Early Cenozoic uplift of the Puna Plateau, Central Andes, based on stable isotope paleoaltimetry of hydrated volcanic glass

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ABSTRACT

Uplift of the Central Andes is largely thought to have occurred during the past 10 m.y. based on paleoaltimetry studies from the Altiplano of Bolivia. However, the spatio-temporal uplift history may not be uniform across the Central Andes. We present new stable isotopic results from the Salar de Antofalla, Salina del Fraile, and Arizaro Basin on the Puna Plateau (24°–26°S) of northwestern Argentina. Samples of volcanic glass give δDglass values and modeled paleoaltitudes that indicate an uplifted (~4 km) Puna Plateau since ca. 36 Ma with limited (<1 km) changes in elevation since then and, unlike the Altiplano, no evidence of large-magnitude uplift during the Miocene.

INTRODUCTION

Today the Central Andes constitute the largest barrier to zonal moisture transport on Earth, having a profound effect on South American precipitation patterns (Strecker et al., 2007). Their uplift would have had a dramatic impact on regional climate and faunal and floral diversification (Ehlers and Poulsen, 2009; Hoorn et al., 2010). Thus, an outstanding issue not only in the field of Andean geodynamics but also in the fields of climate sciences, biology, and ecology is the timing of uplift of the Andes. From a geological perspective, understanding the timing of uplift could help in evaluating the importance of a variety of mountain-building processes, such as arc magmatic activity (Kay et al., 1994), crustal thickening in the retro-arc region (Kley and Monaldi, 1998; McQuarrie et al., 2005), and cyclical lithospheric removal (DeCelles et al., 2009).

The timing and mechanism of uplift of the Central Andes remains hotly debated. Stable isotope and paleobotanical methods of paleoaltimetry used in the Altiplano (15°–20°S, Fig. 1) suggest a 2.5–3 km increase in elevation between ca. 10.3 and 6.8 Ma as a result of wholesale removal of mantle lithosphere (Gregory-Wodzicki, 2000; Garzione et al., 2008). Isotopic evidence from the Bolivian Eastern Cordillera (Leier et al., 2013) finds more spatial and temporal (between 24 Ma and 15 Ma) variability in the history of elevation change. Isotopic evidence of seasonal precipitation during the Miocene in the Subandean foreland is used as evidence of the deflection of the low-level jet due to uplift of the Eastern Cordillera between 10 Ma and 8 Ma (Mulch et al., 2010). In contrast, climate models (Ehlers and Poulsen, 2009) have been used to argue for a more gradual uplift of the Andes since 25 Ma. A major impediment to understanding the paleoelevation history of the Central Andes is the lack of data from the Puna Plateau, which forms the southern half of the Central Andes (24°–26°S, Fig. 1). We present a stable isotope paleoaltimetry study of the Puna Plateau, and demonstrate that elevations of ~4 km have persisted in this region since 36 Ma.

This observation has significant implications for the causes of high elevation throughout the Central Andes.

GEOLOGICAL AND CLIMATIC SETTING OF THE CENTRAL ANDES

The Central Andes include the Western Cordillera (modern arc), the Central Andean Plateau (Altiplano and Puna Plateau), and the Eastern Cordillera (Fig. 1). Moisture drawn from the Amazon lowlands and Paraguayan Chaco Basin produces intense rain-out (>2 m/yr) below 3 km elevation along the eastern flank of the Andes, and aridity westward into the Andean interior (Strecker et al., 2007; Ehlers and Poulsen, 2009) (Fig. 1). On average, the Puna Plateau is higher (4400 m above sea level [masl]), drier, and more rugged than the Altiplano (3800 masl) today (Allmendinger et al., 1997; Strecker et al., 2007; Garzione et al., 2008; Fig. 1). Despite these modern differences, the two regions share similar deformation histories. Both the Altiplano and Puna Plateau were occupied by a regional foreland basin during the early Cenozoic, and transformed into the modern Andean orogen through crustal shortening and magmatic activity from the Paleogene onward (Sempere et al., 1997; DeCelles and Horton, 2003; DeCelles et al., 2011; Carrapa et al., 2011). Geophysical and geochemical studies suggest that the mantle lithosphere beneath the Puna Plateau and Altiplano is gravitationally unstable and subject to local or regional-scale foundering events (Kay et al., 1994; Bianchi et al., 2013; Duca et al., 2013). Crustal thickness is highly variable (45–75 km) under the Puna Plateau and is commonly thinner compared to that under the Altiplano (60–80 km) (Yuan et al., 2002; Bianchi et al., 2013) (Fig. 1). Thicker crust and greater shortening estimates for the Altiplano and Eastern Cordillera of Bolivia (McQuarrie et al., 2005; Kley and Monaldi, 1998) suggest that if shortening alone were responsible for uplift, elevations should be higher in the Altiplano. That this is not the case suggests that other processes, such as underplating and lithospheric removal, may be controlling regional elevation differences between the Puna Plateau and Altiplano.

STABLE ISOTOPE PALEOALTIMETRY

Stable isotope paleoaltimetry is based on the observation that precipitation at higher elevations is progressively more depleted in both 18O and deuterium (D) isotopes due to orographic uplift and associated Rayleigh distillation (Rowley, 2007). To quantify paleoaltitudes on the Puna Plateau, deuterium isotope ratios of hydrated volcanic glass (δDglass) from 33 volcanic tuff and ignimbrite samples were determined from the Salar de Antofalla (~3600 masl), Salina del Fraile (~3700 masl), and Arizaro Basin (~3900 masl) (Fig. 1). The host deposits are fluvial and eolian sandstones, and perennial and ephemeral lacustrine shales with a total stratigraphic thickness of more than 1000 m in the Salina del Fraile area and 3000 m in the Arizaro Basin (Figs. DR1–DR8 in the GSA Data Repository1). New U-Pb zircon ages from 23 sampled tuffs in the Salar de Antofalla and Salina del Fraile areas range from 37.1 ± 8.4 Ma to 0.5 ± 0.5 Ma (Tables DR1 and DR2 in the Data Repository). Previous U-Pb zircon ages from tuffs in the Arizaro Basin range from 34.8 ± 2.3 Ma to 0.4 ± 1.0 Ma (Boyd, 2010). Combined, these data provide a stable isotope paleoaltimetry record from late Eocene (ca. 37 Ma) to modern for the Puna Plateau.

1GSA Data Repository item 2014157, detailed methods and discussion supported by 14 figures and 2 data tables, is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.
The δD of Volcanic Glass

Primary volcanic glass contains 0.1–0.3 wt% magmatic water. Within 1000–10,000 yr after eruption, it becomes hydrated by meteoric water until it reaches ~5 wt% water (Friedman et al., 1993; Cassel et al., 2012). During hydration, hydrolysis and proton (H+) exchange results in a bonded silica “gel” layer which seals the isotopic composition of initial hydration waters (Cailleteau et al., 2008; Mulch et al., 2008; Cassel et al., 2012; Dettinger, 2013).

We found a large range in δD_{glass} values (Fig. 2; Table DR1), from −127.7‰ ±3.3‰ to −67.9‰ ± 2.4‰. This spread in δD_{glass} values is greater than that found today (−125‰ to −74‰) based on δD_{glass} values from modern (<0.55 Ma) samples (n = 7) collected on the Puna Plateau from 2973 masl to 4757 masl (Dettinger and Quade, 2014). This greater range in δD_{glass} values likely reflects varying contributions of different source water (evaporatively enriched surface waters, snow melt, or high-elevation waters depleted in deuterium). Results from two samples (Eastash 1 and Eastash 2) have especially high δD_{glass} values, ~6‰ higher than the upper range of modern glass from the Puna Plateau (Fig. 2). These values are likely the result of glass hydration by evaporatively enriched surface waters, producing underestimated elevations. These scenarios are plausible as Eastash 1 is from the upper Miocene Sijes Formation, which contains numerous evaporites, and Eastash 2 is from a bedded tuff in Quaternary alluvium. A similar scenario may account for the higher-than-expected δD_{glass} values for one young Arizaro Basin sample (ARB09-9, −80.4‰ ± 2.2‰, 0.4 ± 1.0 Ma), which was also collected from Quaternary alluvium, although its δD_{glass} value is still within the range of other modern glasses (Fig. 2).

Water content is highly variable in our glass samples with most falling between ~4 and 12 wt% (Fig. DR9). Samples 3SF4, 6SF664, and 9SF3.8 have H_2O >10 wt% and similar H concentration to a kaolinite standard (Figs. DR9–DR12), suggesting contamination by adhering clay particles, and therefore were not used to interpret paleoelevations. The δD_{glass} values were converted to δD_{paleowater} (δD_{pw}) values using the empirically derived fractionation factor (α) of 1.0343 published by Friedman et al. (1993). This fractionation factor has been applied successfully in previous paleoelevation studies using hydrated volcanic glass (e.g., Mulch et al., 2008).

Figure 1. A: Geologic map of Salar de Arizaro, Arizaro Basin, Salar de Antofalla, and Salina del Fraile region of Puna Plateau, northwestern Argentina. Localities for hydrated volcanic glass samples are depicted. For locations of measured stratigraphic sections where more than one sample was collected, section number is indicated. These include sections in Salina del Fraile: 2SF (n = 5), 6–7SF (n = 4), 3–4, 8–9 SF (n = 4); and in Salar de Antofalla: 2LQ (n = 2) and 1SJ (n = 2). B: Shaded relief map with mean annual precipitation and major tectonomorphic provinces, modified after Strecker et al. (2007). S.B. Ranges east of the Puna shows the Santa Barbara Ranges. Black box denotes area mapped in A. White lines indicate locations of cross sections in C and D. C,D: Cross sections showing topography and subsurface structural differences between Altiplano and Puna plateaus (Gregory-Wodzicki, 2000; Allmendinger et al., 1997; Kay et al., 1994; Yuan et al., 2002; Bianchi et al., 2013). Units are in kilometers; note change in vertical scale at sea level.
Using this fractionation factor, our δDfw values range from −97.8‰ ± 3.4‰ to −35.9‰ ± 2.5‰, which is a greater range in δD values than what is found today (−75.1‰ to −40.3‰) in sampled meteoric waters (Dettinger, 2013). See the Data Repository for further discussion of weight percent H2O and the fractionation factor.

**Modeled Paleoelevations**

Predictions of paleoelevation are made using an atmospheric thermodynamic model based on a change in δ18Oprecipitation between low-altitude and high-altitude locations (isotopic lapse rate), written as Δ(δ18O) (Rowley, 2007). This model has been used to reconstruct paleoelevations for the Sierra Nevada (California), Colorado Plateau, Altiplano, and Himalaya (Cassel et al., 2012; Mix et al., 2011; Rowley and Garzione, 2007). Due to the large error inherent in the model because of its theoretical basis, paleoelevation estimates should be interpreted as a potential range as opposed to absolute elevation. The strength of the Rowley (2007) model is that it can incorporate different temperatures (T) and relative humidities (RH), making it useful for time periods for which an isotopic lapse rate is unknown. However, it is important to choose appropriate values of RH and T to reduce error in the paleoelevation results. Because a major hydrologic shift occurs at the Eocene-Oligocene boundary as Earth moved from a greenhouse climate toward the cooler and drier climate of today, we used the Rowley (2007) model with Eocene parameters (RH = 80% ± 3%, T = 300 K) modeled from Huber and Caballero (2003). To estimate paleoelevations for Oligocene and younger samples, we use the modern model from Rowley (2007) which utilizes all possible modern values of starting temperature and relative humidity for Δδ18O between 0‰ and −25‰.

In order to generate the model’s Δδ18O term, we followed the method of Mix et al. (2011), which involves converting δD values to δ18O values using the global meteoric water line (GMWL) (Craig, 1961). In an effort to minimize error associated with this conversion, we modified the method by using a local meteoric water line (LMWL: δD = 88δ18O + 12.6) for northwestern Argentina from Dettinger (2013). We opted to use the LMWL for the Oligocene to recent samples and the GMWL for the Eocene samples, because the “icehouse” conditions of the Oligocene and younger time periods are closer to today’s conditions than the “greenhouse” conditions of the Eocene. In addition, model results suggest that the Eocene GMWL was very similar to that of today (Speelman et al., 2010).

To account for changing climate throughout the Cenozoic, we use epoch-specific low-elevation modeled or fossil δ18O and δD values to generate the Rowley (2007) model’s Δδ18O term. These low-elevation values were chosen based on their proximity to the assumed source of moisture, which today is the subtropical Atlantic (Ehlers and Poulsen, 2009). We assume the same source in the past because South America has maintained a relatively constant latitude throughout the Cenozoic (Sdrolias and Müller, 2006). See the Data Repository for further discussion of MWLs and moisture source.

For the Eocene, we used a modeled δDfw value of −20‰ from Speelman et al. (2010) as our low-altitude input, which converted to a δ18Ofw value of −3.8‰ using the GMWL. Rodent fossils from <1 km elevation in Brazil provided a δ18Ofw value of −3‰ (Bershaw et al., 2010) for the Oligocene, and a modeled δ18Ofw value of −9‰ was used for the Miocene (Poulsen et al., 2010). For both the Pliocene and Quaternary samples we used an average δ18Ofw value of −2.4‰ from three coastal Brazilian Global Network of Isotopes in Precipitation (GNIP) stations (Belém, Salvador, and Rio de Janeiro; IAEA/WMO, 2006). Any error in model results due to using this modern average for Pliocene samples is within the limits of uncertainty of the model itself.

Modeled mean elevation estimates indicate a Puna Plateau height of between 3.4 ± 1.7 km and 5.3 ± 2.8 km in the late Eocene. For the most part, these upper Eocene samples come from the Salina del Fraile and Salar de Antofalla, with only one sample from the Arizaro Basin, which gives a much lower average paleoelevation estimate; however, given the ±2.3 m.y. error for this sample’s U-Pb zircon age, it may be more appropriate to group it with the Oligocene sample subset. Paleoelevation estimates are between 3.2 ± 1.4 km and 3.7 ± 1.5 km in the Oligocene, up to 4.7 ± 1.8 km in the Miocene, and 3.9 ± 1.6 km to 4.1 ± 1.6 km in the Pliocene and Quaternary (very near modern elevations) (Fig. 3).

**DISCUSSION AND CONCLUSIONS**

Paleoelevation estimates from the Puna Plateau presented here indicate high elevation (~4 km) since at least 36 Ma. High elevations could be supported by a thickened crust beneath the Puna Plateau associated with shortening and eastward migration of the orogenic strain front into the westernmost Eastern Cordillera during the late Eocene (Carrapa and DeCelles, 2008) and by magmatic underplating which is thought to have increased crustal thickness below the Puna Plateau from ~30–35 km during the Jurassic to ~45 km by the late Eocene (Hasc hike et al., 2002). Speculations of small-scale (<1 km) uplift events are tenuous because they are within error of the model, but nonetheless, they may be significant as they correlate with other geological evidence for uplift. For example, the Miocene (ca. 25–18 Ma) event (<500 m; Fig. 3) corresponds to deformation and exhumation in the Eastern Cordillera (Coutand et al., 2006; Carrapa et al., 2011), which is consistent with significant crustal thickening and uplift in the region at this time.

Unlike the Bolivian Altiplano and Eastern Cordillera, there is no evidence of large-magnitude uplift of the Puna Plateau during the Miocene (Garzione et al., 2008; Leier et al., 2013), and small-scale lithospheric removal under the Puna Plateau (Ducea et al., 2013) can only account for limited uplift (<1 km). It is clear from our results that there are important along-strike variations in uplift between the Puna Plateau and Altiplano. In turn, this implies a complex lithospheric structure under the Central Andes as monitored by geophysical data (Yuan et al., 2002; Bianchi et al., 2013). High elevation since 36 Ma of part of the Central Andes should be
incorporated into future models of Andean tectonics, geodynamics, and paleoclimate.

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