Tectonics and Climate of the Southern Central Andes

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South American Monsoon, orogenic plateaus, orographic barrier, aridification, erosion

Abstract

The history of the southern central Andes, including the world's second largest plateau and adjacent intermontane basins and ranges of the Eastern Cordillera and the northern Sierras Pampeanas of Argentina and Bolivia, impressively documents the effects of tectonics and topography on atmospheric circulation patterns, the development of orographic barriers, and their influence on erosion and landscape evolution at various timescales. Protracted aridity in the orogen interior has facilitated the creation and maintenance of the Puna-Altiplano plateau. Contraction and range uplift, filling of basins, and possibly wholesale uplift of the plateau increased gravitational stresses in the orogen interior, which caused the eastward migration of deformation into the foreland and successive aridification. The uplift of the Andean orogen has also had a far-reaching influence on atmospheric and moisture-transport patterns in South America. This is documented by the onset of humid climate conditions on the eastern side of the Andes in late Miocene time, which was coupled with the establishment of dramatic precipitation gradients perpendicular to the orogen, and changes in tectonic processes in the Andean orogenic wedge.

INTRODUCTION

The topography of mountain belts results from tectonic uplift generated by plate-boundary forces and the efficiency of erosional processes mainly determined by climate and rock type. There are numerous examples in Earth's history in which protracted mountain building caused topography to interfere with atmospheric and oceanic circulation patterns, resulting in climate change, pronounced gradients in precipitation, and spatially variable surface processes (e.g., Koons 1989, Isacks 1992, Horton 1999). Furthermore, field, thermochronologic, and modeling studies indicate that the interaction between topographic construction, deformation, climate, and erosional processes may also control the structural evolution of orogens (e.g., Davis et al. 1983, Willett 1999, Hilley et al. 2004, Whipple & Meade 2004, Thiede et al. 2004; Hilley & Strecker 2004, 2005, Reiners & Brandon 2006). Indeed, these studies suggest that sustained precipitation and erosion may influence the kinematics and locus of tectonic activity in orogens. Therefore, the spatiotemporal changes in deformation and uplift, amount and frequency of precipitation, and the erosional removal of material from an orogen cannot be understood in isolation because feedbacks may exist (e.g., Koons 1989, Masek et al. 1994, Willett 1999, Zeitler et al. 2001, Burbank 2002, Reiners et al. 2003).

The noncollisional Cenozoic Andes (**Figure 1**) exemplify the influence of tectonism on the long-term behavior of climate. With a north-south length of approximately 7000 km, peak elevations in excess of 6 km, marked tectonic activity, and strikingly different climatic regimes across and along the strike of the mountain chain, this orogen lends itself to an analysis of the coupled processes of deformation, climate, and erosion. The morphotectonic provinces of the Andes are perpendicular to moisture-bearing winds that impinge on the eastern and western flanks of the orogen, resulting in pronounced gradients (**Figure 1**). For example, in the southern central Andes, including Bolivia and NW Argentina, the Subandean, Interandean, and Eastern Cordillera (**Figure 2**) ranges block moisture-bearing winds originating in the Amazon basin and the Atlantic, leading to humid eastern flanks and aridity within the Puna-Altiplano Plateau and the Western Cordillera (**Figure 1**). A mirror image of this situation exists farther south at approximately 27°S, where the Southern Hemisphere westerlies cause high rainfall on the western flanks of the Principal Cordillera and the Patagonian Andes and semiarid conditions in the lee of the ranges. The asymmetry in precipitation is also reflected in differences in weathering, erosion, and sediment transport rates on opposite sides of the orogen (e.g., Ziegler et al. 1981, Bangs & Cande 1997, Haselton et al. 2002, Hartley 2003, Blisniuk et al. 2005).

In the central Andes, where moisture transport is from the northeast and east-southeast, the topographic and climatic configuration is closely linked because of the eastward migration of tectonic activity (Strecker et al. 1989, Ramos et al. 2002) that successively starves the leeward western portions of the orogen of moisture (e.g., Kleinert & Strecker 2001, Starck & Anzótegui 2001, Sobel & Strecker 2003, Coutand et al. 2006). Consequently, discharge within channels, and hence incision and landscape-lowering rates in the arid interior of the orogen, are expected to be reduced (e.g., Sobel et al. 2003). In contrast, erosion, and perhaps the locus of tectonic

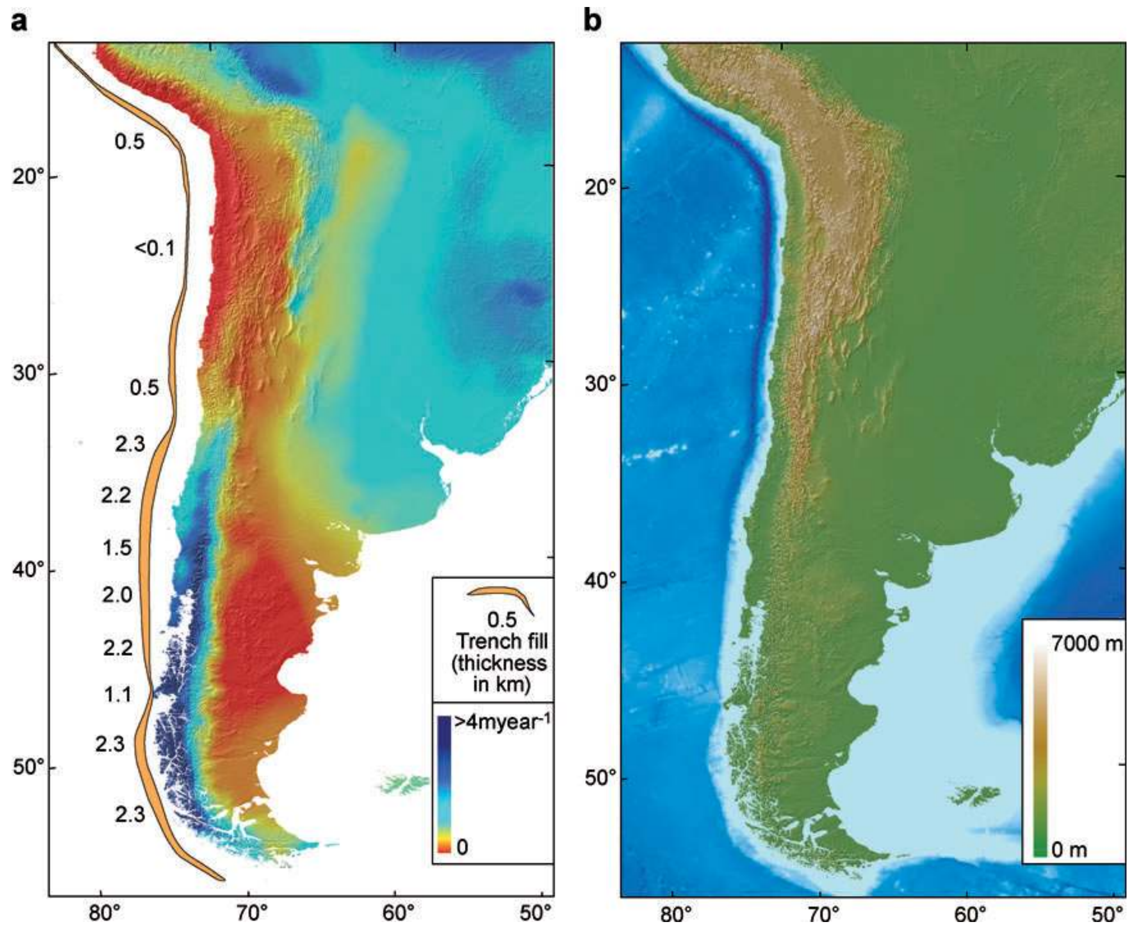


Figure 1

(*a*) Shaded relief map and precipitation in the central and southern Andes and adjacent areas (WMO 1975). Sediment-fill thickness in trench from Bangs & Cande 1997. (*b*) Shaded relief of the central and southern Andes.

activity, owing to removal of material, will likely be strongest where precipitation impinges on the flanks of the orogen or where structurally controlled topographic lows funnel moisture farther into the orogen (Masek et al. 1994, Horton 1999, Sobel et al. 2003, Trauth et al. 2003a, Barnes & Pelletier 2006). In light of the long-term aridity of the western slope of the orogen Lamb & Davis (2003) suggested a positive feedback between protracted aridity and tectonic uplift of the central Andes. In their view, uplift of the orogen is coupled to enhanced aridity due to global cooling and aridity and may result from a high degree of plate coupling with resulting high shear stresses caused by the lack of sediment input into the trench. There are thus several salient climatic and tectonic characteristics of the central Andes that appear to be

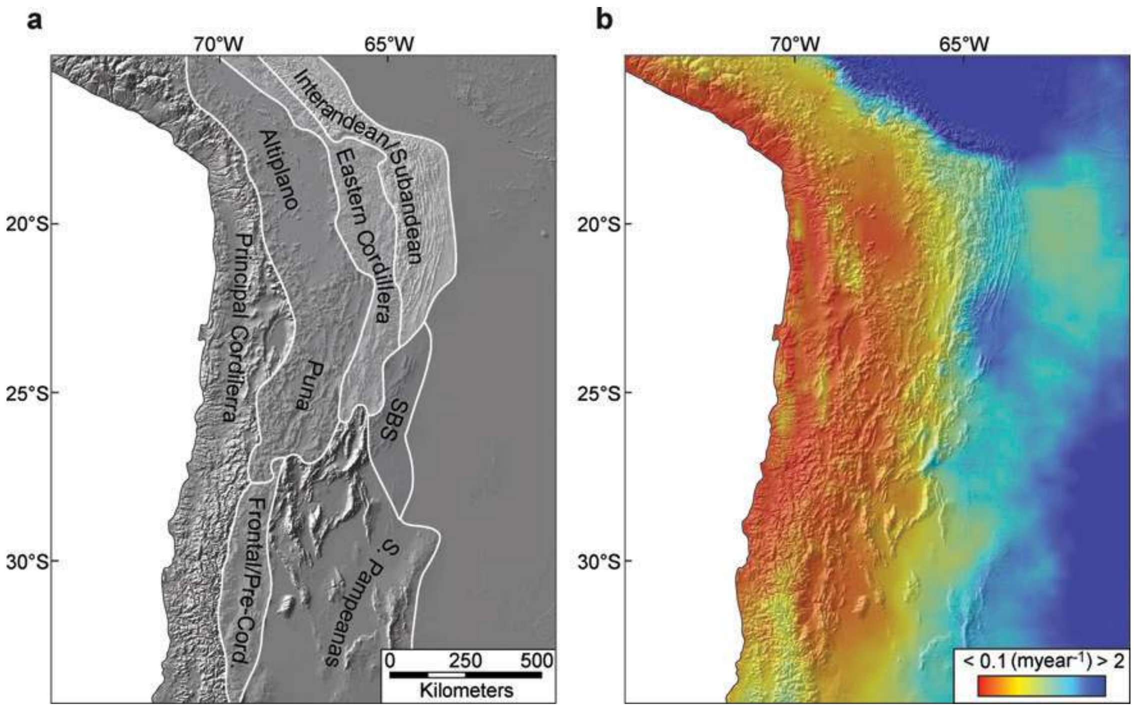


Figure 2

(*a*) Shaded relief map and principal morphotectonic provinces of the southern central Andes (after Jordan et al. 1983). Santa Barbara System, SBS; Sierras Pampeanas, S. Pampeanas; Frontal and Precordillera, Frontal/Pre-Cord. (*b*) Mean annual rainfall distribution derived from the Tropical Rainfall Measurement Mission (TRMM) satellite. Rainfall amounts were calibrated with ground-control stations reported in Bianchi & Yañez (1992) according to methods described in Bookhagen & Burbank (2006).

closely connected and that may have mutually influenced each other during orogenic evolution.

In this review, we examine the relationships between tectonics and climate in the southern central Andes of Argentina between approximately 22°S and 27°S, with particular emphasis on the Puna-Altiplano Plateau and its immediate neighboring morphostructural provinces in Argentina, Chile, and Bolivia. First, we summarize the late Cenozoic geologic and climatic evolution of this sector of the Andes to show that tectonic uplift has had a far-reaching impact on the generation of orographic barriers and the spatiotemporal distribution of precipitation and erosional exhumation. Second, we review the interactions between climate-driven surface processes and tectonism and their influence on developing intraorogenic plateau morphology. Third, we assess the characteristics of the South American Monsoon in relation to the tectonic evolution of the Puna-Altiplano Plateau and the bordering high-elevation ranges to the east. Finally, we evaluate the long-term climate and deformation patterns with respect to potential climatic forcing of tectonism.

Geologic Setting

The principal morphotectonic provinces of the southern central Andes comprise the western slope of the Andes, the active magmatic arc broadly defining the border between Argentina and Chile, the Puna-Altiplano plateau, the Eastern Cordillera (Cordillera Oriental), the Santa Barbara System, the Sierras Pampeanas, and the Subandean foreland fold-and-thrust belt, which borders the Chaco foreland basin (**Figure 2*a***).

With an average elevation of 3700 m, the Puna plateau comprises broad, internally drained depocenters with intervening north-south-oriented mountain ranges, often between 5000 and 6000 m elevation, primarily bounded by high-angle reverse faults (Turner 1972). The basins in the Puna-Altiplano contain continental evaporites and volcanic and clastic deposits typically between 3 and 5 km thick (Jordan & Alonso 1987, Alonso et al. 1991). Eo-Oligocene to Miocene contraction in this region and adjacent areas in the Eastern Cordillera has caused the formation of closed depocenters, now located within the plateau (Kraemer et al. 1999; Coutand et al. 2001; Carrapa et al. 2005, 2006; Deeken et al. 2006; Ege et al. 2007). Structurally similar, although transiently closed, sometimes coalesced intermontane basins in the adjacent morphotectonic provinces of NW Argentina are associated with contraction during late Miocene to Pleistocene time (Strecker et al. 1989, Marrett & Strecker 2000, Reynolds et al. 2000, Bossi et al. 2001, Hilley & Strecker 2005, Coutand et al. 2006, Mortimer et al. in press). The topographically lower and tectonically less compartmentalized Altiplano of Bolivia is also characterized by coalesced sedimentary basins that record a complex history of early Cenozoic extension and thermal subsidence, as well as subsequent widespread shortening and tectonic subsidence (Kennan et al. 1997, Elger et al. 2005). Faulting and folding were associated with bivergent thrust systems mainly active between 33 and 8 Ma (Lamb et al. 1997, Ege 2004, Ege et al. 2007, Elger et al. 2005). The Oligocene to Miocene shortening may have significantly contributed to the building of topography and surface uplift (Gubbels et al. 1993, Kennan et al. 1997, McQuarrie 2002). However, lower crustal and lithospheric delamination may have played a key role in the final uplift of these highlands (Kay et al. 1994, Allmendinger et al. 1997, Garzione et al. 2006), which was coincident with a general waning of shortening within the plateau and the Eastern Cordillera and the onset of thrusting in the Interandean and Subandean regions (Gubbels et al. 1993, Kley 1996, Ege 2004, Elger et al. 2005).

The Eastern Cordillera straddling the margin of the Puna in Argentina consists of late Proterozoic metasedimentary rocks, early Paleozoic sedimentary units, and Paleozoic to Miocene intrusives (e.g., Turner 1972, Jezek et al. 1985) that were uplifted along bivergent, north-northeast striking thrust and reverse faults (Mon & Salfity 1995). The southern part of the Eastern Cordillera is transitional with the eastern ranges constituting the eastern border of the Puna, the northern Sierras Pampeanas, and the Santa Barbara thrust belt, an inverted extensional province related to the Cretaceous Salta Rift basin (Allmendinger et al. 1983, Grier et al. 1991, Mon & Salfity 1995, González & Mon 1996, Kley & Monaldi 2002, Marquillas et al. 2005).

The Sierras Pampeanas occur east of the amagmatic sector of the Andes between 27 and 33°S lat and spatially coincide with the shallow-subducting segment of the oceanic Nazca Plate (Jordan et al. 1983) (**Figure 2**). Structurally, these Laramide-style crystalline basement uplifts (e.g., Jordan & Allmendinger 1986) are akin to the ranges in the transition to the southern Puna and the Cordillera Oriental (e.g., Mon 1979, Strecker et al. 1989, González & Mon 1996, Carrapa et al. 2006). Uplift of the Sierras Pampeanas began in late Miocene time after the last Tertiary transgression into the Andean foreland (e.g., Ramos & Alonso 1995, Ramos et al. 2002, Sobel & Strecker 2003), accelerated after about 4 Ma, and culminated after 3 Ma when intermontane basin deposits were folded and partly overthrust (Strecker et al. 1989, Bossi et al. 2001). To the west, the Sierras Pampeanas are bordered by a thin-skinned foreland fold-and-thrust belt that originated at approximately 20 Ma (e.g., Ramos et al. 2002).

Quaternary Climate and Surface Processes

Being situated in a subtropical high-pressure region with atmospheric subsidence and cold upwelling along the western coast of the continent, the southern central Andes are extremely arid between about 15 and 27°S lat (Houston & Hartley 2003), comparable to the deserts of western Africa and Australia at the same latitudes (**Figures 2** and **3**). This aridity is especially pronounced in the Atacama Desert, which receives about 20 mm/year rainfall (**Figures 2** and **3**) and constitutes a region with one of the lowest erosion rates on Earth that may be lower than 0.2 m/Myr (Nishiizumi et al. 2005, Dunai et al. 2005). With hyperaridity on the western flank, and less than 200 mm/year rainfall on the Puna-Altiplano plateau and in the intermontane basins east of the Puna (Werner 1971, Garleff & Stingl 1983, Bianchi & Yañez 1992), the southern central Andes comprise the most arid sector of the orogen. The primary source of precipitation in this region is associated with Atlantic moisture recycled via the Amazon and moisture related to the South Atlantic Convergence Zone during summer (Garreaud et al. 2003). Minor amounts of precipitation also reach this arid zone via incursions of the Southern Hemisphere westerlies during winter (Vuille 1999). The eastern margin of the Altiplano exhibits a pronounced disparity in the amount of precipitation, runoff, and effectiveness of erosional processes in the northwest versus the more meridionally oriented parts of the Eastern Cordillera and the adjacent morphotectonic provinces (Masek et al. 1994, Barnes & Pelletier 2006). The ranges along the eastern plateau margin in NW Argentina receive 1000 to 3000 mm/year rainfall (**Figures 2** and **3**). These ranges form formidable topographic barriers that shield the interior of the orogen from the moisture-laden easterly winds that impinge on their flanks (**Figures 4** and **5**). This configuration consequently generates an efficient erosional regime (Haselton et al. 2002). The high precipitation on the eastern flanks in the southern central Andes is a direct result of the South American monsoon (Zhou & Lau 1998). The extensive eastern slopes of the Andes are associated with an atmospheric low-pressure system at low levels (Northwestern Argentinean Low) and an upper-air anticyclone (Bolivian High), which develops in summer and attracts moist air from the Amazon lowland (Bianchi & Yañez 1992,

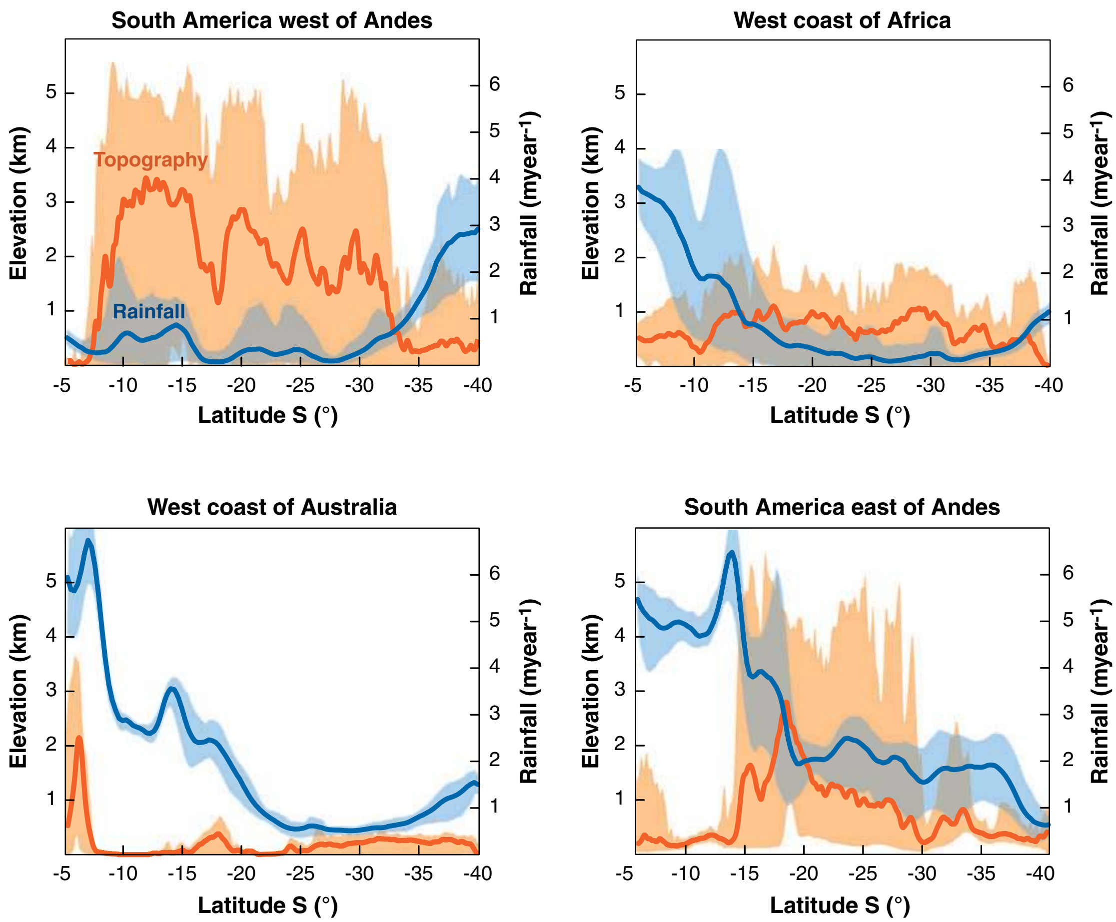


Figure 3

Two hundred and fifty–km-wide N-S swath profiles showing rainfall and topography at similar latitude from the west coasts of South America, Africa, and Australia, as well as the east side of the Andes. Mean value is depicted by the heavy line, whereas shading denotes minimum and maximum rainfall and topography. The diagrams from western South America, Africa, and Australia emphasize the similarity of these coastal deserts. Note the pronounced N-S and E-W asymmetry (**Figure 2**) in the distribution of precipitation on both flanks of the Andes.

Seluchi et al. 2003, Vera et al. 2006). Thus, 80% of the annual precipitation falls in summer from November to February (Rohmeder 1943, Halloy 1982, Bianchi & Yañez 1992, Garreaud et al. 2003) and is transported to these regions by the Andean low-level jet (e.g., Nogués-Paegele & Mo 1997). High summer-monsoon precipitation along the eastern flanks of the Andes in northwestern Argentina is thus not triggered by heating of the plateau region and the development of low pressure conditions, as in the case of the Tibet Plateau and the Indian Summer Monsoon realm (e.g., Prell & Kutzbach 1997, Duan & Wu 2005). In contrast to the Tibetan region,

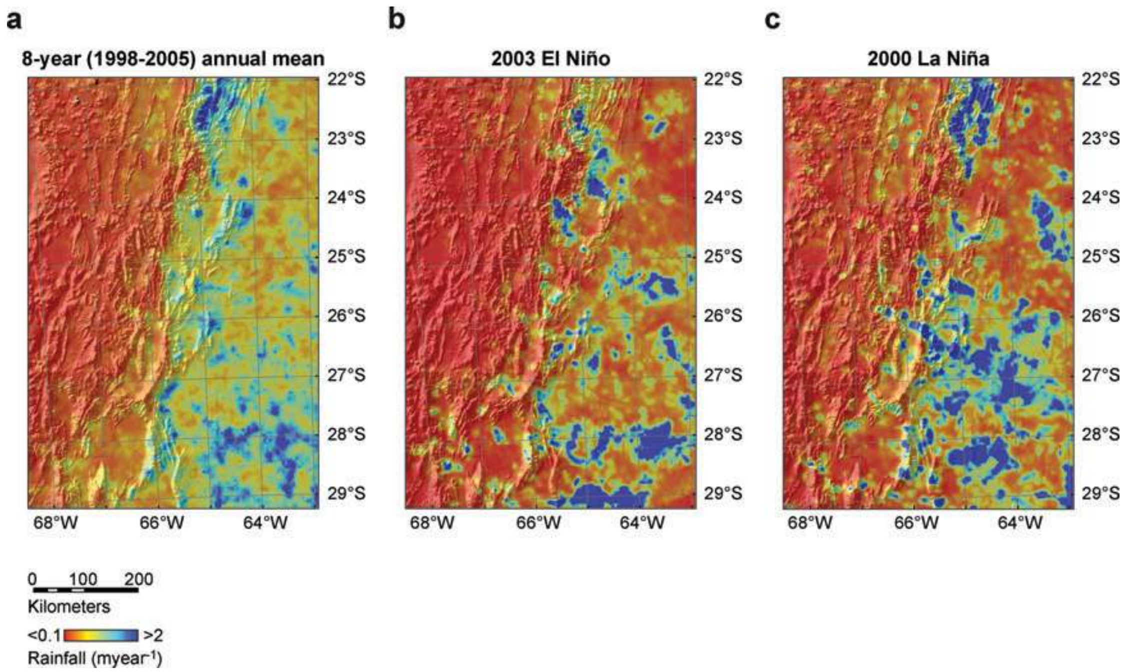


Figure 4

Rainfall amounts in the northwestern Argentine Andes derived from TRMM (Tropical Rainfall Measurement Mission) satellite data (for location see **Figure 2**). Calibration and processing of remotely sensed data are described in Bookhagen & Burbank (2006). (*a*) Mean annual rainfall averaged over 8 years from 1998 to 2005. Note the arid, high-elevation Puna Plateau in the west and the orographic barriers straddling the eastern margin of the plateau; (*b*) mean annual rainfall distribution during the 2003 El Niño. During positive ENSO anomalies, rainfall amounts are reduced in the northwestern Argentine Andes; (*c*) mean annual rainfall distribution during the 2000 La Niña season. Rainfall penetrates farther westward through low-elevation outlets of intermontane basins located at the border of the Puna Plateau. The amount of rainfall in the foreland was atypically high during the year 2000, but penetration of moisture into the drier orogen is characteristic for these episodes and probably protracted phases of higher moisture availability in the past.

the Puna-Altiplano Plateau is apparently not areally extensive enough and it is latitudinally stretched out as a narrow zone that such pronounced heating effects and could be determined by the plateau surface.

The interannual seasonal change in the tropospheric temperature gradient between low and mid-latitudes causes the subtropical westerly jet to extend farther north during winter, reaching its northernmost position around 27°S (Prohaska 1976, Hastenrath 1991). This results in a dry winter climate as regional moisture transport over the eastern flanks of the Andes is prevented (Prohaska 1976, Hastenrath 1991, Bianchi & Yañez 1992). On interannual timescales, these patterns can be significantly modulated by the El Niño Southern Oscillation (ENSO) (Ropelewski & Halpert 1987, Kiladis & Diaz 1989, Bianchi & Yañez 1992, Vuille et al. 2000, Garreaud

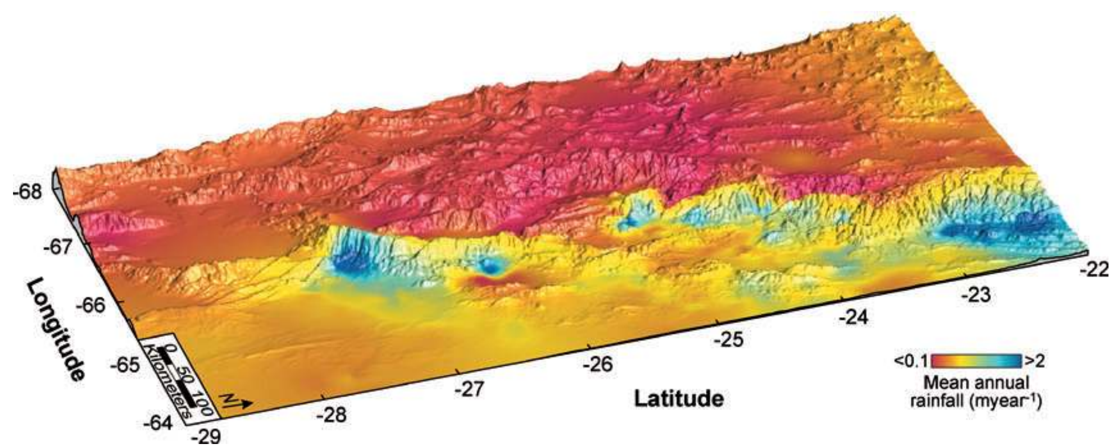


Figure 5

Digital elevation model of the NW Argentine Andes with superposed precipitation patterns. See **Figures 2** and **4** for location.

& Aceituno 2001, Garreaud et al. 2003). For example, in northwestern Argentina and southern Bolivia, strengthened westerly flow with reduced easterly moisture content characterizes El Niño summers, rendering the highlands more arid and the foreland region wetter (**Figure 4**). In contrast, significantly enhanced easterly moisture transport occurs during La Niña summers (Bianchi & Yañez 1992; Trauth et al. 2000, 2003a; Marwan et al. 2003). Moisture transport into the drier parts of the orogen during La Niña episodes may be especially effective in those areas that form natural low-elevation thoroughfares, such as structural transfer zones or outlets of large drainage basins at fault-bounded mountain fronts with diminishing throw (**Figures 4** and **5**).

The southern central Andes have also been subjected to important variations in climate on timescales of 10^3 to 10^5 years Staircase morphologies of multiple lake terraces (e.g., Sylvestre 1999), hydrologic changes reflected in lacustrine sediments (Seltzer et al. 2003, Placzek et al. 2006), plant-fossil data (Latorre et al. 1997, Maldonaldo et al. 2005), and multiple moraine generations (Ammann et al. 2001, Haselton et al. 2002) document recurrent environmental changes. In this respect it is interesting that local late Pleistocene glacial maxima in the Andes of Bolivia and Peru were asynchronous and offset by 10 ka compared with Northern Hemisphere continental glaciation and marine paleotemperature records (e.g., Smith et al. 2005). This emphasizes the pivotal role of moisture and efficient moisture transport into these arid highlands (e.g., Klein et al. 1999), which has also been inferred from hydrologic lake-balance modeling (Bookhagen et al. 2001). There appears to be a broad coincidence between periods of maximum summer insolation on the Puna-Altiplano Plateau and increased availability of moisture, indicating an orbital control of the intensity of the South American Summer Monsoon (Baker et al. 2001, Fritz et al. 2004). However, the spatiotemporal significance of the wetter/drier or cooler/warmer periods varies considerably (e.g., Markgraf & Seltzer 2001, Maldonaldo et al. 2005), and the ultimate cause for pronounced moisture changes may be more complex (e.g., Kull & Grosjean 1998, Betancourt et al. 2000, Rech et al. 2002, Maldonaldo et al. 2005, Cook & Vizy 2006). For example, while there appears to be a good correlation between summer insolation maxima and humid periods on the Altiplano during the last 70 ka (e.g., Bobst et al. 2001), this relationship is more tenuous for earlier episodes of high-lake levels (Fritz et al. 2004). Evaporite and mud-layer alternations, as well as diatom assemblages from sediment cores in the Salar de Uyuni, Bolivia, suggest greater aridity during the penultimate glacial compared to more humid conditions during the last glacial maximum (Fritz et al. 2004). Similarly, efflorescent halite crusts in sediment cores from the Salar de Atacama of Chile document protracted hyperarid conditions for the interval between 325 ka to 53 ka (Lowenstein et al. 2003), whereas an increase in moisture characterizes the past 53 ka (Bobst et al. 2001).

A link between precessional cycles and greater availability of moisture has been inferred from periodic increases in terrigenous detritus in marine deposits at about 27°S during the past 120 ka (Lamy et al. 1998). This was explained by northward shifts of the northern limit of the moisture-bearing westerlies during glacial episodes. However, oxygen isotope data obtained from fluid inclusions in halite from sediment cores in various sedimentary basins in the Puna-Altiplano indicates that increased moisture in this region mainly originated in the tropical Atlantic (Godfrey et al. 2003). This interpretation is supported by a pronounced eastward depression of the Pleistocene snowline in the arid core of the southern central Andes (Haselton et al. 2002). In addition, the absence of west-facing glacial systems north of 27°S document a rather stationary position of the westerlies (Ammann et al. 2001), and modeling studies predict an increase in precipitation along the eastern flanks of the orogen during the last glacial maximum (Cook & Vizy 2006).

Wetter conditions in the intermontane basins along the border of the Puna during the past 50 ka may be responsible for increased lateral fluvial scouring in narrow valleys and an increase in pore pressures that resulted in landsliding in preconditioned rocks prone to failure near faults with Quaternary activity (Fauqué & Strecker 1988; Trauth & Strecker 1999; Trauth et al. 2000, 2003a; Bookhagen et al. 2001; Hermanns & Strecker 1999; Marrett & Strecker 2000; Hermanns et al. 2000, 2001; Fauqué & Tschilinguirian 2002). These conditions were also responsible for maintaining high lake levels in landslide-dammed lakes (Bookhagen et al. 2001) and triggering cut and fill cycles that generated multiple fluvial terraces (Tschilinguirian & Pereyra 2002, Robinson et al. 2005). The landslide deposits (sturzstroms) have volumes $>10^9$ m^3 and cluster at $\sim$35 kyr BP (^{14}C age) and 4.5 kyr BP (calibrated ^{14}C ages), contemporaneous with humid periods in the Puna-Altiplano (Trauth et al. 2003a). The voluminous, low-frequency landslide events in the arid interior of the orogen contrast with the humid eastern flanks, where no such features have been reported. Instead, numerous small rotational slumps and earthflows that are generated annually during the rainy season characterize this region (Hermanns & Strecker 1999). These extreme spatial differences in erosional processes are clearly related to the disparate precipitation, whose impact on erosion processes is manifest at various levels. For example, high precipitation and runoff on the eastern flanks allow effective sediment transport through the fluvial system and export of large volumes of sediment into the foreland

(Masek et al. 1994, Iriondo 1993, Sobel et al. 2003, Hilley & Strecker 2005). This contrasts with sustained internal drainage conditions and ephemeral low-efficiency fluvial systems that may have been initiated as early as Eo-Oligocene time (Schwab 1985, Voss 2002; Adelmann 2001). All drainages in the Altiplano are internal, with the exception of the Río de la Paz and the Río de Consata in Bolivia, which have captured formerly isolated drainages in the Altiplano by headward erosion (Zeilinger et al. 2006). Similar conditions apply to the eastern Puna, where the Río San Juan de Oro has begun to erode into the plateau.

The principal driver for increases in precipitation in the southern central Andes is an intensification of the South American Summer Monsoon, but the mechanistic explanation for this phenomenon is still subject to speculation (Vera et al. 2006). Despite remarkable advances over the past decade, paleoclimate and geochronologic data from this region are still often ambiguous, sometimes contradictory, and limited, especially with respect to timescales exceeding the last ice age. In other orogens, such as the Himalayan-Tibetan region, orbitally controlled radiative processes have been invoked as having caused changes in monsoonal strength, increased moisture, and sediment transport (Fleitmann et al. 2003; Bookhagen et al. 2005, 2006). In contrast, on the Altiplano-Puna plateau amplified insolation obviously did not always result in more moisture because other forcing factors were superposed or modulated the strength of the summer monsoon. This may be associated with intrinsic processes in the Amazon lowland and its moisture source in the tropical Atlantic, perhaps linked with interhemispheric teleconnections (Kull & Grosjean 1998, Rech et al. 2002) a variety of timescales these factors may involve positive feedbacks between precipitation and increased sea-surface temperatures in the Atlantic, possibly related to variations in coastal upwelling (Liebmann et al. 2004); more pronounced meridional sea-surface temperature gradients leading to stronger easterlies and more effective convection and moisture transport (Ruehlemann et al. 1999); or enhanced continental moisture availability due to a stronger Brazil Current (Arz et al. 1998). Alternatively, more humid episodes in the Puna-Altiplano and in the Atacama Desert that were antiphased with insolation may have been related to extrinsic factors associated with increased methane levels in the atmosphere or warming associated with La Niña conditions during enhanced easterlies (Betancourt et al. 2000).

In summary, the eastern flanks of the orogen support an effective moisture and erosional regime, which was maintained or even intensified during episodes of past climatic change (Trauth et al. 2000, Bookhagen et al. 2001, Haselton et al. 2002, Cook & Vizy 2006). In contrast, the western flanks, the plateau region, and intermontane valleys in the lee of the outermost eastern ranges are hyperarid to semiarid, respectively (**Figure 2**). Even small to moderate changes in precipitation and runoff that occur at decadal to millennial time scales lead to significant changes in surface processes in these regions (e.g., Coppus & Imeson 2002, Houston 2005, Amsler et al. 2005). We thus consider these dry environments to be near their geomorphic process thresholds (**Figure 2**). However, variability in the erosional regime has neither been voracious enough to force a general fluvial connectivity with the foreland and the Pacific coast nor sufficiently strong to have triggered major changes in erosional exhumation of these highlands.

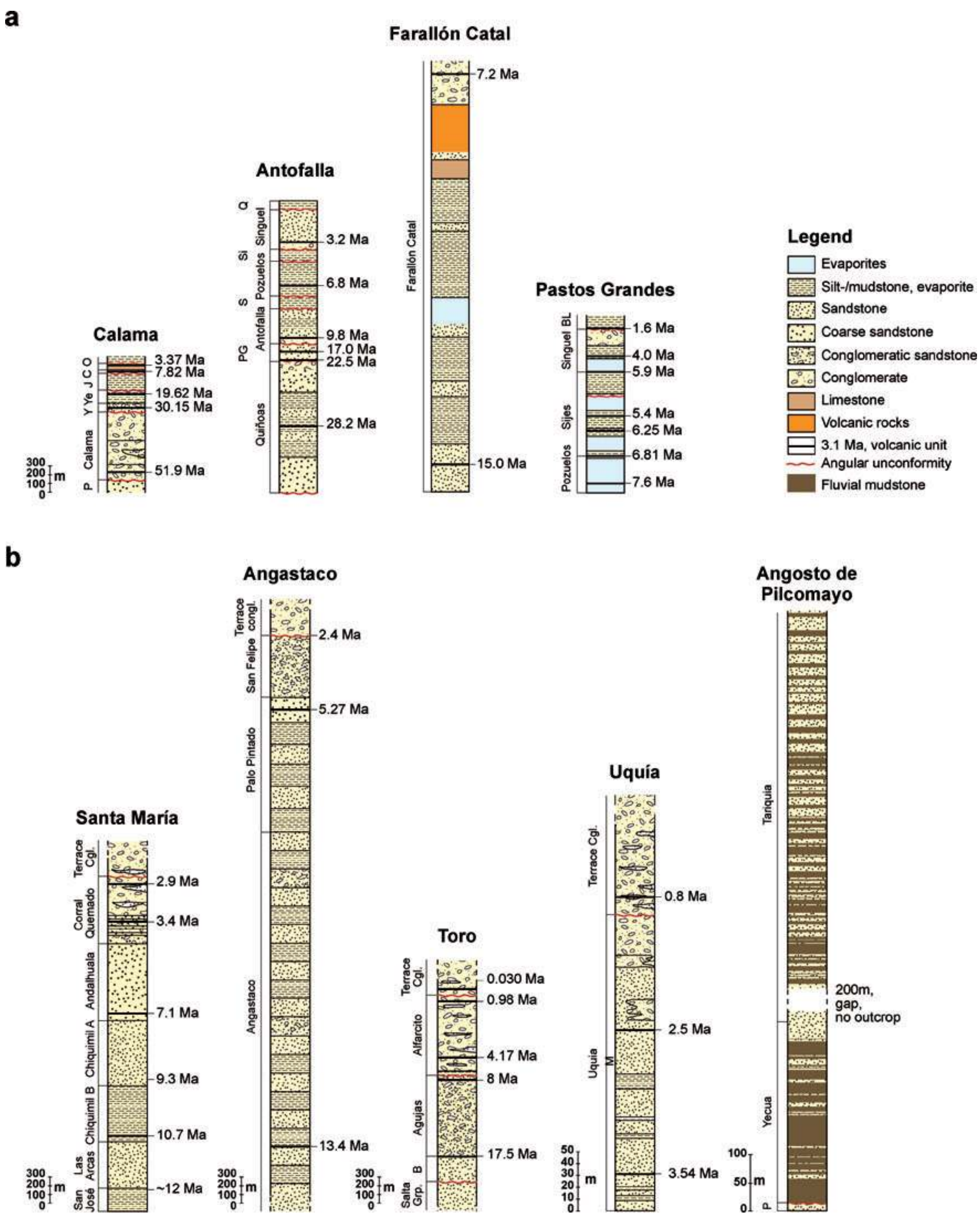
Tertiary Climate Characteristics

According to Hartley & Chong (2002) and Houston & Hartley (2003), the effects of the Hadley circulation, the cold Humboldt current along the South American coast, and the great distance to potential moisture sources in the equatorial region conspire in the Atacama region to maintain one of the driest deserts on Earth. Furthermore, Hartley (2003) suggested that the establishment of the Panama land bridge may have triggered the transition from an arid to a hyperarid climate after 4 Ma. In any case, aridity in the southern central Andes has been a long-lived phenomenon, as the latitudinal position of South America appears to have been relatively stable at its present location during the last 18 Ma (Scotese et al. 1988). In addition, the present atmospheric and oceanic circulation system causing cold upwelling and low evaporation and precipitation on the west coast of South America may have already been established during the Paleogene (e.g., Parrish et al. 1982) owing to the existence of the proto-Humboldt current (e.g., Zachos et al. 2001). Hence, pronounced aridity in this region is to be expected, and in fact may date back to the Mesozoic (Hartley et al. 1992).

In Chile, the Cretaceous Purilactis Group at the western margin of the Salar de Atacama between approximately 22 and 24°S constitutes alluvial fan, sheetflood, dune, and evaporitic playa sediments that were deposited under semiarid conditions (Hartley et al. 1992). Separated by an angular unconformity, these units are onlapped by alluvial fan and evaporative playa units, 57.9 million years old, which are deformed themselves and overlain by Eo-Oligocene alluvial fan deposits (Hartley et al. 1992, Gardeweg et al. 1994) that were deposited east of the Cordillera Domeyko (Mpodozis et al. 2005). This range began exhuming at approximately 64 Ma (Andriessen & Reutter 1994) and experienced rapid exhumation between 50 and 30 Ma (Maksaev & Zentilli 1999), providing sediments that were shed eastward. Conglomerates, approximately 700 m thick, and exposed at the base of a sedimentary sequence in the Calama Basin between 22 and 23°S lat (**Figure 6*a***), were deposited between approximately 52 and 37 Ma and are cemented by gypsum and halite (Blanco et al. 2003). However, there are also sections in this profile that correspond to streamflood deposits, with a matrix characterized by oxidized silts and absence of evaporites. These units are superseded by sand- and siltstones with gypcretes, and recorded aridity with

Figure 6

(*a*) Simplified stratigraphic profiles showing generally arid conditions along the western flank of the Andes at Calama and in the Puna Plateau at Antofalla (Adelmann 2001, Kraemer et al. 1999, Voss 2005), Farallón Catal (Alonso et al. 1991), and Pastos Grandes (Alonso et al. 1991, Strecker 1987). (*b*) Simplified stratigraphic profiles showing the transition from semiarid to humid conditions in the intermontane basins of NW Argentina (Santa María, Angastaco, Toro, Uquía) and in the Subandean Belt of Bolivia (Angosto de Pilcomayo). Santa María (Strecker et al. 1989, Bossi et al. 2001, Kleinert & Strecker 2001), Angastaco (Díaz 1985, Grier et al. 1991, Coutand et al. 2006; M.R. Strecker, unpublished data), Toro (Schwab & Schäfer 1976, Marret & Strecker 2000, Alonso et al. 2006; K. Schwab, oral communication, 1999), Uquía (Reguero 2003; M.R. Strecker, unpublished data), Angosto de Pilcomayo (Uba et al. 2005, 2006). For locations, see **Figure 7**.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Calama
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Silt-/mudstone, evaporite
Sandstone
Coarse sandstone
Conglomeratic sandstone
Conglomerate
Limestone
Volcanic rocks
3.1 Ma, volcanic unit
Angular unconformity
Fluvial mudstone
3.37 Ma
7.82 Ma
19.62 Ma
30.15 Ma
51.9 Ma
3.2 Ma
6.8 Ma
9.8 Ma
17.0 Ma
22.5 Ma
28.2 Ma
7.2 Ma
15.0 Ma
1.6 Ma
4.0 Ma
5.9 Ma
5.4 Ma
6.25 Ma
6.81 Ma
7.6 Ma
Q
Si
Singuel
Pozuelos
S
Antofalla
PG
Quiñoas
Farallón Catal
Singuel BL
Sijes
Pozuelos
Calama
b
Santa María
Angastaco
Toro
Uquía
Angosto de Pilcomayo
Terrace Cgl.
Corral Quemado
Andalhuala
Chiquimil A
Chiquimil B
Las Arcas
San José
2.9 Ma
3.4 Ma
7.1 Ma
9.3 Ma
10.7 Ma
~12 Ma
Terrace congl.
San Felipe
Palo Pintado
Angastaco
2.4 Ma
5.27 Ma
13.4 Ma
Alfarcito
Agujas
Salta Grp.
0.030 Ma
0.98 Ma
4.17 Ma
8 Ma
17.5 Ma
Uquia
0.8 Ma
2.5 Ma
3.54 Ma
Tariquia
Yecua
200m, gap, no outcrop
300
200
100
0
m
50
40
30
20
10

precipitation probably did not exceed 250 mm/year until approximately 6 Ma (May et al. 2005). Toward the top these units are superseded by diatomites and marls, indicating a freshwater lacustrine environment, whereas the basin margins are characterized by calcretes (May et al. 1997, 2005). Interestingly, while the latter deposits suggest greater availability of moisture in late Miocene to early Pliocene time, the opposite could be inferred from data collected in the Lauca and Quillaga-Llamara basins at 18°S and between approximately 21 and 22°S, respectively (Gaupp et al. 1999, Sáez et al. 1999). Kött et al. (1995) and Gaupp et al. (1999) report transient freshwater lake environments in the Lauca Basin related to semihumid conditions in late Miocene time and a drastic switch to arid conditions between 6.4 and 3.7 Ma. This was followed by a return to more variable, semihumid conditions that lasted until 2.6 Ma, after which the present hyperarid conditions were established. These observations emphasize that the environmental signals contained in these basins have to be viewed with care. Tectonically and geomorphologically controlled influences, including changes in the size of watersheds or the connection with high-altitude sectors governed by different climatic conditions, must be taken into account before variations in sedimentary patterns can be used in the context of climate change (Gaupp et al. 1999, Sáez et al. 1999, Allmendinger et al. 2005). Taken together, all available data show that the western flanks of the southern central Andes have had a history of sustained yet variable arid conditions. These were modulated, however, by the effects of locally relevant tectonic or geomorphic controls as well as transient shifts in higher amounts of moisture. The source of these moist air masses reaching the western slopes is not known. Incursions of easterly moisture or changes in sea-surface temperatures facilitating evaporation in the eastern Pacific (Molnar & Cane 2002) may have controlled precipitation in the Atacama region.

The assessment of a variable arid climate is constrained by other data sets. Many workers have referred to the termination of supergene alteration and copper-sulfide enrichment in the Atacama Desert at 24°S and between 14 and 8.7 Ma (Alpers & Brimhall 1988) and in the volcanic arc immediately west of the Puna at 27°47′S at 13.5 Ma (Sillitoe et al. 1991) as evidence for a changeover from semiarid to hyperarid conditions on the western flank of the orogen. Optimal conditions for supergene enrichment exist in precipitation regimes with rainfall >100 mm/year, with a delicate balance between precipitation, the presence of meteoric water circulation available for near-surface chemical weathering, and incision processes (Clark et al. 1967). Sillitoe & McKee (1996) suggested a time window for supergene oxidation and enrichment between 34 and 14 Ma and subsequent hyperaridity. Recent data from the Central Depression of Chile at approximately 25°S lat and at elevations of approximately 1000 m suggest that supergene mineralization took place under semiarid conditions between 33 and 9 Ma and were followed by hyperaridity and a termination of supergene mineralization between 9 and 5 Ma (Arancibia et al. 2006). Furthermore, in their synopsis on supergene enrichment of copper-porphyry deposits in the arid corridor of the Andes between 17 and 27°S lat, Hartley & Rice (2005) present evidence for supergene oxidation and enrichment between 44 and 6 Ma along the western flank of the Andes, whereas the climatic conditions in the Puna-Altiplano region and the Eastern Cordillera have supported conditions for continued mineralization since 11 Ma. The semiarid environments in the Andean highlands apparently have provided environmental conditions conducive to these processes, whereas an increase in aridity on the western flanks after 6 Ma prevented further supergene mineralization (Hartley & Rice 2005). This is in line with an analysis of drainage patterns between 18 and 22°S, suggesting that a transition to hyperaridity may have occurred on the western Andean flanks between approximately 10 and 5.8 Ma (Hoke et al. 2004). These inferences are also corroborated by the termination of the formation of alluvial fans at about 26°S at approximately 9 Ma (Nishiizumi et al. 2005). Rech et al. (2006) suggest that the transition between semiarid and hyperarid conditions began even earlier and may span the interval between 20 and 13 Ma based on a changeover from calcic vertisols to nitrate bearing soils, which only form under extreme hyperaridity. Based on the fact that easterly moisture-bearing winds in the central Andes today precipitate at an elevation of approximately 2000 m, these authors related the pronounced changes in environmental conditions to Andean uplift to a paleoelevation of approximately 2000 m at that time.

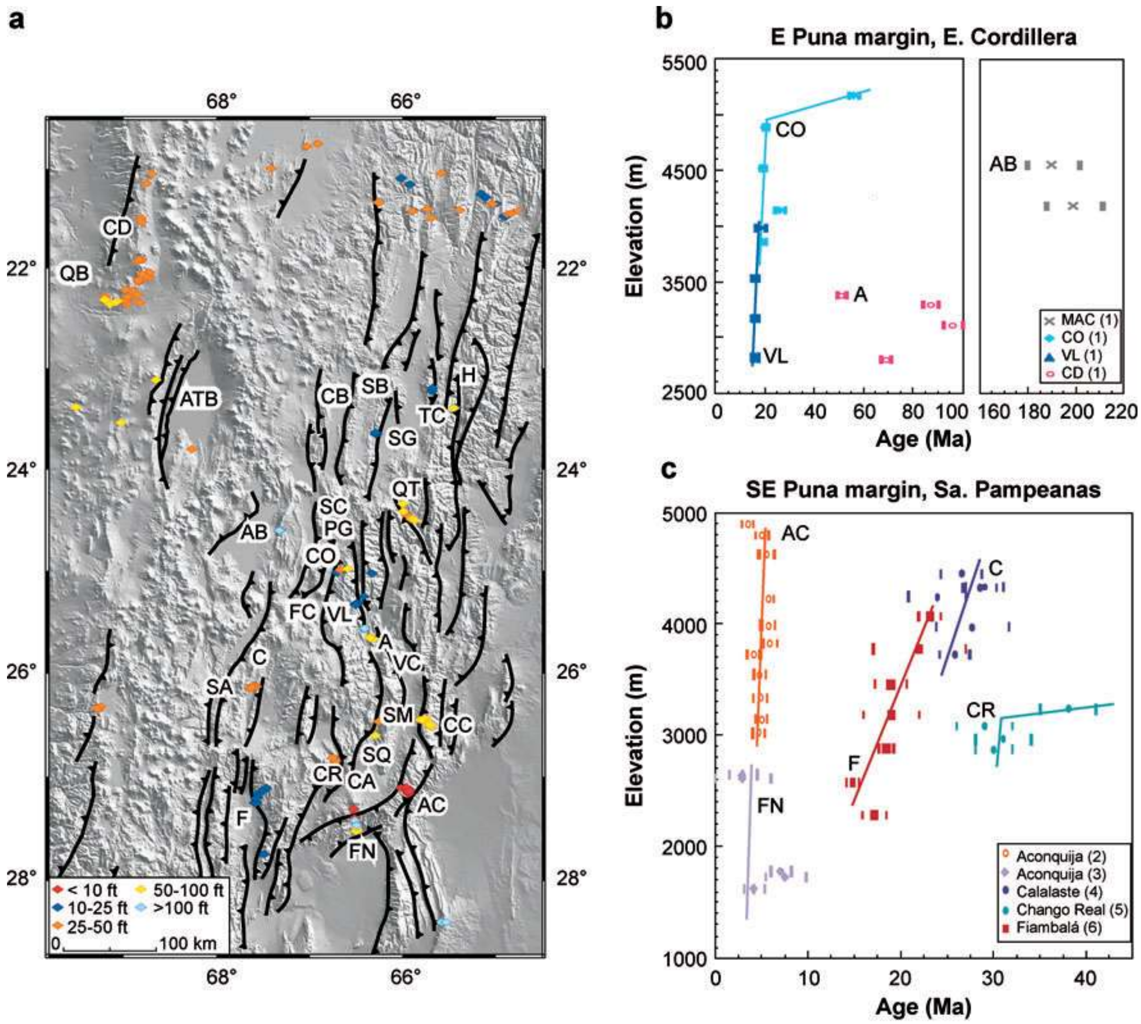
In the Puna, the onset of aridification has been commonly linked with the onset of internal drainage and the deposition of halite and gypsum-bearing units between 24 and 15 Ma (e.g., Alonso et al. 1991, Vandervoort et al. 1995), often several hundreds of meters thick (**Figure 6*b*,*c***). Borate-bearing units in these basins are younger than 7 Ma (Alonso et al. 1992). However, similar to the western Andean flank, semiarid environments may have already existed in the Puna in Eo-Oligocene time, as indicated by eolian deposits and gypsum-bearing playa deposits in the Salar de Antofalla at approximately 26°15′S (**Figure 6*b***) (Adelmann 2001, Voss 2002). In the Salar de Arizaro region, 2000-m-thick Eocene sandstones were deposited in sandflats and dune environments in a foreland basin that had developed east of uplifting ranges on the present western Andean flank (Jordan & Mpodozis 2006). These units are overlain by Oligo-Miocene sand- and siltstones with intercalated layers of gypsum, indicating a semiarid environment (Jordan & Mpodozis 2006). At least 1500 m of sand- and siltstones intercalated with layers of gypsum and halite crop out in the Salar Cauchari area at approximately 24°S may be Eocene in age (Schwab 1973), although other authors consider these units to be Miocene (Alonso et al. 1984). An Eo-Oligocene semiarid environment characterized by transient episodes with increased available moisture has been demonstrated for the playa sandflat, evaporitic playa, and sheetflood deposits in the Peña Colorada and Casa Grande formations of the northern Puna at approximately 22°S lat (Adelmann 2001). Halite and borate-bearing units in the Puna at Farallón Catal and Pastos Grandes at approximately 25°S (**Figure 6**) document that arid conditions have prevailed at least since the middle Miocene (Alonso 1986, Alonso et al. 1991, Vandervoort et al. 1995).

East of the Puna border, sediments in the intermontane Santa María, Calchaquí, Toro, and Humahuaca valleys furnish various climate proxy indicators that document pronounced environmental change during Mio-Pliocene time. Presently, these environments receive approximately 200 mm/year rainfall. In the Santa María Valley (**Figure 6*b***), paleosols in 9 Ma units are characterized by illuvial clays, whereas paleosols in strata formed after 7.1 Ma (Andalhuala Formation) contain authigenic clays, indicating greater availability of moisture. This is in line with palynomorphs and plant fossils that suggest a hot, semihumid environment (Anzótegui 2007) characterized by a mixed vegetation comprised of C4 and C3 plants (Kleinert & Strecker 2001). Faunal assemblages, including giant running birds, ground sloths, and a variety of ungulates (Marshall & Patterson 1981) characterize a subtropical tree savanna (Pascual et al. 1985, Pascual & Ortiz Jaureguizar 1990, Nasif et al. 1997). Overall, the climate after approximately 7 Ma appears to have been more humid than today and with a pronounced precipitation seasonality, similar to the present Chaco region in the undeformed foreland. After 2.5 $\pm$ 0.6 Ma, however, the entire Santa María Group was folded and overthrust by basement rocks of the Aconquija and Calchaquí ranges (Strecker et al. 1989), leading to the establishment of a very effective orographic barrier blocking easterly moisture. From this time onward, semiarid conditions have been sustained in the valley indicated by thick $CaCO_3$-cemented conglomeratic gravels, whereas the windward flanks receive large amounts of moisture (Strecker et al. 1989, Kleinert & Strecker 2001, Haselton et al. 2002).

North of 26°S lat in the Calchaquí Valley (**Figures 6*b*** and **7**), Díaz (1985) and Starck & Anzótegui (2001) describe coarse sandstones of the Angastaco Formation that were deposited in an arid paleoenvironment at approximately 13.4 $\pm$ 0.4 Ma (Grier et al. 1991). This was followed by more humid conditions and the deposition of fluvial sandstones and fluvio-lacustrine siltstones and mudstones rich in organic material, which comprise the Palo Pintado Formation (Starck & Anzóteguí 2001). The Palo Pintado contains a rich mammal fauna, pollen, leaves, and tree remnants that all indicate a significantly more humid climate (Anzótegui 1998, 2007;

Figure 7

Structures and exhumation ages for the southern central Andes. (*a*) Digital elevation map of the southern central Andes based on SRTM data, including important structures, modified from Alonso et al. (2006), Sobel et al. (2003) and references therein, and Mpodozis et al. (2005). Colored dots show apatite fission-track (AFT) cooling ages; data is tabulated in Alonso et al. (2006) and in Carrapa et al. (2006). AFT age distribution reflects the role of aridity in minimizing magnitude of erosion since tectonically driven exhumation occurred. (*b*) AFT age-elevation profiles from the eastern margin of the Puna and the Eastern Cordillera. The oldest profile shows that little exhumation has occurred in the plateau center since the Jurassic. The two young profiles show that rapid exhumation in the Eastern Cordillera commenced around 21 Ma. The intermediate profile represents an exhumed partial annealing zone; thermal modeling suggests that final exhumation commenced between 12 and 7 Ma. Therefore, exhumation has propagated from the plateau center eastwards. Locations of the 4 profiles on (*a*) are denoted by bold abbreviations. Data are all from (1) Deeken at al. (2007); locality names in this reference are Sierra de Macón (MAC), Cerro Durazno (CD), Complejo Oíre (CO), and Valle Luracatao (VL). (*c*) AFT age-elevation profiles from the southeastern margin of the Puna and the Sierra Pampeanas. The Calalaste and Fíambalá profiles were being exhumed between 29 and 24 Ma and 24 Ma, respectively, whereas the Chango Real profile was being exhumed during the Oligocene. The two young profiles show that the northern Sierras Pampeanas were exhumed commencing in the latest Miocene–early Pliocene. Therefore, exhumation commenced within the plateau during the Oligocene and stepped outward much more recently. Locations of the 4 profiles on (*a*) are denoted by bold abbreviations. Data are from (2) Sobel & Strecker (2003), (3) Coughlin et al. (1998), (4) Carrapa et al. (2005), (5) Coutand et al. (2001), and (6) Carrapa et al. (2006).



Locality abbreviations:

A, Angastaco basin
AB, Salar de Arizaro
AC, Sierra Aconquija
ATB, Atacama basin
C, Sierra Calalaste
CA, Valle del Cajón
CB, Cauchari basin
CC, Calchaquí range
CD, Cordillera Domeyko
CO, Complejo Oire
CR, Sierra Chango Real
F, Bolsón de Fiambalá
FC, Farallón Catal
FN, Farallón Negro
H, Quebrada de Humahuaca
PG, Salar de Pastos Grandes
QB, Quillaga basin
QT, Quebrada del Toro
SA, Salar de Antofalla / Antofalla Basin
SB, Susques basin
SC, Siete Curvas
SG, Salinas Grandes basin
SM, Santa María Valley
SQ, Sierra de Quilmes
TC, Tres Cruces basin
VC, Calchaquí Valley
VL, Valle Luracatao

Starck & Anzótegui 2001). For example, the presence of moisture-loving pteridophytes, angiosperms, and floating ferns in these strata indicates that drastically different environmental conditions prevailed here compared to the present and preceding episodes. The identified plant communities characterize a subtropical, hot, and humid environment with minor xeric components compatible with a maximum of 1500 mm/year rainfall, comparable to present-day conditions in northeastern

Argentina and southern Brazil (Cabrera 1976, Starck & Anzótegui 2001, Anzótegui 2007). In contrast, the formation of thick calcretes in the upper part of the superseding San Felipe Formation demonstrate that the humid conditions reverted back to aridity sometime after 5.2 and before 2.4 Ma, which was also associated with range uplifts farther east (Coutand et al. 2006).

In the semiarid intermontane Quebrada del Toro basin (**Figures *6b*** and **7**) at the eastern Puna margin (24°30′S lat), plant-fossil-bearing fluvio-lacustrine silt- and claystones are found that are younger than 8 Ma and older than 4.6 Ma, based on lateral correlations with dated volcanic ash horizons (Marrett & Strecker 2000, Hilley & Strecker 2005, Alonso et al. 2006). The plant fossils include *Typha* sp. *(Thyphaceae)*, *Equisetum* sp. *(Equisetaceae)*, and *Thelypteris* sp. *(Thelypteridaceae)*. This association is identical to recent plant communities along the humid eastern flanks of the mountain ranges in northwestern Argentina that constitute the present-day orographic barriers (*Selva pedemontana* or *Selva de transición*). These plants are found at elevations <1000 m characterized by rainfall between 700 and 1000 mm/year (Cabrera 1976; Brown 1995, Grau 1999). The presence of the Thelypteris fern particularly is important because it is limited to humid environments where waters have an acid pH and trees provide sufficient shade (Collinson 2002, Martínez 2003).

To the northeast, sedimentary rocks in the now semiarid intramontane Quebrada de Humahuaca at approximately 23°30′S lat and 65°20′W long (**Figures 6** and **7**) record a humid environment, probably with some wet-dry seasonality permitting an open forest environment. The presence of *Hydrochoeropsis dasseni*, a capybara-like rodent, suggests the presence of permanent water bodies (Reguero et al. 2003). All fossils occur in the Uquía Formation, >2.78 million years old (Marshall et al. 1982), which conformably overlies gypsum-bearing mud, sandstones, and conglomeratic sandstones of the Maimará Formation. This unit is older than 3.5 Ma (Walther et al. 1998), contains clasts of the eastern Puna margin indicating eastward transport, and is interpreted to have been deposited in a foreland setting. Both units are folded, faulted, and overthrust by basement rocks that constitute the orographic barrier to the east, which is responsible for the aridity in this sector of the Eastern Cordillera. Both units are in turn overlain by a >400-m-thick early Pleistocene calcrete-cemented conglomerate-fill unit that originally covered the erosional paleotopography of this basin.

Farther north, on the eastern slopes of the Subandean region of Bolivia, Uba et al. (2005, 2006) report on paleoclimate indicators in sediments now deformed as part of the foreland fold-and-thrust belt of southern Bolivia. These authors inferred aridity for this environment until approximately 10 Ma, followed by more humid conditions. Similarly, Hernandez et al. (1996) report on the transition from arid to humid environments in this region during the late Miocene. This is in line with the late Miocene begin of megafan deposition (Uba et al. 2006) and the results of carbon isotope studies on Miocene soil carbonates. Theses studies document a moisture stressed C3 vegetative cover inferred to be associated with the onset of monsoon-related precipitation patterns in southern Bolivia between about 12 and 10 Ma (Strecker et al. 2006).

Tectonics, Orographic Barriers, and Strong Climatic Gradients

Paleoaltimetry data helping to constrain the evolution and the impact of orographic barriers that effectively influence climatic conditions is very limited in the Andes (e.g., Gubbels et al. 1993, Hinojosa & Villagrán 1997, Kennan et al. 1997, Gregory-Wodzicki 2000, Blisniuk et al. 2005, Garzione et al. 2006). However, thermochronology data from uplifted ranges combined with sedimentologic data help unravel the paleotopographic development of orographic barriers and associated changes in climate.

Apart from the volcanic arc, the western flank of the Puna also includes the Chilean Cordillera Domeyko (Chilean Precordillera), which extends from 20 to 26°S lat, with peak elevations between 3500 and 5000 m (**Figure 7**). Apatite fission-track data demonstrate that this range experienced a strong pulse of exhumation during the Eocene, when several kilometers of rock were eroded at rates of 0.1 to 0.2 mm/year; during the past 30 Ma, the exhumation rate decreased to approximately 0.05 mm/year (Maksaev & Zentilli 1999). Based on a combination of structural and thermochronologic data, these authors concluded that much of the present relief had formed during the Eocene and that it has been preserved owing to sustained aridity. These early manifestations of deformation and uplift are recorded by the eastward transport of sediments into a former semiarid foreland-basin setting corresponding to the now compartmentalized Puna-Altiplano region (Kraemer et al. 1999, Jordan & Mpodozis 2006). Cosmogenic nuclide and apatite (U-Th)/He dates from the western flank of the northern Andes of Chile (Juéz-Larré et al. 2005, Dunai et al. 2005) indicate minimal erosional modification of the landscape since Oligo-Miocene time and therefore protracted aridity, although other authors suggest that the transition to extremely low surface-process rates associated with hyperaridity took place in Miocene time (e.g., Nishiizumi et al. 2005, Rech et al. 2006). Apatite fission-track cooling ages and sedimentological data from the Sierra de Calalaste in the southern Puna Plateau document the onset of uplift and exhumation of this range during mid to late Oligocene time, which may have initiated the compartmentalization of this sector of the Puna region and caused internal drainage conditions (Voss 2002, Adelmann 2001, Carrapa et al. 2005). Apatite fission-track analysis of rocks from the interior of the Puna plateau east of the Salar de Arizaro at 25°S yields Jurassic ages, also documenting minimal exhumation in this region. This was locally followed by burial of previously exhumed blocks that were finally uplifted during Miocene shortening (Deeken et al. 2007). An Oligocene onset of exhumation associated with shortening and range uplift is reported along the southeastern margin of the Puna between 24.3 and 27°S (Andriessen & Reutter 1994, Coutand et al. 2001, Deeken et al. 2007), which is also recorded in Oligocene detrital apatite fission-track ages from the Angastaco Basin at the eastern Puna margin (Coutand et al. 2006). However, in the Eastern Cordillera between 23 and 25.5°S, apatite fission-track data from a series of vertical profiles in basement rocks document further exhumation and the eastward migration of deformation and uplift between approximately 25 and 15 Ma (Deeken et al. 2007).

To the south, the northwestern Sierras Pampeanas have been instrumental in enhancing the aridification of the orogen. Today, there is a marked decrease in precipitation at elevations in excess of 2000–2500 m on the windward, eastern flank of the ranges (Bianchi & Yañez 1992, Hilley & Strecker 2005). Without other upwind topographic barriers, Sierra Aconquija at approximately 27°S is more than 5000 m above the foreland and constitutes the windward barrier of the semiarid Santa María basin and other intermontane basins (**Figure 7**), which became aridified after 4 Ma (Kleinert & Strecker 2001). Apatite fission-track data documents that rapid exhumation of the western flank of Sierra Aconquija commenced at approximately 6 Ma with a rate of 1 mm/year. The rate slowed significantly at about 3 Ma (Sobel & Strecker 2003), and surface uplift rates have increased from 0.1–0.5 mm/year to approximately 1.1 mm/year, whereas exhumation on the now arid western flank of the range has decreased.

Immediately to the north, the Cumbres Calchaquíes range (**Figure 7**) is structurally identical, but defines a different exhumation setting. Importantly, the sedimentary fill of the adjoining Santa María basin to the west of both ranges and the deformation history of the basin strata suggest that the Aconquija and Calchaquí ranges were uplifted and exhumed at about the same time (Strecker et al. 1989, Sobel & Strecker 2003). However, in contrast to Sierra Aconquija, where precipitation maxima reach 3100 mm/year (Bianchi & Yañez 1992), the Cumbres Calchaquíes range preserves remnants of a regional Cretaceous erosion surface along its crest, similar to the tilted basement block of Sierra de Quilmes to the west (**Figure 7**). This is in marked contrast to Sierra Aconquija, where up to 4300 m of basement rocks were removed from above the highest peaks. The mean exhumation rate over the past 6 Ma for the Cumbres Calchaquíes was between 0.4 and 0.5 mm/year (Sobel & Strecker 2003). The differences in exhumation correlate with significant differences in precipitation (**Figures 4** and **5**). Compared to Sierra Aconquija, the precipitation maximum at Cumbres Calchaquíes is reduced by 50% (Bianchi & Yañez 1992) and located within a broad swath of lower-elevation upwind ranges.

In summary, the earlier uplifts within the Puna and the Eastern Cordillera may have constituted orographic barriers for moisture transport into this region, either from the west or east. The available data from the interior basins documents that similar to the Quaternary, and regardless of transiently increased availability of moisture, semiarid to arid conditions have prevailed in this core region of the orogen since at least the Oligocene. Furthermore, the setting of the easternmost ranges of the orogen underscores that uplifting areas with low precipitation are slowly exhumed, allowing high topography to be rapidly constructed and old basement-erosion surfaces to be preserved. In contrast, along those ranges that are under the direct influence of high, sustained precipitation, erosional exhumation is focused and pronounced, similar to observations in the northwestern and southern sector of the eastern margin of the Bolivian Andes (Gubbels et al. 1993, Horton 1999, Barnes & Pelletier 2006). In both cases, over geologic time, the uplifts have caused the orographic precipitation gradients to become steep, progressively starving the leeward areas of moisture. This extreme asymmetric distribution in precipitation is closely coupled with the precipitation patterns of the South American Monsoon, which provides high amounts of moisture to these parts of the orogen despite their latitudinal position.

Morphology of the Intra-Andean Plateau, Asymmetric Precipitation Patterns, Basin Filling, and Basin Exhumation

The world's orogenic plateaus, including the Puna-Altiplano region, are first-order tectonic and topographic features that are characterized by several unifying similarities. For example, despite different thermomechanical processes responsible for uplifting these regions (e.g., Allmendinger et al. 1997), all of the Cenozoic plateaus fundamentally impact climatic conditions and influence zonal wind and precipitation patterns (Hastenrath 1991). Orogenic plateaus often comprise an amalgamation of internally drained basins that have coalesced over time, hosting thick sedimentary deposits (Meyer et al. 1998, Sobel et al. 2003). Importantly, all Cenozoic plateaus are located in arid to semiarid regions and their flanks constitute efficient orographic barriers, giving rise to pronounced precipitation gradients. High precipitation, runoff, high-density stream networks, and high-erosional capacity are typical for the windward flanking slopes of these regions. Conversely, arid to hyperarid conditions with internal, low-erosional-capacity fluvial networks, and often evaporite deposition in the basin centers characterize plateau interiors and leeward plateau rims (e.g., Alonso et al. 1991, Fielding et al. 1994). As a result, fluvial systems incise inefficiently, especially if these regions coincide with low-erodibility rocks. The low relief typical of these regions may reflect isolation of the local base-level of plateau basins from the low-elevation foreland. During this process, basin aggradation replaces incision and transport, whereas erosion of the surrounding peaks reduces the internal relief of the region (e.g., Sobel et al. 2003).

The disruption of the fluvial system in the present-day Puna region beginning in Eo-Oligocene time and its continuation throughout the Miocene probably triggered a series of interrelated tectonic, climatic, and erosional processes that generated the geomorphic character of the plateau. Paleocurrent indicators, sedimentary provenance, and apatite fission track thermochronology within the plateau region show that flow within this once contiguous foreland basin was disrupted by widespread, diachronous range uplifts that formed the foundation for internal drainage conditions in a semiarid environment (Alonso 1986, Jordan & Alonso 1987, Coira et al. 1993, Marrett 1990, Vandervoort et al. 1995, Kraemer et al. 1999, Carrapa et al. 2005, Jordan & Mpodozis 2006). The spatial coincidence between the onset of topographic construction, generally low exhumation rates, and deposition of evaporites suggests that the widespread range uplifts caused fluvial systems to be disconnected from the foreland. Indeed, the prevailing zonal semiarid climate was not conducive to effective downcutting by permanent fluvial systems that maintained foreland connectivity, and basin isolation eventually ensued (Alonso et al. 1991, Vandervoort et al. 1995, Sobel et al. 2003). This assessment is not only corroborated by thick gypsum, halite, and borate deposits but also by the absence of volcanic clasts sourced in the Puna highlands in sediments in the Toro (<8 Ma; Marrett & Strecker 2000), Angastaco

(<13 Ma; Coutand et al. 2006), and Campo Arenal-El Cajón (10 Ma; Strecker 1987) intermontane basins along the eastern margin of the plateau. Similarly, sediment provenance in the Subandean region of Bolivia is characterized by source regions within the Eastern Cordillera (Uba et al. 2005), and internal drainage of the Altiplano region may date back to the Oligocene (Lamb et al. 1997, Horton et al. 2002).

In contrast to the protracted internal drainage and sediment accumulation in the basins of the Puna-Altiplano, structurally similar basins within the Eastern Cordillera, the Santa Barbara province, and the northernmost Sierras Pampeanas have alternated between internal drainage conditions similar to those of the plateau region and open drainage with a connection to the foreland. These basins currently drain into the foreland, but many of them contain multiple thick conglomeratic fill units that formed in a closed-basin setting or under conditions of reduced transport capacity (Hilley & Strecker 2005). Their ubiquity and similarity along the plateau margin (**Figure 8**) suggests that common processes are responsible for their formation and destruction. Importantly, all of the intermontane basin outlets at the eastern Puna margin coincide with structurally complex parts of the orogen where along-strike changes in displacement allow the fluvial system to either remain or be easily reconnected to the foreland despite the tectonic activity in these areas. In addition, all low-elevation outlets coincide with high-precipitation gradients (**Figures 4** and

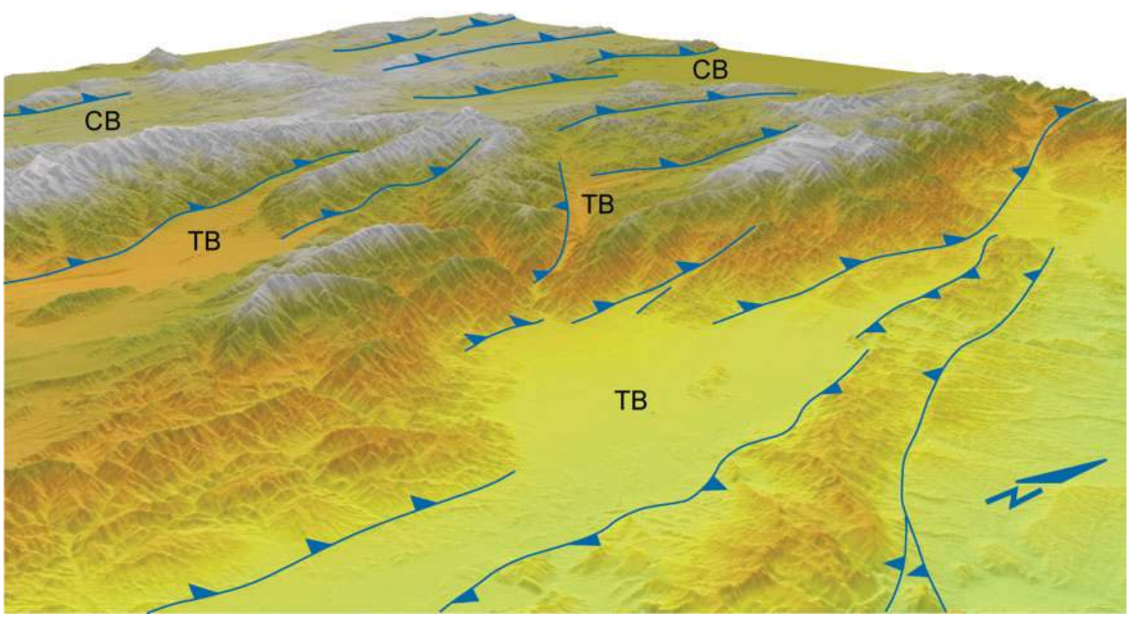


Figure 8

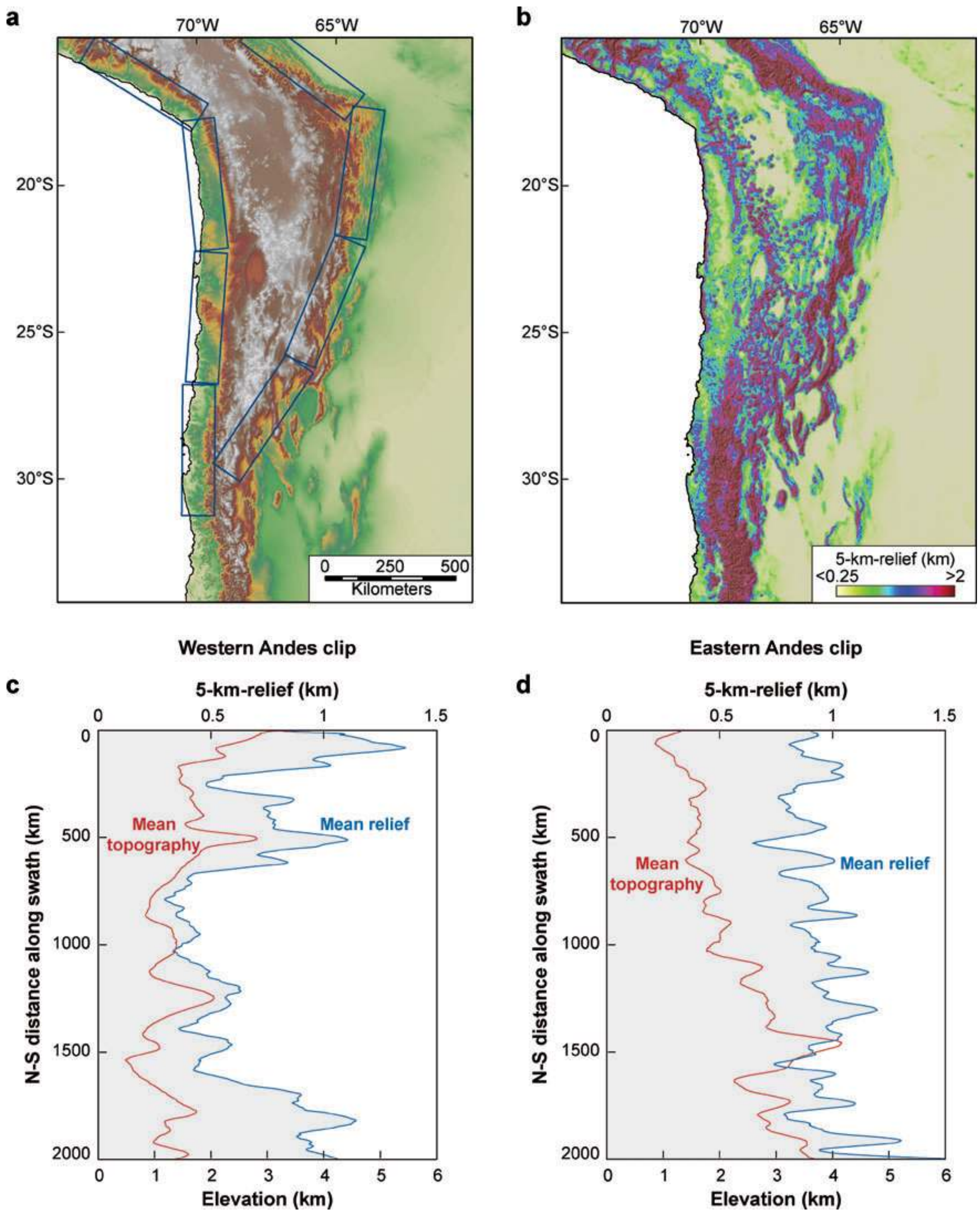
Three-dimensional NW view of closed, fault-bounded sedimentary basins in the Puna (CB) and currently external-draining basins within the eastern border of the Puna (TB). These transiently closed basins at the Puna border were subject to repeated cycles of tectonic activity that transformed the former foreland areas into intermontane basins.

5). During episodes of climatic variability with increased precipitation, these outlets must have been even more effective in funneling moisture into the orogen, helping to maintain external drainage conditions. However, changes in climate, tectonic rates, or the unroofing of resistant units must have repeatedly reduced the transport capacity of the streams or completely halted sediment evacuation, leading to basin isolation and sediment storage. With basin aggradation and reduction of relief contrasts during such episodes, these basins were morphologically part of the Puna realm, and with the exception that no evaporites were found, became virtually indistinguishable from the plateau morphology and the basins there. In some cases, two complete cycles of reduced sediment evacuation or basin isolation following exhumation and reestablishment of external drainage conditions can be observed (Hilley & Strecker 2005). The intermontane basin filling stage is recorded in each of these basins by the arrival of 800- to 1000-m-thick conglomeratic boulders that were deposited over deeply deformed and eroded Tertiary sediments (Penck 1920, Strecker et al. 1989, Tschilinguirian & Pereyra 2001, Alonso et al. 2006). In some cases, lacustrine units are also preserved (Marrett & Strecker 2000, Salfity et al. 2004). The onset of deposition of these conglomerates, locally known as Punaschotter (Penck 1920); the beginning of the erosive cycle that destroyed these fill units; as well as the onset of filling, renewed erosion, and reintegration of the fluvial network in the second cycle are highly diachronous. The largest, and most impressive, intermontane basins of this type include the Humahuaca (<2.7 Ma), Toro (<8 and >4.1 Ma), Calchaquí (<2.4 Ma), Santa María/El Cajón (<2.9 Ma), Corral Quemado (<3.6 Ma), and Fiambalá (<3.6 Ma) basins (Marshall & Patterson 1981, Strecker et al. 1989, Alonso et al. 2006).

Possible Feedbacks Between Erosion and Tectonics in the Central Andes

The southern central Andes comprise a unique setting to evaluate the long-term effects of protracted tectonic activity and climate on landscape evolution. The available data sets suggest that crustal thickness (Kay et al. 1994) and present elevations of this region may have been attained by late Miocene time (Gregory-Wodzicki 2000, Garzione et al. 2006). Similarly, the extreme climatic contrasts between the eastern and western flanks existed at that time (Kleinert & Strecker 2001, Starck & Anzótegui 2001, Montgomery et al. 2001, Houston & Hartley 2003). The asymmetric precipitation patterns are mimicked by the pronounced differences in fluvial and glacial dissection on both flanks of the orogen (**Figure 9**), which is not controlled by major differences in rock erodibility (Haselton et al. 2002).

To compare the erosional evolution of the eastern and western flanks of the Andes, we present a semi-quantitative analysis of the distribution of relief and topography in two 125-km-wide swath profiles (**Figure 9*a***). Relief is a proxy for landscape dissection through erosion, such as fluvial and glacial processes. Generally, relief is constantly high on the eastern flanks of the Andes indicating pronounced fluvial incision and dissected landscapes (**Figure 9*d***). In these presently nonglaciated ranges, episodic increases in moisture during glacial stages has helped shape the erosional relief of


a
70°W
65°W
20°S
25°S
30°S
0 250 500
Kilometers
b
70°W
65°W
20°S
25°S
30°S
5-km-relief (km)
<0.25
>2
Western Andes clip
c
5-km-relief (km)
0 0.5 1 1.5
N-S distance along swath (km)
0 500 1000 1500 2000
Mean topography
Mean relief
Elevation (km)
0 1 2 3 4 5 6
Eastern Andes clip
d
5-km-relief (km)
0 0.5 1 1.5
N-S distance along swath (km)
0 500 1000 1500 2000
Mean topography
Mean relief
Elevation (km)
0 1 2 3 4 5 6

the east-facing slopes (Klein et al. 1999, Haselton et al. 2002), and fluvial incision and denudation rates are high (Sobel & Strecker 2003, Barnes & Pelletier 2006). Interestingly, mean topography in these profiles shows a southward increase in mean elevation (from 1 to ~3.5 km), while mean relief remains roughly the same. This phenomenon may be explained by the incomplete removal of transient Plio-Pleistocene sedimentary fills in the intermontane basins along the Puna. In contrast, relief on the western flanks of the Andes is significantly lower, especially in the Atacama desert region between 20 and 25°S (**Figure 9*c***). On the western flank it is also noteworthy that the transition between a few, deeply incised allochthonous rivers and well-preserved interfluve regions of Miocene age and a more densely dissected landscape with pronounced relief contrasts is located between 27 and 28°S (Haselton et al. 2002). This observation supports the notion that the negligible amounts of sediment in the trench north of 27°S and higher values south of there are a reflection of a relatively stationary position of the westerlies during the past few million years.

Late Miocene aridity on the western flank and in the present plateau regions, as well as the onset of humid conditions along the plateau margins, are compatible with paleoaltimetry studies in the Altiplano of Bolivia and inferences made from apatite fission-track studies in the southern Puna (e.g., Garzione et al. 2006, Carrapa et al. 2006). The onset of humid conditions east of the Puna is in marked contrast with global cooling and concomitant aridity in the late Miocene (e.g., Cerling et al. 1993, Zachos et al. 2001), and it is particularly at odds with the Miocene development toward hyperarid conditions along the western flanks (Rech et al. 2006, Strecker et al. 2006). Therefore, this climatic transition points to a tectonic, not a global-change, origin. This change was associated with the establishment of effective orographic barriers that blocked moisture-bearing winds, which were forced to precipitate on the windward side of the developing central Andean plateau and its flanking ranges. At that time, the Puna-Altiplano region and the Eastern Cordillera may have had the necessary elevation to modify the South American paleo-monsoon by causing a seasonally changing low-level inflow of moisture into regions that would have been otherwise characterized by aridity. This agrees with climate modeling by Lenters & Cook (1997), suggesting that the present-day distribution of rainfall requires a minimum elevation of 2000 m in the Andes. This assessment concerning the interception of moisture-laden easterly winds and their southward deflection along the southern central Andes is also compatible with the position of the Northwestern Argentinean Low (NAL)

Figure 9

Topographic and relief distribution for the southern central Andes. Relief is based on the difference between minimum and maximum topography in a 5-km radius. Note the constantly high amounts of relief on the eastern side of the Andes that is contrasted by very low relief on the western Andes side. To quantify this pronounced contrast, we analyze mean topography and relief from a 125-km-wide swath from the eastern and western side of the Andes. Mean relief is on the order of ~0.9 km in the eastern side, while it drops to less than 0.5 km on the western side of the Andes. The contrast in relief on both flanks of the orogen is even more pronounced, if the punctiform constructional relief of the volcanic edifices in the magmatic arc were removed.

along the eastern slopes of the orogen (Seluchi et al. 2004). The NAL attracts these air masses and furthermore causes their clockwise transport into the intermontane basins of the northernmost Sierras Pampeanas and the southern Puna (Vera et al. 2006b), underscoring the role of tectonically created topography in modifying rainfall patterns.

The history of the southern central Andes not only illustrates the impressive effects of tectonism regarding orographic barriers, precipitation patterns, and the modification of zonal climate characteristics it also demonstrates how precipitation may modulate the tectonic evolution of an orogen. Sustained aridity (e.g., Hartley & Chong 2002, Hartley 2003) in combination with tectonic and a specific set of geomorphic processes have conspired to create and maintain the Puna-Altiplano Plateau. Under these circumstances it is thus very difficult to verify the climate-driven uplift model of Lamb & Davis (2003) for this region, which links uplift of the orogen to closer plate coupling owing to low sediment input into the trench due to aridification in the course of major glaciations in Antarctica.

The uplift of the Andean plateau region and its adjacent ranges has caused a highly asymmetric E-W distribution of precipitation associated with the westward transport of moist air masses in the South American monsoonal realm. This a setting may ultimately result in the demise of the plateau morphotectonic province, if the voracity of headward erosion on the eastern slopes is maintained. To determine the conditions that create and maintain internal drainage and isolation, Sobel et al. (2003) and Hilley & Strecker (2005) analyzed the conditions causing hydrologic isolation versus basin connectivity for the Puna Plateau and adjacent intermontane basins by determining the roles of rock uplift, precipitation, and rock type. Channel cut-off from the foreland and ensuing basin isolation were determined by a threshold that depends on the ratio of uplift (U) in the developing bedrock barrier to the square root of the product of the fluvial transport constant (D_c), which encapsulates the climatic conditions and the bedrock erodibility constant (K). Within tectonically uplifting basement blocks, a fluvial system will incise a bedrock barrier through knickpoint migration and aggrade areas upstream from the developing constriction to adjust to the tectonic disturbance of the longitudinal profile. Thus, if the rock-uplift rate increases relative to the capability of the stream to effectively remove material and incise into bedrock, internal drainage will be established. Furthermore, if the building of topography is coeval with establishing steep orographic precipitation gradients, reduced amounts of moisture will reach the leeward interior basins and diminish discharge. Aggradation and uplift may therefore outpace incision, isolating the basin from the foreland. Therefore, if it is true that the present-day Puna-Altiplano plateau became uplifted in the course of mantle delamination during late Miocene time (Allmendinger et al. 1997, Garzione et al. 2006), it had already inherited all structural and morphologic elements that characterize this region now. The sustenance of long-term aridity and the occurrence of widely distributed range uplifts that disrupted the drainage caused a fluvial system virtually disengaged from its foreland regions. In addition, the exposure of low-erodibility rocks and the inability of rivers to remove material led to the accumulation of thick sediment fills within the intermontane basins, which helped reduce relief contrasts and the formation of coalesced basins (**Figure 8**). With the exception of the headward erosion of the La Paz, Catana, and Juán de Oro rivers into the plateau realm, the Puna-Altiplano has maintained its character as an internally drained tectono-geomorphic entity for at least the past 10 Ma owing to the blocking of easterly moisture.

In the Puna-Altiplano, the failure of sediment evacuation, protracted crustal shortening and thickening of the crust within the plateau, as well as plateau uplift must have fundamentally altered the distribution of crustal stresses in the orogen. Royden (1996) and Willett (1999) demonstrated that such conditions ultimately lead to an increase in the lithostatic load. Despite ongoing overall shortening in the Andean orogen, the gravitational stresses related to the combined effects of shortening, basin filling, and surface uplift may eventually inhibit contractional tectonic activity in the plateau, initiate normal faulting, and force deformation toward lower elevations in the foreland (e.g., Dalmayrac & Molnar 1981). In the Bolivian Andes the inferred late Miocene uplift of the Altiplano was linked with an eastward migration of tectonic activity documented in thermochronologic, structural, and sedimentologic data (Kley 1996, McQuarrie 2002, Ege 2004, Horton et al. 2002). In the Puna region, shortening became predominately focused on the plateau margin and foreland regions at that time (Strecker et al. 1989, Coutand et al. 2006), thus widening the orogen and causing the ongoing uplift of basement blocks. It is important to note, however, that the width of the Puna foreland broadly correlates with the Cretaceous Salta Rift (e.g., Allmendinger et al. 1983). Thus, the pronounced width of the southern sector of the investigated area is not only a result of the protracted aridity (e.g., Montgomery et al. 2001) and the evolution of deformation patterns associated with high topography in the core of the orogen (e.g., Sobel et al. 2003) but also to structural inheritance. In contrast to the margins of the plateau, Pliocene and Quaternary shortening has been negligible or nonexistent on the plateau and was replaced by extension (Allmendinger et al. 1989, Marrett et al. 1994, Cladhous et al. 1994, Elger et al. 2005).

In contrast to the fortuitous combination of conditions constructing a plateau in this arid environment, a set of climate-related conditions can also be envisioned to prevent plateau formation or to cause plateau destruction. This may include extremely large drainage-basin areas to provide runoff and erosional power, the filling of basins to the spill point to reintegrate drainage and to evacuate basin fills, or an increase in precipitation and runoff to promote headward erosion and basin capture. Consequently, if recapture of such areas should occur and material were rapidly evacuated, deformation could migrate back into the interior of the orogen as sediment loads decrease. Such processes may have occurred in the intermontane basins along the eastern Puna margin owing to their vicinity to strong precipitation gradients and the ability of moisture to penetrate into the orogen at low-elevation outlets (Sobel et al. 2003, Hilley & Strecker 2005). The diachronous faulting, filling, and incision in the intermontane basins adjacent to the Puna may reflect these processes. However, more tight timing constraints are needed to document an unambiguous causative relationship between phases of basin filling, variations in lithostatic loading, and tectonic activity along the high-angle fault-bounded basement ranges in this environment that do not constitute an integral part of an orogenic wedge as in fold and thrust belts (Mortimer et al. 2007), where such relations are known to exist (e.g., Dahlen & Suppe 1988, Hilley et al. 2004).

Observations from the Andean orogenic wedge in Bolivia document the climatic forcing of tectonic processes that cause shortening to be accommodated along low-angle structures (Mugnier et al. 1997, Horton 1999, Leturmy et al. 2000). Here, in the northern Sierras Subandinas, the orogenic wedge is narrow, out-of-sequence deformation is common, and precipitation and erosional processes are efficient. In contrast, in the southern Sierras Subandinas between 20 and 24°S lat, precipitation is less pronounced, the fold-and-thrust belt is wider, regional erosion surfaces are preserved, and deformation has successively migrated eastward. These differences are also brought out by different depths of the underlying detachment faults. Therefore, unlike the Puna where internal drainage and tectonism have generated a threshold beyond which erosional mass export is stopped, the externally drained Sierras Subandinas fold-and-thrust belt is being eroded, and deformation and geometry may change spatially and temporally as topography is being constructed (Horton 1999, Barnes & Pelletier 2006).

In conclusion, the inherent aridity in the southern central Andes has favored the development and maintenance of the second largest plateau on Earth. Aridity in the orogen is primarily a function of latitudinal position and cold upwelling along the Chilean coast, but it has been exacerbated by the construction of orographic barriers during the eastward-migrating deformation front that prevents westward moisture transport and successively starves the interior of the orogen of moisture. This provided the conditions for low erosional denudation rates in the interior, basin isolation, and the generation of thick sedimentary fills that may have aided the outward propagation of basement uplifts into the foreland regions. The plateau uplift may be ultimately responsible for the onset of a humid climate along the eastern border of the orogen, which provides high precipitation and runoff to maintain an efficient erosional regime, which in turn profoundly influences the tectonic and landscape evolution of the mountain belt at various timescales.

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