the central Andean Plateau. Previously published paleoaltimetry proxies used to suggest a rapid rise of plateau (thick dashed line) are shown with 1- and 2-sigma uncertainties (Gregory-Wodzicki et al., 1998; Gregory-Wodzicki, 2000; Garzione et al., 2006; Ghosh et al., 2006). On the basis of our climate model results, we estimate a >0.3–4.5 km paleoclimate correction should be applied to the paleoaltimetry data. The gray shading represents our estimated range of possible plateau elevations after application of a paleoclimate correction to the rapid-rise model. The solid line represents our estimated average elevation history after correction. See Section 5.4 for details.

5.5. Predicted paleoprecipitation and fluvial incision of the plateau flanks

A secondary, and intriguing, model result is that simulated precipitation rates along the western and eastern flanks of the plateau are higher (by 0.5–3 mm/day) when plateau elevations are reduced (e.g., Fig. 2B, C). Increased precipitation rates on the plateau flanks could cause an increase in fluvial discharge, and increased fluvial incision into the plateau margins. Interestingly, the locations of our predicted increased precipitation on the plateau flanks correspond with locations where numerous studies have documented >1 km fluvial incision sometime between 2 and 12 Ma (Barke and Lamb, 2006; Schildgen et al., 2007; Thouret et al., 2007; Hoke et al., 2007). Some of these studies concluded that rapid surface uplift was the cause of increased fluvial incision through the steepening of river channel slopes. Our results provide an alternative explanation, and suggest that increased incision rates could be explained by higher paleo-discharge that existed when plateau elevations were lower than the modern. Thus, we caution that it may be premature to argue solely for rapid surface uplift as the driver for plateau incision, and suggest that more detailed studies of the influence of climate change and fluvial incision (e.g., Whipple and Tucker, 1999; Schaller and Ehlers, 2006) are needed on the flanks of the plateau.

5.6. Model limitations and implications for paleoelevation estimates

In closing, we emphasize that our analysis represents a first-order estimate of the influence of climate factors on Andean precipitation δ^{18}O. We calculate a range of possible δ^{18}O depletion to provide a sense of the possible magnitude and importance of these effects on paleoaltimetry estimates. However, we acknowledge that our maximum depletion estimates may be unreasonable. For example, using modern Andean Plateau δ^{18}O values of −15‰ (e.g. Gonfiantini et al., 2001) and ignoring isotopic distillation due to airmass lifting, our maximum depletion would suggest pre-uplift precipitation δ^{18}O in the Andean region >3.6‰. Modern low-latitude (20°), low-elevation (<200 m) meteoric δ^{18}O is typically −4 to −5‰ (Bowen and Wilkinson, 2002). We also note that factors related to the Cenozoic climate evolution would have affected Andean precipitation δ^{18}O. We expect that most of these factors, including the decline of global atmospheric pCO_2 and the development of high-latitude ice sheets, would have had a relatively small (1–2‰) direct effect on Andean precipitation δ^{18}O. For example, using an atmospheric GCM with water isotope capabilities, Poulsen et al. (2007) estimate a ~2‰ increase in low-latitude precipitation δ^{18}O with a 4× increase in atmospheric CO_2 from 8 to 2× present atmospheric levels. Other influences, including the evolution of Pacific and Atlantic sea surface conditions, may have altered regional circulation patterns, and could possibly have had a more substantial, but secondary, impact on Andean precipitation δ^{18}O. The range of possible depletions, and their impact on paleoelevation reconstructions, highlight the need for future paleoclimate studies to use isotope tracking regional climate models (e.g. approach of Sturm et al., 2005 for modern conditions). These improved simulations are needed to calculate the climate effects and to refine the estimates made here.

6. Conclusions

This study predicts significant changes in South American temperature, vapor source, and precipitation associated with the rise of the Andean Plateau that have important implications for interpreting paleoaltimetry proxies. Most importantly, these paleoclimate changes would have substantially altered Andean precipitation δ^{18}O and surface air temperatures, complicating paleoaltimetry estimates based on carbonate δ^{18}O, clumped isotopes, and paleobotany techniques. Because these paleoclimate changes have not been considered, previous elevation reconstructions of the plateau have likely overestimated the rapid rise of the plateau. When paleoclimate corrections to the apparent elevation history are considered (Fig. 6), constant slow growth of the Andean plateau over the last 25 Ma is a viable model for plateau evolution. A slow and steady rise of the Andean Plateau draws into question recently proposed models of lithospheric mantle delamination as the driving mechanism for rapid late Miocene uplift. Finally, future studies in the Andean and Tibetan Plateaus, or other orogens, that are interested in accurately quantifying the elevation of history need to consider the effects of paleoclimate and changing orography on paleoaltimetry data rather than assuming elevation change alone is responsible for the observed signal.

Appendix A

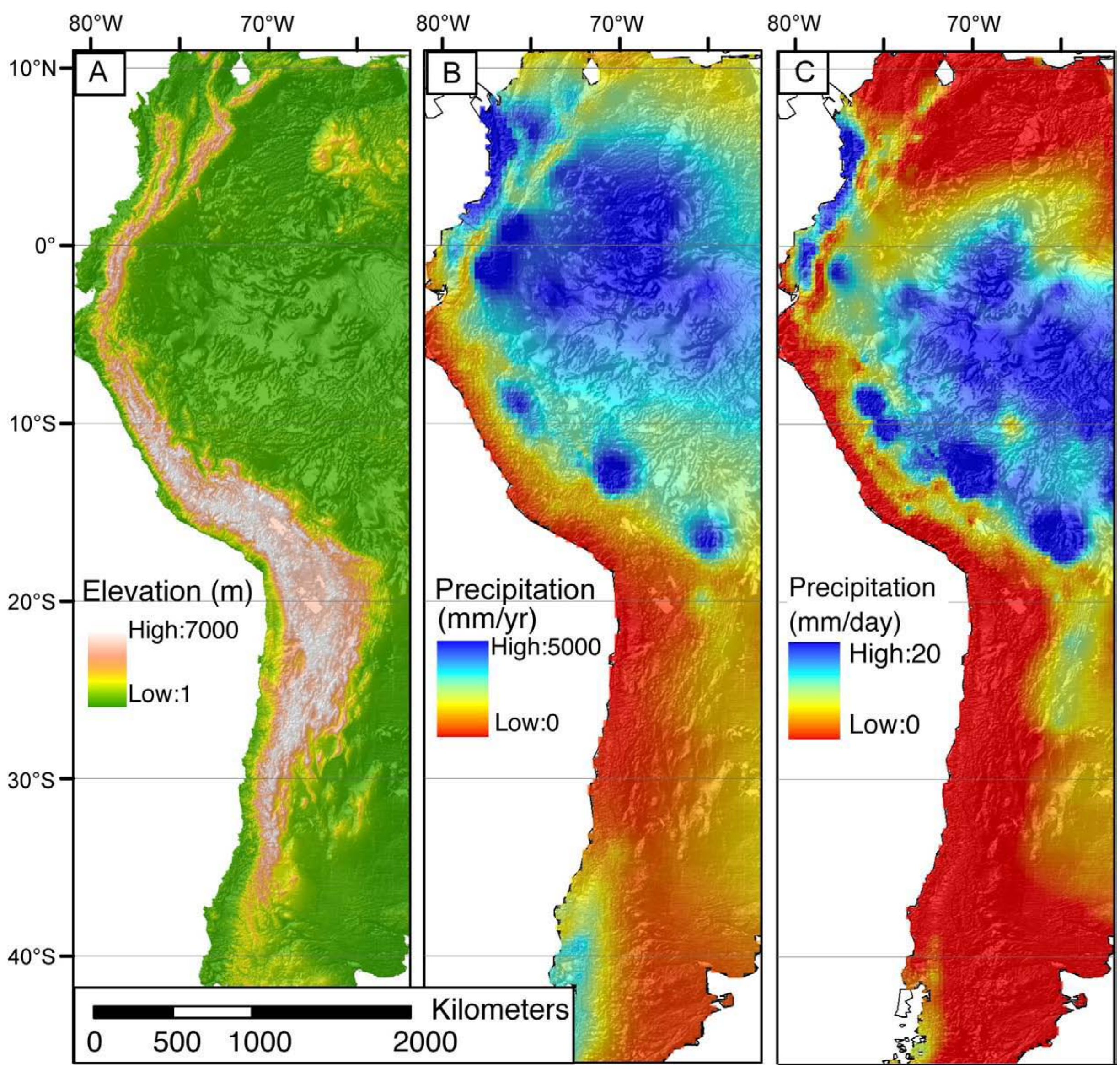

Fig. A1. (A) Modern topography of western South America based on a 90-m digital elevation model. (B) Modern observed mean-annual precipitation (mm yr^{-1}). (C) Modern observed summer (December–January–February) precipitation (mm day^{-1}). Note that RegCM simulates reasonably well the magnitude and spatial distribution of the observed summer precipitation (compare Figs. 2 A and A1C).

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