Neogene climate change and uplift in the Atacama Desert, Chile

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ABSTRACT

The relationship between Andean uplift and extreme desiccation of the west coast of South America is important for understanding the interplay between climate and tectonics in the Central Andes, yet it is poorly understood. Here we use soil morphological characteristics, salt chemistry, and mass independent fractionation anomalies ($\Delta^{17}O$ values) in dated paleosols to reconstruct a middle Miocene climatic transition from semiaridity to extreme hyperaridity in the Atacama Desert. Paleosols along the southeastern margin of the Calama Basin change from calcic Vertisols with root traces, slickensides, and gleyed horizons to an extremely mature salic Gypsisol with pedogenic nitrate. We interpret this transition, which occurred between 19 and 13 Ma, to represent a change in precipitation from >200 mm/yr to <20 mm/yr. This drastic reduction in precipitation likely resulted from uplift of the Central Andes to elevations >2 km; the uplift blocked moisture from the South American summer monsoon from entering the Atacama. The mid-Miocene Gypsisol with pedogenic nitrate is located at elevations between 2900 and 3400 m in the Calama Basin, significantly higher than modern nitrate soils, which occur below ~2500 m. Modern and Quaternary soils in this elevation zone contain soil carbonate and lack pedogenic gypsum and nitrate. We infer that >900 m of local surface uplift over the past 10 m.y. displaced these nitrate paleosols relative to modern nitrate soils and caused a return to wetter conditions in the Calama Basin by decreasing local air temperatures and creating an orographic barrier to Pacific air masses.

Keywords: Atacama Desert, Andes, paleosols, Calama Basin, soil nitrate.

INTRODUCTION

The extreme aridity of the Atacama Desert in northern Chile has been associated with the uplift of the Central Andes and the subsequent blocking of moisture from the Amazon during the mid-Miocene (e.g., Alpers and Brimhall, 1988; Sillitoe and McKee, 1996). Recent paleoclimatic studies from the Atacama, however, have questioned this relationship. Hartley and Chong (2002) suggested a late Pliocene age for the initiation of hyperaridity and argued that desiccation of the west coast of South America was unrelated to Andean uplift. Dunai et al. (2005) proposed that hyperaridity commenced during the late Oligocene (ca. 25 Ma) and that this aridity could have been a significant contributor to the regional uplift in the Cordillera, as proposed by Lamb and Davis (2003). Hoke et al. (2004), however, identified a major shift toward more arid conditions ca. 10 Ma and suggested that desiccation at that time was related to the uplift of the Altiplano plateau. In order to determine the relation and potential feedbacks between Central Andean uplift and climate change, we need records of Neogene precipitation and elevation. Here we present evidence from paleosols in the Calama Basin for extreme hyperaridity by ca. 12 Ma along the eastern margin of the Atacama Desert. As extreme hyperaridity in the Atacama today is maintained through the existence of a strong orographic rain shadow, the mid-Miocene initiation of hyperaridity has implications for the uplift history of the Andes.

STUDY AREA

The Atacama Desert is between the central Andes and Pacific Ocean in northern Chile (Fig. 1). Several factors produce the extreme hyperaridity of the Atacama today, including the Andean rain-shadow effect, the coastal temperature inversion, and the latitudinal position of this region (Houston and Hartley, 2003). The hyperarid core of the Atacama receives <3 mm/yr precipitation, does not support vascular plants, and contains soils with high concentrations of soluble salts, including sulfates, chlorides, and nitrates. We use the term extreme hyperaridity to refer to comparable climatic conditions in the geologic record.

Occasional precipitation events in the Atacama generally result from Pacific air masses that migrate northward from the westerly precipitation belt. Along the eastern margin of the Atacama (~2500 m), precipitation is >20 mm/yr and is associated with the South American summer monsoon (SASM). SASM air masses spill over the central Andes and generate precipitation on the eastern Atacama at elevations above ~2800 m, but do not cause rainfall in the central Atacama.

The Calama Basin is located on the eastern margin of the Atacama, ~150 km from the Pacific Coast at elevations between 2200 and 3500 m (Fig. 1). Precipitation in the center of the basin (2200 m) is ~4 mm/yr, whereas along the eastern margin (3350 m) precipitation is ~50 mm/yr.

PALEOSOLS IN THE CALAMA BASIN

We examined Miocene strata and Quaternary landforms along the southeastern margin of the Calama Basin for evidence of pedogenesis. Miocene gypcretes were first reported in this region by Hartley and May (1998). We identified Miocene and Quaternary paleosols developed on substrates of alluvial fan and flood-plain deposits, and basement bedrock.

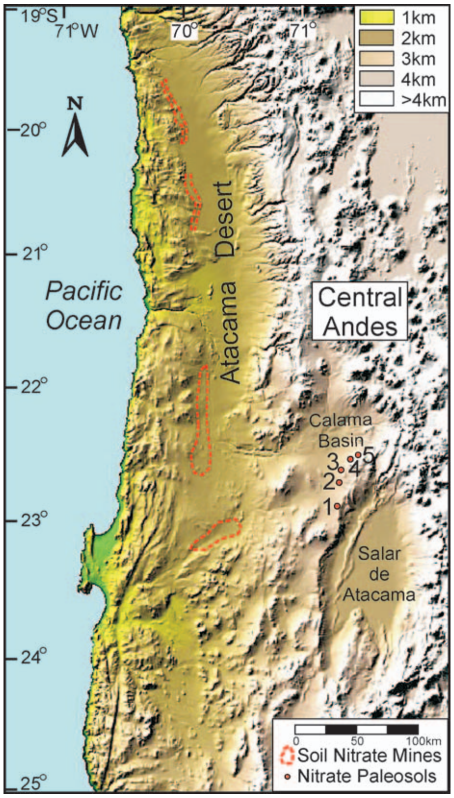

Figure 1. Location of Atacama Desert, Calama Basin, and soil nitrate mines in northern Chile. Red circles (1–5) identify locations of stratigraphic sections (Fig. 2). Shaded relief digital elevation model was created by NASA/JPL/NIMA with data from Shuttle Radar Topography Mission.

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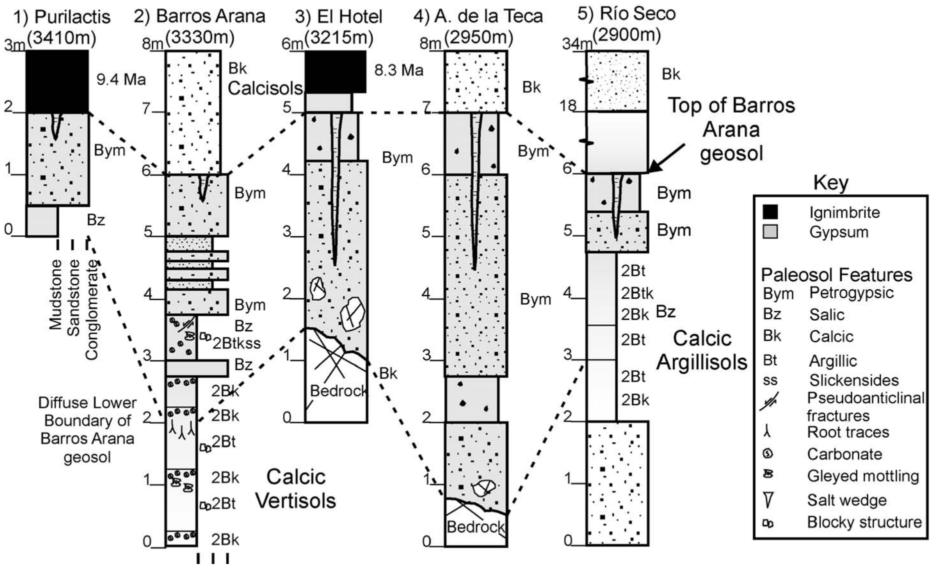

Figure 2. Stratigraphic sections of Neogene paleosols from southeastern margin of Calama Basin. Bym—petrogypsic horizon.

All paleosols were classified using the system of Mack et al. (1993). Four classes of paleosols are present. Middle-upper Miocene strata contain calcic Vertisols and calcic Argillisols, which are overlain by a thick, well-developed salic Gypsisol in all locations examined. Calcisols are present within Quaternary deposits and geomorphic surfaces.

The oldest paleosols we investigated are interbedded in the lower-middle Miocene El Loa Formation (ca. 20 Ma). These paleosols are exposed at the Barros Arana and Río Seco localities (Fig. 2) and consist of calcic Vertisols and calcic Argillisols, respectively. The calcic Vertisols are ∼1 m thick and contain red argillic horizons with montmorillonite and illite clay minerals. Argillic horizons have angular blocky structure with clay skins, slickensides, and pseudoanticlinal fractures. These soils also have soil carbonate (stage II; 1–3 cm nodules), gleyed mottling and gleyed horizons (5–15 cm thick), and root traces with manganese oxides and organic matter. At the Río Seco locality, calcic Argillisols occur in a gravelly parent material. These paleosols have diffuse calcic horizons (stage I) and argillic horizons composed of montmorillonite and illite.

A well-developed salic Gypsisol, hereafter defined as the Barros Arana geosol, is preserved across the southeastern margin of the Calama Basin (Figs. 1 and 2). This paleosol formed on top of basement bedrock (Paleozoic volcanics) at the El Hotel and Agua de la Teca localities, and in Miocene alluvial fan deposits at other localities (Fig. 2). The Gypsisol is >3 m thick and contains a 1.5–6-m-thick petrogypsic horizon (Bym). The petrogypsic horizon has a high bulk density, >2.0 g/cm^2, and contains 10–45 wt% SO_4, or ∼20%–90% gypsum (Fig. 3). Sulfate concentrations are highest at the top of the paleosol profile, where clasts commonly float in a gypsum matrix. The paleosol has large v-shaped salt fractures, or sand dikes, that are as much as 35 cm across and 2 m deep. These fractures are generally filled with eolian silts and sands as well as salts, and are similar to fractures in modern Atacama salic soils (Ericksen, 1981). Petrographic analysis of the Bym horizon shows a complex history of precipitation and dissolution of pedogenic salts, also analogous to modern soils in the Atacama. Between 1.5 and 3.5 m depth in the salic Gypsisol, and in some locations superimposed on top of older paleosols, is a salic horizon that contains as much as 2.5% NO_3, 0.5% Cl, and trace amounts of perchlorate (Fig. 3).

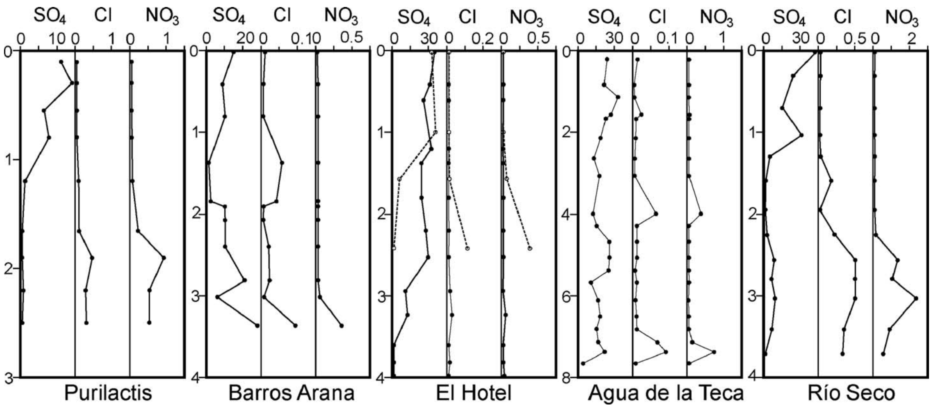

Figure 3. Anion concentrations with depth of soluble salts from top of Barros Arana geosol from the five measured localities identified in Figure 2.

Soils on top of Quaternary landforms along the southeastern margin of the Calama Basin are Calcisols that contain pronounced calcic or petrocalcic horizons (stage II to stage IV development within ∼25 cm of the surface), but lack argillic, gypsic, and salic horizons. These soils generally occur on the surface of alluvial fans and fluvial terraces.

DISCUSSION

Calcic Vertisols with gleyed horizons and root traces form in poorly drained, vegetated flood plains with seasonal precipitation. These soils do not occur in the Atacama today, but are present to the south in central Chile (∼lat 28°–34°S). The calcic Argillisols, with a coarse-grained parent material, form in basin margin environments (alluvial fan and bajadas) where the water table is lower.

The Barros Arana geosol formed across the southeastern margin of the Calama Basin, on top of pediments, alluvial fans, and bajadas. This paleosol was originally described by Hartley and May (1998, p. 361), who interpreted it as "a subsurface crust formed by a combination of hydromorphic and illuvial processes and subject to periodic exhumation and weathering." This interpretation was based primarily on evidence of surficial exposure (salt fractures) and hydromorphic precipitation of gypsum (poikilitic textures). Poikilitic textures, however, are common in salic soils in the Atacama, most of which form in regions with low water tables and are not influenced by hydromorphic processes. Poikilitic textures in these soils likely result from dissolution and recrystallization of soil salts produced by meteoric waters that infiltrate and evaporate in pore spaces and fractures. These pedogenic salt textures, although recognized in previous studies, have caused confusion and debate concerning the origin and genesis of Atacama nitrate soils (e.g., Searl and Rankin, 1993; Ericksen, 1994). We interpret the formation of the Barros Arana geosol to be comparable to modern salic soils in the Atacama (Ericksen, 1981). These soils generally occur on stable landscape surfaces such as alluvial fans and pediments along basin margins, similar to the interpreted paleolandscape setting of the Barros Arana geosol. The majority of regions with well-developed salic horizons do not have high water tables and are not influenced by upward capillary migration of soluble salts. Rather, salts originate from eolian dust that is translocated down into the soil profile during rare precipitation events (Rech et al., 2003). In addition, the Barros Arana geosol has many of the characteristics of extremely mature nitrate soils in the modern Atacama, including large vertical salt fractures, angular salt-shattered gravels floating in a gypsum matrix, and a comparable distribution of soluble salts (i.e., high concentrations of near-surface sulfates and greater concentrations of chlorides and nitrates at depth; Ericksen, 1981; Rech, unpublished data).

Although the duration of soil development for modern nitrate soils in the Atacama is not precisely known, a maximum duration of ~5–10 m.y. has been postulated on the basis of Ar/Ar ages of underlying tuffs (Alpers and Brimhall, 1988; Sillitoe and McKee, 1996). The Barros Arana geosol is a paleolandscape surface in the Calama Basin, and therefore represents varying amounts of time at different localities depending on local deposition histories. The geosol is directly overlain by the Sifon and Artola Ignimbrites (Fig. 2), which have been dated as 8.3 and 9.4 Ma, respectively (de Silva, 1989). These ignimbrites provide a minimum age (9.4 Ma) and duration (1.1 m.y.) for soil development. We estimate, based on the degree of pedogenesis relative to modern nitrate soils, that the Barros Arana geosol likely represents several million years (2–5 m.y.) of soil development.

PALEOPRECIPITATION IN THE CALAMA BASIN

Comparison of paleosols in the Calama Basin with modern soils in the Atacama Desert and surrounding region allow us to reconstruct time-averaged precipitation values during periods of soil formation in the Neogene. Vertisols are not found in the Atacama Desert today, but do occur in central Chile (~lat 29°–32°S), where precipitation is ~250 mm/yr. We therefore estimate minimum precipitation values during the formation of these calcic Vertisols to be >200 mm/yr.

We estimate precipitation during the formation of the Barros Arana geosol to have been <20 mm/yr. This estimate is based on the high concentrations of nitrate and the Δ^{17}O anomalies of nitrate. Nitrate within the upper few meters of soils is only known to occur in the world's driest deserts today, such as the Atacama, the dry valleys of Antarctica, and parts of the Namib in Africa. Modern Atacama soils with concentrations of nitrate similar to values observed at depth in the Barros Arana geosol (0.4%–2.4%; Fig. 3) are restricted mostly to localities where precipitation is 5–10 mm/yr. Nitrate concentrations of soils in the hyperarid core of the Atacama are generally greater than this. Mass independent fractionation anomalies of oxygen (Δ^{17}O) in soil nitrate also allow us to constrain paleoprecipitation values (Michalski et al., 2004). The Δ^{17}O values of soil nitrate measures the relative proportion of nitrate derived from photochemical reactions in the atmosphere (+23‰; Michalski et al., 2003) to nitrate formed or modified by moisture-dependent microbial processes (0‰; Michalski et al., 2004). Soil nitrate in the hyperarid core of the Atacama, where precipitation is <3 mm/yr, has Δ^{17}O values that range from 13.7‰ to 21.6‰ (mean 17.6‰) (Michalski et al., 2004). Samples (n = 10) of nitrate from the Barros Arana geosol have values that range from 4.6‰ to 14.4‰ (mean 9.9‰), suggesting slightly wetter conditions than occur today in the hyperarid core of the Atacama (see GSA Data Repository Table DR1[1]). Combined, soil nitrate concentrations and Δ^{17}O anomalies of the Barros Arana geosol suggest paleoprecipitation values of 5–10 mm/yr.

We estimate precipitation during the formation of Quaternary Calcisols in the Calama Basin by using modern precipitation values and precipitation values inferred from Quaternary rodent middens. Modern precipitation ranges from ~30 to 50 mm/yr at the elevations of the Miocene paleosols (2900–3400 m). Pleistocene rodent middens, however, indicate precipitation between 50 and 100 mm ca. 10 ka (Betancourt et al., 2000). We estimate average precipitation values associated with Pleistocene Calcisols to be ~50 mm/yr.

ANTIQUITY OF THE ATACAMA

The transition from ca. 20 Ma calcic Vertisols with root traces, soil carbonate, and gleyed horizons to the ca. 13–8 Ma Barros Arana geosol with pedogenic nitrate clearly marks the onset of extreme hyperaridity, with precipitation <20 mm/yr, in the Calama Basin. This middle Miocene age for the onset of extreme desiccation in the Atacama Desert broadly supports ages for the initiation of hyperaridity inferred from the cessation of supergene mineralization and erosion of ca. 15 Ma in northern Chile (Alpers and Brimhall, 1988; Sillitoe and McKee, 1996) and is slightly older than proposed ages of 10 and 9 Ma for relict geomorphic landforms in the Atacama (Hoke et al., 2004; Nishiizumi et al., 2005).

Our results do not support the late Oligocene age for the cessation of erosion proposed by Dunai et al. (2005) on the basis of ^{21}Ne concentrations in fluvial gravels, nor with the late Pliocene age postulated by Hartley and Chong (2002) based on the sedimentological record. Hartley and Chong (2002) concluded that Miocene fluvio-lacustrine and alluvial fan deposits in the Atacama are indicative of a semiarid climate, the formation of thick evaporites after 6 Ma identifies a phase of increased aridity, and the cessation of fluvio-lacustrine and alluvial fan sedimentation between 4 and 3 Ma indicates the beginning of hyperaridity. Unfortunately, there is not a good correlation between these depositional environments and precipitation in the Atacama today. Fluvio-lacustrine sediments are being deposited across the modern Atacama in deposystems supported by groundwater recharged in the high Andes (Rech et al., 2002). Hence, these wet depositional environments bear no direct climatic implications for the valley settings in which they are being deposited. Evaporite salars are found in many locations in the hyperarid Atacama, but also occur throughout the Altiplano in regions that are much colder and receive as much as 300 mm/yr precipitation (Stoertz and Ericksen, 1974).

ANDEAN UPLIFT AND CLIMATE CHANGE

We suggest that the initiation of hyperaridity in the Calama Basin, which occurred between ca. 19 and 13 Ma, was the result of uplift of the Andes to elevations that were high enough to block moisture entrained in the SASM from entering the Atacama. We estimate that a minimum Andean paleoelevation of ~2 km was necessary to cause extreme hyperaridity along the eastern margin of the Calama Basin. This estimation is based on examination of modern rainfall between lat 22° and 25°S across the Central Andes, which shows a significant decrease in precipitation at an elevation of ~2 km on the eastern slope of the Andes (Fig. DR1; see footnote 1). Atmospheric modeling of precipitation along the eastern flank of the Andes also indicates that most moisture is precipitated out of air masses at elevations <2 km (Masek et al., 1994). This middle Miocene paleoelevation estimate of ~2 km prior to 12 Ma for the Central Andes suggests a slightly earlier uplift history than those based on the isotopic composition of soil carbonate near lat 18°S (Garzione et al., 2006; Ghosh et al., 2006).

We suggest that the return to wetter condi-

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[1] GSA Data Repository item 2006159, Table DR1, Δ^{17}O values of soil nitrate, and Figure DR1, precipitation data for the Central Andes, is available online at www.geosociety.org/pubs/ft2006.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

tions in the Calama Basin, evidenced by Quaternary Calcisols, rodent middens, and modern precipitation values that are greater than middle Miocene paleoprecipitation estimates, was the result of local surface uplift. The upper elevation limit of modern nitrate soils along the eastern margin of the Atacama is $\sim$2500 m. At the Purilactis locality, the Barros Arana geosol crops out at an elevation of 3410 m. This suggests a minimum upward displacement of 900 m over the past 9.4 m.y. This minimum uplift estimate is in accord with river-profile data from the western flank of the central Andes (18.5°–22°S) that indicate $\sim$1 km of post–10 Ma uplift (Hoke et al., 2005). As late Neogene uplift developed in the absence of significant local upper-crustal deformation along the western flank of the Andes, it is likely driven by ductile thickening of the lower crust and convective removal of the underlying lithospheric mantle (Isacks, 1988; Lamb et al., 1997; Garzione et al., 2006).

Several factors associated with uplift likely contributed to the wetter conditions observed in the Calama Basin during the Quaternary. Decreased air temperatures would have increased effective moisture. Continued uplift also would have created an orographic barrier capable of intercepting air masses that migrate northward from the southern Westerlies.

CONCLUSIONS

The initiation of extreme hyperaridity along the eastern margin of the Atacama Desert is clearly recorded by the transition from calcic Vertisol soils with root traces to salic (nitrate) Gypsisols in Miocene strata of the Calama Basin. This transition represents a change in precipitation from $>$200 mm/yr to $<$20 mm/yr and occurred well before 9.4 Ma. We interpret this dramatic climate change to have taken place between 19 and 13 Ma and suggest that it was the result of Andean uplift to a paleoelevation ($>$2 km). Although the initiation of hyperaridity could also be the result of global climate cooling and the strengthening of the Humboldt Current at that time, hyperaridity could not have developed along the eastern margin of the Atacama without a strong Andean rain-shadow effect in place. Therefore, the Barros Arana geosol places a minimum age for the development of a strong Andean rain shadow in the Central Andes, which we estimate to be $\sim$2 km.

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