Does the topographic distribution of the central Andean Puna Plateau result from climatic or geodynamic processes?

M.R. Strecker[1]*, R. Alonso[2], B. Bookhagen[3], B. Carrapa[4], I. Coutand[5†], M.P. Hain[1], G.E. Hilley[5], E. Mortimer[6], L. Schoenbohm[7], and E.R. Sobel[1]

[1]Institut für Geowissenschaften, Universität Potsdam, 14476 Potsdam, Germany
[2]Universidad Nacíonal de Salta, 4400 Salta, Argentina
[3]Geography Department, University of California–Santa Barbara, Santa Barbara, California 93106, USA
[4]Department of Geology, University of Wyoming, Laramie, Wyoming 82071, USA
[5]Université des Sciences et Technologies de Lille 1, Laboratoire Géosystèmes, 59655 Villeneuve d'ascq, France
[6]School of Earth and Environment, Leeds University, Leeds LS21 1AH, UK
[7]School of Earth Sciences, The Ohio State University, Columbus, Ohio 43210, USA

ABSTRACT

Orogenic plateaus are extensive, high-elevation areas with low internal relief that have been attributed to deep-seated and/or climate-driven surface processes. In the latter case, models predict that lateral plateau growth results from increasing aridity along the margins as range uplift shields the orogen interior from precipitation. We analyze the spatiotemporal progression of basin isolation and filling at the eastern margin of the Puna Plateau of the Argentine Andes to determine if the topography predicted by such models is observed. We find that the timing of basin filling and reexcavation is variable, suggesting nonsystematic plateau growth. Instead, the Airy isostatically compensated component of topography constitutes the majority of the mean elevation gain between the foreland and the plateau. This indicates that deep-seated phenomena, such as changes in crustal thickness and/or lateral density, are required to produce high plateau elevations. In contrast, the frequency of the uncompensated topography within the plateau and in the adjacent foreland that is interrupted by ranges appears similar, although the amplitude of this topographic component increases east of the plateau. Combined with sedimentologic observations, we infer that the low internal relief of the plateau likely results from increased aridity and sediment storage within the plateau and along its eastern margin.

INTRODUCTION

Orogenic plateaus are extensive high-elevation, low internal relief regions that have been explained by models that invoke geodynamic processes (e.g., Isacks, 1988) or combined deep-seated and geomorphic processes that act to limit erosion (e.g., Masek et al., 1994; Métivier et al., 1998; Sobel et al., 2003). In the former view, density contrasts driven by thermal or compositional effects within the lithosphere and asthenosphere or crustal thickening result in high mean elevations and reduce the internal topographic relief of plateaus (e.g., Fielding et al., 1994). Alternatively, contraction along the plateau margins may uplift ranges that successively incorporate foreland areas into intermontane sedimentary basins, which subsequently fill with sediment as their erosional base level is disconnected from the foreland (Sobel et al., 2003). Where moisture transport is perpendicular to these structures, leeward aridification reduces fluvial efficiency to keep pace with range uplift. This causes progressive defeat of fluvial systems, internal drainage, the overfilling of basins, and ultimately their coalescence (e.g., Métivier et al., 1998; Sobel et al., 2003). Consequently, basins within the plateau interior that were assimilated first should show lower internal relief and thicker basin fills than younger intermontane basins flanking the plateau. In this scenario, the plateau margin migrates systematically outward as basins are progressively isolated, and mean elevations increase to match those of the plateau interior as basins fill and coalesce. This concept may apply to intermontane basins close to the Puna Plateau of northwestern Argentina. These basins are structurally akin to basins within the plateau, but occur in areas of pronounced rainfall gradients. Here we test the model of climate-controlled, erosionally moderated lateral plateau growth by determining the timing of regional-scale uplift patterns, changes in fluvial connectivity, and basin filling with sedimentary and chronostratigraphic data from basins along the Puna Plateau.

GEOLOGIC HISTORY OF THE EASTERN PUNA MARGIN AND THE BROKEN FORELAND BASINS

On average, the Puna is ~3.7 km high, and comprises reverse-fault bounded ranges and closed basins with <5-km-thick sedimentary fills (e.g., Allmendinger et al., 1997). To the east, Cenozoic contractional inversion of structures associated with the Cretaceous Salta Rift province has created a broken foreland, ~200 km wide. This region is between the Subandean fold-and-thrust belt south of lat 23°S and the thick-skinned Sierras Pampeanas at ~26°S (Fig. 1). To the west, this province transitions into the Eastern Cordillera and the Puna (Allmendinger et al., 1983).

Uplift along the present-day margin of the Puna dates back to the Oligocene (Coutand et al., 2001), and since 10 Ma ago this region must have constituted high topography, which shielded the region now corresponding to the plateau from easterly moisture (for reviews, see Marrett and Strecker, 2000; Starck and Anzótegui, 2001; Strecker et al., 2007). The fragmentation of the formerly contiguous foreland into individual intermontane basins by basement uplift is revealed by the timing and nature of conglomeratic basin fills (Fig. 1). These basin fills record the loss or severence of fluvial connectivity with the foreland, and so their chronology allows us to determine if range uplift and basin partitioning vary systematically or randomly with longitude. It also helps illuminate the broad relationships between the timing of basin formation within the high-elevation Puna and the currently low-elevation basins along its margin.

In the Quebrada de Humahuaca, at 24°S (Fig. 1), foreland fragmentation is recorded by the older than 3.5 Ma Maimará Formation, which consists of gypsum-bearing mud, sandstones, and conglomeratic sandstones, some of the clasts of which are sourced from the Puna (Walther et al., 1998). Following erosion, renewed filling of this intermontane basin is recorded by the older than 2.78 Ma Uquía Formation (Marshall et al., 1982). A 0.8 Ma old, >400-m-thick conglomerate fill unit records the relatively recent reduced connectivity of this basin (Strecker et al., 2007). In the Quebrada del Toro, at 24.5°S, two successive aggradation and incision cycles suggest that hydrologic isolation began after ca. 6 Ma ago (Marrett and Strecker, 2000; Hilley and Strecker, 2005). At ~25°S, uplift of the Sierra de Mojotoro was coeval with the folding of 1.3 Ma

*E-mail: strecker@geo.uni-potsdam.de.
†Current address: Department of Earth Sciences, Dalhousie University, Halifax NS, B3H 4J1, Canada.

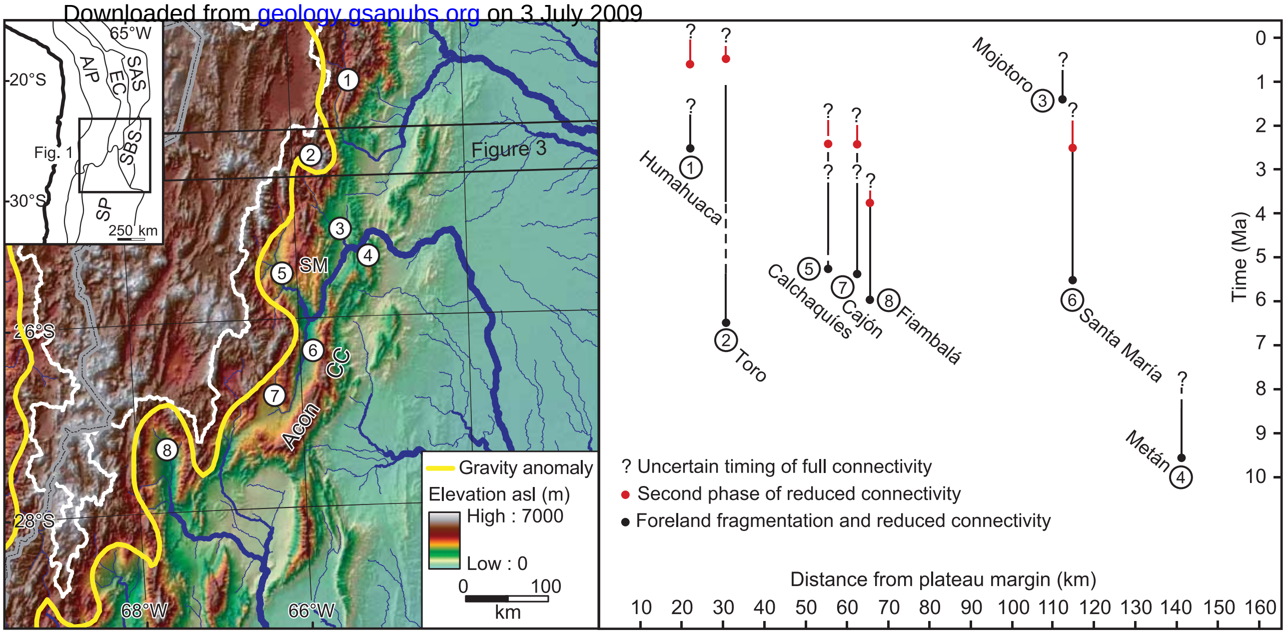

Figure 1. A: Puna Plateau and eastern foreland (left) with spatiotemporal variation of intermontane basin histories (asl—above sea level). White line is margin of internally drained Puna Plateau. Inset outlines major morphostructural provinces of central Andes: SAS—Subandean Belt; EC—Eastern Cordillera; SBS—Santa Barbara system; SP—Sierras Pampeanas; A/P—Altiplano-Puna Plateau. Box denotes location of topographic swath profile shown in Figure 3. Solid yellow line denotes –300 mgal Bouguer gravity anomaly (Tassara et al., 2006). White letters denote locations discussed in text: Acon—Sierra Aconquija; CC—Cumbres Calchaquíes; SM—Sierra de Léon Muerto. B: Basin locations discussed in text, numbered from north to south. Basins are plotted corresponding to average latitudinal position and with respect to distance from Puna border. Black bars show onset of foreland-basin fragmentation and filling of intermontane basins; red bars denote renewed deposition of conglomerates, subsequent to partial removal of basin fills.

old conglomerates (Malamud et al., 1996; Hain, 2008), but unroofing of basement in the Sierra de Metán ~80 km southeast had already begun between 10 and 8 Ma ago, evidenced by an angular unconformity (Cristallini et al., 1997) and clear lithologic provenance signal and westward transport directions (Gonzáles Villa, 2002; Hain, 2008). The intermontane-basin stage in the Valle Calchaquí (25.5°S) was caused by the uplift of the Sierra de León Muerto along the eastern border of the valley, leading to leeward aridification after 5.2 and before 2.4 Ma ago (Coutand et al., 2006; Strecker et al., 2007). Between 26ºS and 27°S the uplift of the Cumbres Calchaquíes (4 km) and Sierra Aconquija (5 km), coupled with the formation of the intermontane Santa María basin, began after 6 Ma ago (Kleinert and Strecker, 2001; Sobel and Strecker, 2003). Transient basin isolation accompanied by basin-wide deposition of thick conglomerates covering tilted and eroded units, 3.4 Ma old, followed after 2.9 Ma ago (Strecker et al., 1989). The intermontane El Cajón basin farther west developed after 5.4 Ma ago (Mortimer et al., 2007). At ~28°S, the Bolsón de Fiambala (Fig. 1) records a broken foreland after 6 Ma ago, and reduced fluvial connectivity by 3.7 Ma ago (Carrapa et al., 2008).

DIFFERENCES BETWEEN THE MORPHOLOGY OF THE PUNA PLATEAU AND ITS EASTERN MARGIN

Nonsystematic basin fragmentation east of the Puna suggests that the plateau has not been growing laterally through progressive assimilation of broken foreland basins (Fig. 1). To elucidate the role that climate and basin isolation may play in forming the plateau's topographic characteristics, we isolated those components of the topography that are Airy isostatically compensated from flexurally supported sectors (Fig. 2). The former components constitute the long-wavelength element of the topography and result from the buoyancy structure within the lithosphere and lithospheric thickness. In contrast, the flexurally supported sectors are composed of topographic features having wavelengths far less than the square of the elastic crustal thickness, and thus may result from shallower processes. In isolating long-wavelength (compensated) components of the topography from short-wavelength components, we assumed an elastic thickness of 15 km (e.g., Tassara et al., 2007), and mantle and crustal densities of 3300 kg/m^3 and 2600 kg/m^3, respectively (see GSA Data Repository[1]). The modeled long-wavelength attributes of the topography of the Puna and

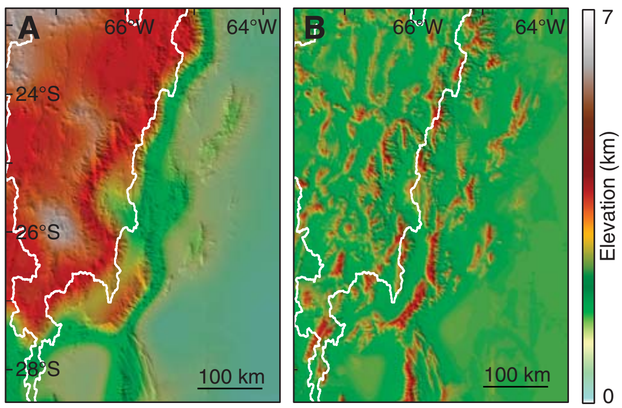

Figure 2. A: Long-wavelength topography of southern Puna. Topography was isolated using two-dimensional Fourier transform flexural model (described in GSA Data Repository; see footnote 1), assuming elastic thickness of crust = 15 km. B: Short-wavelength component of topography was isolated by subtracting observed topography from flexurally compensated topography shown in A. White outline shows location of internal drainage shown in Figure 1.

[1]GSA Data Repository item 2009149, determination of isostatically and non-isostatically compensated components of topography in the Puna and adjacent areas using varying elastic thickness, is available online at www.geosociety.org/pubs/ft2009.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

its flanks show the broad westward rise of plateau topography (Fig. 2A). The high-elevation component of the topography is clearly seen at these low spatial frequencies, suggesting that the high plateau elevations are isostatically compensated. Assuming Airy isostasy with crust and mantle densities of 2600 kg/m^3 and 3300 kg/m^3, respectively, the thick plateau basin fills (e.g., Alonso et al., 1991; Vandervoort et al., 1995) may produce a long-wavelength mean elevation rise that is only about a quarter of the observed elevation difference between the foreland and the plateau (Fig. 2B). Thus, crustal thickening and/or processes affecting the relative buoyancy of the crust appear necessary to explain ~75% of the mean elevation difference between plateau and foreland (e.g., Allmendinger et al., 1997). This is corroborated by the magnitude and spatial extent of low Bouguer gravity anomalies (Fig. 1) that broadly coincide with the plateau (Tassara et al., 2006).

When viewing the modeled short-wavelength, flexurally compensated topographic variations, the morphology of foreland basement ranges extends well into the interior of the Puna. It is often difficult to distinguish the short-wavelength morphology of ranges within the plateau from the broken foreland (Fig. 2B).

The principal difference between the modeled short-wavelength topography within the plateau and along its margins, however, is the amplitude of this topography, which roughly gauges the peak-to-basin relief across the region (Fig. 2B). A west-east topographic swath between 24ºS and 24.5°S highlights the change in short-wavelength relief that accompanies the plateau margin (Fig. 3). Within the plateau interior, peak-to-basin relief is <0.7 km; however, to the east, this relief more than doubles (Fig. 3, dotted line). This transition in relief coincides with the drainage divide between the Puna and the fluvially integrated broken foreland. Within the plateau, erosion products are stored locally within the basins, while within the broken foreland, transiently stored sediments are eventually exported from the orogen (e.g., Sobel et al., 2003).

DISCUSSION AND CONCLUSION

We originally hypothesized that the partially coalesced, high-elevation Puna basins were isolated from the foreland base level and that resulting basin filling increased mean elevations as shortening continued, similar to the predictions for the lateral growth of Tibet (e.g., Métivier et al., 1998) and resulting in a successive eastward expansion of the plateau. Consequently, intermontane sectors closest to internally drained basins in the orogen interior would become increasingly arid, fluvially isolated, and filled with sediment. However, our synopsis of data from the broken foreland of the Puna documents disparate basin filling in space and time that may reflect the diachronous uplift of ranges built along preexisting structural weaknesses (e.g., Hilley et al., 2005). The fact that the earliest fragmentation of the foreland basin occurs ~150 km east of the plateau and the lack of systematic basin filling with proximity to the plateau margin effectively rule out a systematic eastward plateau growth by progressive isolation from the foreland base level. In addition, the modeled isostatically compensated component of the topography comprises most of the elevation variation across the region (Fig. 2A), suggesting that this compensation must be occurring at the scale of the plateau and its margins. While geophysical data beneath the Puna do not yet clearly image the status of the lower crust and upper mantle, some studies suggest at least its partial convective removal (Schurr et al., 2006). It is intriguing that the time at which basins in the broken foreland begin to fill with sediment (ca. 10 Ma ago; Fig. 1B) corresponds with inferred uplift of the surface of the present greater Andean plateau area thought to coincide with convective removal of the lower crust and upper mantle (e.g., Sobolev and Babeyko, 2005; Garzione et al., 2006; Ghosh et al., 2006). Direct measures of the elevation history of the Puna are not available, and given the uncertainties of the timing of basin filling on the plateau (e.g., Vandervoort et al., 1995), it is difficult to determine if a rapid delamination may have caused rapid surface uplift of the plateau while simultaneously fragmenting the foreland basin. Nonetheless, the temporal coincidence of these events makes such removal a possible, if equivocal, mechanism for producing the long-wavelength attributes of the Puna topography.

When isolating the component of the topography that is not isostatically compensated, the frequency of the contractional basin and range topography within the plateau and the eastern broken foreland is similar. As deformation within these two regions was accommodated by a similar structural style, the modern broken foreland may serve as an analog to the Eocene–Oligocene to Miocene development of basins in the Puna (e.g., Carrapa et al., 2005; Deeken et al., 2006; Jordan and Mpodozis, 2006) that have since undergone surface uplift and are contained within the current plateau. However, the amplitude of basin and range relief within the plateau is much less than that observed along the adjacent plateau flanks and the broken foreland as the result of sediment storage within basins (Figs. 2B and 3). As a result, basin fills in marginal intermontane basins tend to be thinner than those found on the plateau (e.g., Vandervoort et al., 1995; Kraemer et al., 1999; Hilley and Strecker, 2005). In contrast, basins in the broken foreland have large peak-to-basin relief, reflecting erosional mass removal by fluvial systems integrated with the foreland. A topographically controlled decrease in precipitation along the plateau margin reduces the ability of channels traversing the basement ranges of the broken foreland to incise, leading to reduced fluvial connectivity (Hilley and Strecker, 2005; Strecker et al., 2007). This orographic shielding effect becomes more pronounced in the Puna, resulting in hydrologically isolated basins subjected to a progressive decrease in internal relief. Consequently, the low internal plateau relief appears to result from the long-term climatic evolution of the plateau margin, which involves the creation and sustenance of an effective orographic barrier promoting internal drainage and protracted aridity. This analysis thus suggests that the plateau results primarily from geodynamic processes, but the low internal relief reflects basin isolation from the foreland base level as erosional efficiency decreases in the arid plateau. Hence, a synergy of geodynamic processes that increase crustal thickness and/or alter the buoyancy structure of the crust and mantle beneath the plateau and climatically moderated

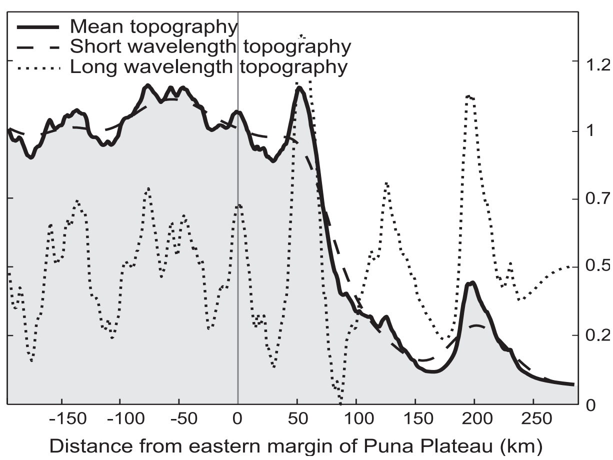

Figure 3. West-east swath (location shown in Fig. 1) showing mean topography (solid line), mean of long-wavelength component of topography (dashed line), and mean of short-wavelength component of topography (dotted line) as functions of distance from eastern Puna margin. Vertical line shows transition from internal to external drainage, which we define as the plateau boundary. Scales for mean observed and long-wavelength component of topography are denoted on left y axis; scale for mean of short-wavelength component of topography is marked on right y axis. Vertical translation of 500 m is applied to short-wavelength topography to make all values positive on graph.

erosional processes that allow the reduction and persistence of internal relief appears to be required to form this Andean plateau landscape.

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