Links between topography, erosion, rheological heterogeneity, and deformation in contractual settings: Insights from the central Andes

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ABSTRACT

Orogenic structure in the central Andes (15°S–34°S) systematically varies with mean annual precipitation, suggesting that erosional processes may be coupled to tectonic processes. We explore the range of possible interactions between deformation, erosional processes, changing geodynamic conditions, and preexisting geologic structures to assess the relative importance of each. Our review and synthesis indicates that the pre-orogenic geologic history and changing geodynamic conditions leave a lasting imprint on the modern structure of the mountain belt. In contrast, changes in precipitation that result from regional-scale atmospheric and oceanic circulation interact with local topography to control the climatic zoning across the central Andes. Changes in erosion, which likely correspond to changes in climate, rock uplift rate, and the exposure of different lithologies, may in turn affect the tectonic development of the region at the scale of each morphotectonic province. In this view, geologic and geodynamic factors set the stage for the nature and strength of coupling between erosional processes, tectonic deformation, and mountain belt structure in the different parts of the central Andes.

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1. Introduction

Over the last three decades, earth scientists have realized that processes that redistribute mass across Earth’s surface may influence deep earth processes (e.g., Davis et al., 1983; Beaumont et al., 1992; Willett et al., 1993; Willett, 1999). For example, global and regional climate changes may enhance or subdue erosion within mountain belts, and these changes result in mass redistribution patterns that may affect the near-surface thermal state of the crust (e.g., Ehlers, 2005), the body forces within the crust (e.g., Dahlen, 1984), and cause changes in mean elevations that result in isostatic adjustments (e.g., Molnar and England, 1990a). Many of the concepts linking surface processes to those deep within the earth arose from observations from Taiwan (e.g., Davis et al., 1983; Dahlen and Suppe, 1988), the Himalayas (e.g., Hodges, 2000; Thiede et al., 2005; Grujic et al., 2006), and geodynamic models (e.g., Beaumont et al., 2004; Stolar et al., 2007). In the central Andes, tectonic deformation and erosional patterns also correlate with one another, suggesting that here too they may be linked (e.g., Horton, 1999; Montgomery et al., 2001).

The South American convergent plate margin mostly consists of the subduction of the Nazca Plate beneath South America along an eastward-dipping subduction zone (Stauder, 1975; Barazangi and Isacks, 1976, 1979; Hasegawa and Sacks, 1981; Bevis and Isacks, 1984; Cahill and Isacks, 1992; Allmendinger et al., 1997). Volcanic and tectonic processes have deformed and uplifted this margin, which hosts a variety of landscapes, including volcanoes in the Western Cordillera, internally drained basins in the high-elevation Altiplano–Puna Plateau, thin- and thick-skinned fold-belts within the Eastern Cordillera and Subandean ranges, and basin-and-range-like topography created by basement-cored uplifts of the Sierras Pampeanas (Fig. 1). In the central Andes, changes in topography and deformation style have been attributed to a variety of tectonic factors, including the geometry of the down-going subducting plate (Isacks, 1988; Gephart, 1994), the thickness of the mantle lithosphere in the central segment of the mountain belt (e.g., Kay et al., 1994; Whitman et al., 1996; Garzione et al., 2006), interplate coupling (e.g., Gephart, 1994; Lamb and Davis, 2003), and pre-existing heterogeneities in the crust and mantle that reflect the pre-Cenozoic geologic history of the area (e.g., Allmendinger et al., 1983; Coutand et al., 2001; Ramos et al., 2002). However, changes in erosion related to climate, and the exposure and erosion of different rock types have been inferred to play a role in determining the patterns of deformation and topography observed within this orogen (e.g., Masek et al., 1994). For example, the intensity of erosion that likely varies...
because of climate patterns related to long-term global-scale atmospheric circulation (e.g., Haselton et al., 2002; Parrish et al., 1982, Hartley et al., 1992; Masek et al., 1994; Strecker et al., 2007), may affect mountain-belt width (e.g., Willett, 1999; Montgomery et al., 2001) or control the interplate frictional coupling by modulating the sediment supply to the subduction trench (e.g., Lamb and Davis, 2003). At a regional scale, studies have argued that erosion plays a key role in controlling the kinematics, geometry and pattern of deformation within the subandean fold-and-thrust belt (e.g., Masek et al., 1994; Horton, 1999), the Argentine Precordillera (Hilley et al., 2004), and the Sierras Pampeanas (Sobel and Strecker, 2003; Hilley et al., 2005).

While erosion may play a role in shaping the central Andes, the first-order characteristics of the topography and deformation of this orogen correlate strongly with changes in geodynamic features such as down-going slab geometry (e.g., Isacks, 1988; Cahill and Isacks, 1992; Gephart, 1994; Allmendinger et al., 1997), and geologic heterogeneities in the crust that include inherited structures and paleogeographic features (e.g., Schmidt et al., 1995; Allmendinger et al., 1983, 1997; Ramos et al., 2002; Carlier et al., 2005). Based on expectations derived from the theoretical framework that links erosion to tectonic processes, the strength of the linkages between these factors should vary spatially depending on the rheology of the crust and its heterogeneities, the lithologies exposed at the surface, and regional to local climatic conditions. In this contribution, we review geological and geophysical observations, as well as modeling studies to examine the strength and nature of coupling between tectonic and erosional processes in the central Andes. The spirit of this work follows that of previous geological studies, in which correlations between natural phenomena are documented and interpreted in the context of experimental (e.g., Persson et al., 2004; Hoth et al., 2006) and theoretical (e.g., Dahlen, 1984) models that rationalize linkages between the observed phenomena. Unfortunately, one cannot manipulate the underlying tectonic processes or the geological history of natural orogens as is the case in a controlled laboratory environment (e.g., Persson et al., 2004; Hoth et al., 2006), and so while not wholly satisfying, we follow the path of others in which we seek correlations between various parameters that can be understood in terms of these simpler systems (e.g., Hoffmann-Rothe et al., 2006).

2. Conceptual framework linking tectonic and erosional processes

In the simplest conception, the state of stress within tectonic plates results from the sum of tractions at the plate boundaries due to plate-tectonic interactions, the buoyancy stresses applied to the base of
the crust that arise from density differences between the crust and mantle, and stresses produced by the overburden of crustal material. The rheology of the crust relates deformation and/or deformation rate to these loading conditions (Fig. 2). The rheology of rocks depends on many factors including composition, temperature, loading stresses, and grain size (e.g., Bürgmann and Dresen, 2008, and references therein), and so deformation may directly modify crustal rheology through at least three mechanisms: 1) heat redistribution in the crust due to magmatism, exhumation, and frictional failure in the upper crust, (e.g., Byerlee, 1968; Byerlee, 1978; Brace and Kohlstedt, 1980; Graham and England, 1976; Lachenbruch and Sass, 1980; Ryback and Cermak, 1982; Edman and Surdam, 1984; Molnar and England, 1990b; Furlong et al., 1991; Royden et al., 1997; Beaumont et al., 2004; Byerlee, 1978) Lachenbruch and Sass, 1980; Ryback and Cermak, 1982; 2) grain size reduction by dynamic recrystallization (e.g., Rybacki and Cermak, 1982; Furlong et al., 1991; Royden et al., 1997; Beaumont et al., 2004; Byerlee, 1978) Lachenbruch and Sass, 1980; Ryback and Cermak, 1982; 2) grain size reduction by dynamic recrystallization (e.g., Rybacki and Cermak, 1982; Edman and Surdam, 1984; Molnar and England, 1990b); and 3) spatial redistribution of rheologically distinct crustal units during deformation (e.g., Allmendinger et al., 1983).

Changes within the deep earth also influence the rheology of rock at shallower levels; for example, changes in mantle temperatures may eventually be manifest by temperature changes within the crust (e.g., Sleep, 2005, and references therein; Fig. 2). Convective removal of mantle lithosphere or magmatic underplating may alter the crust and mantle lithosphere’s thicknesses and consequently, the thermal state of the crust (e.g., Kay et al., 1994; Conrad and Molnar, 1997). Finally, the rheological configuration of the crust prior to orogenesis potentially leaves a lasting imprint on the subsequent evolution of the deforming crust. Thus, in active convergent orogens, the distribution of deformation can be spatially non-uniform and evolve over time as the state of stress and rheology of the crust change in space and time.

Finally, erosion of Earth’s surface can play a fundamental role in deformation by modifying the distribution of lithostatic stresses in the crust (Fig. 2). For example, as surface elevations increase, so do the lithostatic stresses within the crust, and this may favor failure along the extremities of an orogen at the expense of internal shortening (e.g., Dahlen, 1984; Hilley et al., 2005) and may elicit an isostatic response where mean elevations change (Molnar and England, 1990a). By implication, the state of stress and deformation within Earth’s crust is, at least in part, controlled by erosional processes. The strength of these erosional processes is likely moderated by factors including climate (e.g., Knox, 1983; Bookhagen et al., 2005a, b; Strecker et al., 2007) and the lithology of the rocks exposed at the surface (e.g., Gilbert, 1877; Hack, 1960; Stock and Montgomery, 1999; Sklar and Dietrich, 2004). In turn, local climate is strongly influenced by topographic relief that creates orographic gradients in precipitation (e.g., Barry, 1981; Bookhagen and Burbank, 2006), and global climate is perturbed by the uplift of large portions of Earth’s surface such as the Tibetan Plateau (e.g., Raymo and Ruddiman, 1992; Ruddiman, 1997). Thus, surface uplift, climate, erosion, and deformation are inextricably linked in actively deforming orogens.

The lithology of rocks exposed at the surface partially controls erosional efficiency and topographic slopes. Consequently, as orogens evolve, changing distribution of lithologies likely changes the way in which deformation is accommodated. For example, as contractional deformation propagates toward the foreland, the exposure of poorly consolidated basin sediments may enhance erosion, which allows large quantities of material to be eroded even where relief is subtle, and delays the accumulation of lithostatic stresses and advancement of deformation to the orogen’s external portions (e.g., Hilley et al., 2004). Similarly, orogens in which surface processes are veracious due to abundant precipitation may exhum deeper crystalline rocks whose resistance to erosion is high. The surface exposure of such competent rocks promotes the rapid steepening of orogenic slopes, increases lithostatic loads within the orogen’s interior, and forces deformation to migrate towards the external portions of the mountain belt (e.g., Hilley et al., 2004; Hilley and Strecke, 2004). In addition, where erosion is inefficient during the waxing stages of orogenesis, the balance between lithostatic stresses at the base of the crust and the restoring stresses that buoyancy provides may be altered, causing the thickened crust to partially sink into the mantle (e.g., Stuewe, 2002). As orogenesis wanes, lowering of the surface reduces the lithostatic stresses that oppose buoyancy stresses, causing isostatic uplift of rocks (Molnar and England, 1990a). Thus, erosional removal of mass from Earth’s surface can modify the local lithostatic stresses acting along individual structures and within the frictional medium of the crust, as well as the balance between lithostatic loading and the restoring stresses provided by the relative buoyancy of the crust in mantle (Fig. 2).

In the simple scenarios outlined above, deep earth processes interact with surface processes through the alteration of crustal lithostatic loading and isostatic responses in complex ways that...
depend on the history of deformation, erosion, and heat transfer within the crust. In addition, the pre-existing rheological state of the crust likely influences the deformation history during orogenesis. Consequently, the initial distribution of physical properties of the crust when combined with global and local climatic effects, establishes the strength of coupling between erosion and deformation throughout the history of orogenesis.

3. Physiographic, geologic, and climatic framework of the central Andes

3.1. Physiography

The central Andes are located between 70°W and 60°W, and 15°S and 35°S (Fig. 1) along an active subduction margin in which the oceanic Nazca Plate is translated ENE relative to stable South America and is underthrust beneath the South American Plate along an eastward dipping subduction interface (Stauder, 1975; Barazangi and Isacks, 1976; Hasegawa and Sacks, 1981; Bevis and Isacks, 1984; Cahill and Isacks, 1992; Allmendinger et al., 1997). This geodynamic configuration has led to the formation of a broad swath of high topography that roughly parallels the subduction zone (Fig. 1). This topography reaches peak elevations of 6962 m at Mt. Aconcagua (Argentina), the highest point in the western hemisphere. The elevation and extent of the high topography changes with latitude along the subduction zone: at ~18°S, the meridional extent of the Andes reaches widths of ~750 km, whereas it becomes narrower both north and south of this location (Fig. 1).

A number of morphotectonic segments can be defined in the central Andes based on topographic attributes. The forearc domain comprises, from west to east, the continental margin, the Coastal Cordillera (up to 2000 m in elevation), the Central Depression, and the North Chilean Precordillera reaching an average elevation of 3000 m (e.g., Reutter et al., 2006). To the east, the Western Cordillera consists of the Miocene to Quaternary volcanic arc atop the subduction zone (e.g., Isacks, 1988; Fig. 1). Between the latitudes of 15°S and 26°S, the Western Cordillera is flanked on its eastern side by the high-altitude internally drained Altiplano–Puna Plateau (Fig. 1). Elevations within the Altiplano average ~3700 m (Isacks, 1988), whereas the Puna is generally higher and has average elevations of ~4400 m (Whitman et al., 1996). Across the Altiplano, the Plateau surface is mostly flat; however, the Puna is partitioned into a series of smaller basins by generally N–S trending basement ranges that reach elevations of about 5000 m (Coutand et al., 2001; Fig. 1). South of the Puna (~26°S), the Principal, Frontal, and Pre-Cordilleras, which consist of a series of relatively narrow, N–S trending ranges that are externally drained (e.g., Jordan et al., 1993; Fig. 1). To the east at these latitudes, the central Andes consist of the Sierras Pampeanas, which is a zone of discontinuous and widely spaced mountain ranges that are broadly distributed across the foreland (Fig. 1; Stelzner, 1923; González-Bonorino, 1950; Jordan and Allmendinger, 1986). Bounding the eastern edge of the internally drained plateau north of ~26°S is the Eastern Cordillera, which consists of a series of high ranges that generally serve as the drainage divide between the internally drained plateau and the externally drained foreland (Kennan et al., 1995; Kley, 1996). In a transitional area between ~24°S and ~26°S, the Eastern Cordillera and the Sierras Pampeanas intermingle, with some Pampean-style ranges located to the east of the Eastern Cordillera. However, north of ~24°S, these vanish and give way to a series of ranges that closely flank the Eastern Cordillera called the Santa Barbara system (Rolleri, 1976; Baldis et al., 1976). Finally, north of the Santa Barbara system, a series of ranges lying east of the Eastern Cordillera comprise the Interandean and Subandean ranges (Allmendinger et al., 1983; Fig. 1).

On average, elevations taper gently toward the foreland from 6000–5000 to 200–100 m, although high local slopes result from the partitioning of the foreland into a series of meridionally extensive, rhythmic and coherent ridges.

3.2. Relationship of morphology to current plate geometry

The width and height of the orogen mimic changes in the geometry of the subducting Nazca plate. In the portion of the central Andes between 15° and 27°S, the orogen reaches a maximum width of 650–700 km, the downgoing Nazca plate dips steeply (~30°) down to 600 km beneath South America (Fig. 1B). By the time the Altiplano–Puna Plateau tapers both to the south and north, coinciding with a significant narrowing of the orogenic system, the slab partially shoals and dips shallowly (~5°) between depths of 100 and 125 km underneath South America (e.g., Bevis and Isacks, 1984, Barazangi and Isacks, 1976; Cahill and Isacks, 1992). This phenomenon is reflected in the symmetry of the Andean topography—the pole to the great circle about which topography is most symmetric matches that about which slab dip direction is also most symmetric (Gephart, 1994). In fact, this pole also corresponds to the modern rotation pole of the Nazca plate relative to a fixed South American plate, suggesting that the symmetry of the orogen is tightly linked to the subduction of the Nazca plate beneath South America (Gephart, 1994).

We show the topography of the central Andes projected about the great circle defining its symmetry axis in Fig. 3A. When looking at the average difference between elevations located north and south of this symmetry axis, the orogen is remarkably symmetric (Fig. 3B), although important excursions from this symmetric view of the Andes exist. When considering the average difference between elevations (e.g., Gephart, 1994), differences within the external portions of the margin are naturally de-emphasized, since they lie at low elevations and are unlikely to have the corresponding differences in elevation expected in steep terrains. To adjust for this, we normalized the difference between points north and south of this symmetry axis to their mean elevation. This allows small differences in low relief areas to have similar weight as large differences at high elevation (Fig. 3C). Using this approach, we see that the center portion of the mountain belt constituted by the Altiplano–Puna Plateau is highly symmetric; however, topography along the margins and within the foreland displays marked asymmetries (Fig. 3C). In particular, the enclave into the eastern-most portion of the Andes north of the symmetry axis that is absent to the south constitutes a highly asymmetric feature that was obscured using the dimensional difference (compare Fig. 3B and C). In addition, the margins of the Eastern Cordillera, as well as the whole of the Pampean and Santa Barbara ranges are highly asymmetric with respect to their counterparts to the north. Thus, while the extent and location of the Plateau appears highly symmetric, the external portions of the mountain belt, including the Eastern Cordillera, Principal, Frontal, and Pre-Cordilleras, the Sierras Pampeanas and Santa Barbara system, and the Interandean and Subandean ranges are far less so.

3.3. Pre-Cenozoic geologic history

The oldest rocks exposed within the back-arc area, south of 15°S are located within the Sierras Pampeanas (Fig. 1) and include Mesoproterozoic high-grade metamorphic basement consisting of gneisses and migmatites (Mon, 1979; Ramos et al., 1986) that may be locally intruded by several generations of late Neoproterozoic to early Paleozoic granitoids (Halpern and Latorre, 1973; Lork and Bahlburg, 1993). Late Neoproterozoic to Early Cambrian green-schist-grade turbidites and pelagic clays of the Puncoviscana Formation appear within the Puna and adjacent Eastern Cordillera (NW Argentina and southernmost Bolivia) and were deposited in a deep marine basin.
thickness is ~2000 m; Ruiz Huidobro, 1960; Turner, 1960a,b; Aceñolaza, 1973; Omarini, 1983). Importantly, the transition between the crystalline basement and the Puncoviscana Formation is abrupt, and corresponds to the boundary between the Sierras Pampeanas and Eastern Cordillera at ~26°S (Fig. 1). Whether this sharp transition represents a buttress unconformity between a southerly crystalline basement high (potentially the Pampia terranes, e.g., Ramos, 2008) within the Early Cambrian, or motion along a fault is unclear; however, it certainly marks a persistent and profound change in the paleogeography of the central Andes that occurred at least as early as Cretaceous time, and perhaps during the Proterozoic to early Paleozoic (Allmendinger et al., 1983). These units are locally unconformably covered by Cambrian quartzites (Hausen, 1925) deposited in an extensional back-arc basin (Sallity et al., 1976) associated with an extensional episode affecting the southwestern margin of Gondwana (Ramos et al., 1986; de Wit and Ransome, 1992; Bahlburg and Furlong, 1996). During the Ordovician, an up to 4000-m-thick turbiditic section is deposited across the Puna and Eastern Cordillera in Argentina (Bahlburg and Breitkreuz, 1991; Bahlburg and Furlong, 1996). In the late Ordovician, these units were deformed and metamorphosed to greenschist grade (Turner and Méndez, 1979; Coira et al., 1982; Mon and Hongn, 1987) and subsequently intruded by a granitoid and rhyodacitic belt known as the Faja Eruptiva de la Puna Oriental (FEPO; Mendez et al., 1972). The FEPO is interpreted as representing a late Ordovician to Silurian magmatic arc that resulted from a subduction zone located to the west of the Puna, which likely dipped to the east (Coira et al., 1982). In northwestern Argentina, Silurian–Devonian sedimentation is mainly localized in the present-day Subandean ranges and eastern part of the Eastern Cordillera, and comprises about 2000 m of marine sandstones and shales (Turner, 1960a,b; Padula et al., 1967). To the west of the FEPO, Carboniferous and Permian plutons indicate that this margin was active during the Late Paleozoic.

The activity of the convergent margin during the Paleozoic is partially represented within the sedimentary sequence lying east of the current Altiplano–Puna Plateau margin. These Paleozoic strata are present north of the crystalline basement/Puncoviscana transition around ~26°S (Allmendinger and Gubbels, 1996, their Fig. 6). In parts of the Sierras Pampeanas to the south, lower Paleozoic strata consist of both marine and continental strata deposited in an extensional to...
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trans-tensional tectonic environment, while Permian strata record filling and amalgamation of these sub-basins (Fernandez-Seveso and Tankard, 1995; Fisher et al., 2002). In total, these Paleozoic strata generally reach thicknesses of ~2.5 km. As an exception, Paleozoic sediments currently exposed within the Precordillera of Argentina (Fig. 4F) reach thicknesses of up to 7 km (e.g., Zapata and Allmendinger, 1996). Between ~26°S and ~24°S, Paleozoic sedimentary rocks are often present, but generally of limited thickness. However, north of ~24°S, a thick, eastwardly tapering section of Paleozoic clastic rocks reaches a thickness of ~12 km in the Eastern Cordillera of Bolivia and thins to ~4–9 km at the edge of deformation in the Chaco Plain (Allmendinger et al., 1983; Allmendinger and Gubbels, 1996). These rocks were apparently deposited in the retro-arc basin of the Paleozoic volcanic arc (e.g., Ramos, 2008). Thus, as with the Precambrian and earliest Cambrian rocks in the central Andes, the Paleozoic section starts to thicken around ~26°S, reaches maximum thicknesses in the current Bolivian foreland (e.g., Allmendinger and Gubbels, 1996; McQuarrie and DeCelles, 2001, and references therein) and finally tapers again northward to disappear around 17.5°S (Baby et al., 1994).

Mesozoic strata exposed within the central Andes generally record an episode of intracontinental rifting starting in the Neocomian (Salfity, 1982). This rift basin system has a complex geometry with N–S to E–W elongated depocenters (see Fig. 1 in Marquillas et al., 2005) filled with strata of the Salta Group (Turner, 1959), which is Early Cretaceous to Late Eocene in age (e.g., Marquillas et al., 2005). Deformed remnants of these basins are preserved in the Altiplano–Puna Plateau, the Eastern Cordillera and the foreland thrust belts. In the latter, Mesozoic strata are of limited thickness or altogether absent south of ~27.5°S, reach a maximum thickness of ~4–5 km between 25° and 26°S in the Santa Barbara system and disappear north of 23°S when the Lomas de Olmedo branch diverges eastward (see Fig. 1 in Marquillas et al., 2005; Salfity and Marquillas, 1994).

A plot of pre-Cenozoic stratigraphic thickness in the modern foreland with latitude shows the abrupt changes in thickness that accompany the widespread exposure of Precambrian basin south of ~26°S (Fig. 4F). Importantly, changes in basin thickness correlate strongly with the different morpho-tectonic provinces in the central Andes: the thick (up to 17 km) pre-Cenozoic foreland basin units are associated with the fold-and-thrust belts in the Subandean and the Precordillera fold-and-thrust belts, the irregular Cretaceous rift basins are associated with thick-skinned deformation in the Santa Barbara system, and relatively thin (or absent) pre-Cenozoic basin strata are often associated with the Sierras Pampeanas basement-cored uplift province (Fig. 4F).

3.4. Cenozoic geologic history and deformation

The style, location, and amount of Cenozoic deformation correlates with zones defined by the current plate-boundary geometry, as well as by pre-Cenozoic paleogeography of the central Andes. While the timing of Andean deformation within the Puna–Altiplano Plateau and bounding Eastern Cordillera are somewhat disputed and certainly highly diachronous, the commencement of modern deformation in this area is broadly constrained to be between late-Eocene and mid-Miocene time (e.g., Allmendinger, 1986; Horton, 1997, 1998; Lamb and Hoke, 1997, Marrett and Streecker, 2000; Coutard et al., 2001, 2006; Hongn et al., 2007).

3.4.1. Altiplano–Puna Plateau

Apatite fission-track thermochronology data from the Altiplano–Puna Plateau indicate that exhumation of this part of the orogen has been limited (Carrapa et al., 2005; Deeken et al., 2006; Ege et al., 2007; Letcher, 2007). The Bolivian Altiplano records exhumation in the Early Oligocene (33–27 Ma) at slow and constant rate of 0.2 mm/yr until the late Miocene (11–7 Ma) (Ege et al., 2007). In the Puna Plateau, the onset of exhumation is recorded by apparent AFT ages between 24 and 29 Ma (Deeken et al., 2006; Carrapa et al., 2005; Letcher et al., unpublished data). If exhumation serves as a crude proxy for total tectonic shortening and rock uplift, these data suggest that shortening within the interior of the Altiplano–Puna Plateau has been limited during the last 20–25 Ma. The deformation that did occur within the Puna is more distributed than the Altiplano Plateau to the north and has uplifted N–S-trending bedrock blocks that partition the plateau into a series of ranges and adjacent basins. These basins contain the detritus that presumably resulted from uplift and exhumation of the ranges in the Plateau interior (Alonso et al., 1991; Vandervoort et al., 1995; Coutard et al., 2001). Basins that record the partitioning of the Puna into disconnected depocenters may have formed as early as the Oligocene, and certainly by the Miocene (Vandervoort et al., 1995). In fact, recent sedimentologic and structural data from the Puna have been interpreted as evidence for Eocene exhumation and depocenter development (Carrapa and DeCelles, 2008). Some of these basins are up to 5-km thick (Coutard et al., 2001), and those basins whose tops define the low-relief high elevation Plateau surface are internally drained (Isacks, 1988; Vandervoort et al., 1995; Sobel et al., 2003). Shortening estimates within the Puna are limited; however, those palinspastic reconstructions that exist suggest that less than 10–15% shortening (~37 km) has occurred within the Puna during the Cenozoic (Coutard et al., 2001).

The location and style of shortening within the interior of the Altiplano Plateau and along its margins differs from that of the Puna (e.g., Allmendinger, 1986; Lamb and Hoke, 1997; Lamb et al., 1997; Coutard et al., 2001; McQuarrie and DeCelles, 2001; Elger et al., 2005). The eastern margin of the Puna–Altiplano Plateau undergoes a transition around ~23°S in which thick-skinned, high angle reverse faults accommodate deformation to the south, while a west-verging back-thrust belt accommodates deformation to the north (Sempere et al., 1990; McQuarrie and DeCelles, 2001; Elger et al., 2005). Within the Altiplano itself, ~65 km of shortening appears to be accommodated between the Oligocene and mid-Miocene, with deformation migrating to the orogen’s margins after that time (Lamb et al., 1997; Lamb and Hoke, 1997; Elger et al., 2005). Deformation within the Plateau and along its margins has created up to a 12-km-thick depocenter (Sempere et al., 1990; Kennan et al., 1995; Lamb and Hoke, 1997; Horton et al., 2002) that is currently internally drained, and may have been so since Oligocene time (Horton et al., 2002; Sobel et al., 2003). Deformation within the Altiplano did not partition the depocenter into discrete, isolated
units to the same degree as in the Puna (Allmendinger et al., 1997, and references therein), creating the broad, largely uninterrupted, low relief Plateau surface (Fig. 4A).

Paleobotanical and stable-isotope data suggest that the broad, high-elevation portion of the central Andes, comprised mainly of the Puna–Altiplano Plateau abruptly rose sometimes after 10 Ma. In particular, paleobotanical evidence indicates that the Plateau had obtained no more than a third of its modern elevation at 20 Ma, and had risen to no more than half of its modern elevation by ~11 Ma (Gregory-Wodzicki, 2000). Likewise, fractionation of oxygen isotopes recorded by paleosol carbonates, lacustrine carbonates, sandstone cements, and carbonates formed in paleo-marshes, indicate that the surface of the Bolivian Altiplano was uplifted ~2.5–3.5 km between 10.3 and 6.8 Ma (Garzione et al., 2006). These results are consistent with paleothermometric data from carbonates of the area, which have been interpreted to reflect ~3700 ± 400 km of surface uplift between 10.3 and 6.7 Ma (Ghost et al., 2006). Thus, at least within the Plateau, elevations appear to have increased during the late Cenozoic in what appears to be a relatively short-lived phase of rapid surface uplift (Garzione et al., 2006). This rapid uplift has been attributed to the convective removal of eclogitized lower crust and lithosphere mantle beneath the Altiplano–Puna Plateau (Molnar and Garzione, 2007; Hoke and Garzione, 2008).

3.4.2. Eastern Cordillera

Fission track data from the Eastern Cordillera record the earliest onset of Cenozoic exhumation in the orogen during Late Eocene–Early Oligocene in Bolivia (40 Ma) (Ege et al., 2007) and further south in Argentina (40–38 Ma) (Coutand et al., 2001, 2006). Interestingly, the end of exhumation in the Eastern Cordillera is diachronous in the North and the South. In Bolivia, it ceases in the Late Oligocene–Early Miocene, after which time exhumation resulting from tectonic rock uplift is transferred to the foreland area (Ege et al., 2007), while in Argentina it persists during the Late Miocene in the Angastaco area (Deeken et al., 2006; Coutand et al., 2006) until contractual deformation is transferred eastward to the Santa Barbara System in the Pliocene (Kley and Monaldi, 2002).

The Eastern Cordillera is a basement-involved thrust system (Kley et al., 1996; Kley, 1996). In Argentina, Cenozoic deformation is taken up along moderate- to steeply-dipping reverse faults that accommodate as much as 55–85 km of horizontal displacement relative to stable South America (Grier et al., 1991; Baby et al., 1997; Coutand et al., 2001). Nested in this range, intramontane basins preserve up to 6-km-thick Cenozoic sediments (Grier et al., 1991; Marrett and Strecke, 2000; Coutand et al., 2006) that have been variously cannibalized, reworked and deformed. In the southern Eastern Cordillera, the current margins of the Cenozoic basins closely correspond with the paleo-borders of Cretaceous rift basins, emphasizing the impact of pre-existing structure in the modern architecture of the range (Grier et al., 1991; Deeken et al., 2006; Carrera et al., 2006; Carrera and Muñoz, 2008). Further north at ~22’S, the thickening of the Paleozoic section triggers a change in the style and amount of deformation accommodated within this unit with the development of a back-thrust belt along the western flank of the Eastern Cordillera and a deformation largely accommodated along detachment horizons interlayered in the Paleozoic section (e.g., Sempere et al., 1990; McQuarrie, 2002; Elger et al., 2005). Correspondingly, the Eastern Cordillera absorbs a larger amount of deformation, with total displacement accommodated increasing to ~100–140 km (Baby et al., 1997; McQuarrie and DeCelles, 2001; McQuarrie, 2002; Elger et al., 2005; McQuarrie et al., 2008a). Thus, while the Eastern Cordillera varies little in its meridional location, the style and amount of deformation accommodated by these ranges changes abruptly along the strike. These changes spatially coincide with the distribution of pre-existing Late Paleozoic detachment horizons (e.g., Allmendinger et al., 1983), again highlighting the deep influence of pre-existing paleogeographic features on the modern structure of the range.

3.4.3. Southernmost central Andes

South of the termination of the Puna Plateau, the Aconcagua fold-and-thrust belt, the Frontal Cordillera, and Precordillera define the eastern slope of the Andes (Fig. 4A). The emergence of the thin- and thick-skinned fold-and-thrust belts at this latitude corresponds to the re-emergence of a thick pre-Tertiary sedimentary section in the modern foreland (Allmendinger et al., 1983; Ramos et al., 2002) that appears to have been deformed, and in some places continues to deform, from the mid-Miocene to present (e.g., Allmendinger et al., 1983; Ramos et al., 2002). At ~33’S, geologic relationships indicate that deformation propagated eastward with time, and is broadly constrained to have occurred between 15 and 9 Ma in the Aconcagua fold-and-thrust belt, 9–6 Ma in the Frontal Cordillera, and 5–2 Ma in the Precordillera (Ramos et al., 2002). Further north at ~30.5’S, deformation within the Precordillera seems also to have migrated eastward with time, although the timing of deformation is slightly different than to the south (Allmendinger et al., 1990; Jordan et al., 1993). In particular, deformation within only the Precordillera appears to have begun around ~20 Ma, and propagated eastward in a somewhat discontinuous manner until the present (Jordan et al., 1993). In either case, deformation is broadly bracketed to have commenced in the early- to mid-Miocene in this area, and currently continues to deform rocks at the boundary between the Precordillera and the Sierras Pampeanas provinces (e.g., Zapata and Allmendinger, 1986b). Despite its relatively narrow width, total Cenozoic displacement accommodated within the Precordillera ranges from ~108 to 135 km (relative to fixed South America) (Allmendinger et al., 1990; Cristallini, 1996; Ramos et al., 2004, 1996a,b).

3.4.4. Sierras Pampeanas

The easternmost portions of the central Andes comprise, from south to north, the Sierras Pampeanas, Santa Barbara system, and Subandean fold-and-thrust belt, respectively (Figs. 1 and 4A). Within the Sierras Pampeanas, deformation is accommodated by high-angle reverse faults whose locations often spatially correspond to terrain boundaries from the Paleozoic, or inverted normal faults formed during the Cretaceous extension (Allmendinger et al., 1983; Schmidt et al., 1995; Ramos et al., 2002). Movement along these thrust faults uplifts ranges of Late Proterozoic to Early Paleozoic metamorphic bedrock (Allmendinger et al., 1983; Jordan and Allmendinger, 1986). The utilization of pre-existing crustal weaknesses, whose along-strike extent is often limited, causes these ranges to be discontinuous (e.g., Schmidt et al., 1995; Allmendinger and Gubbelts, 1996). Intramontane depocenters associated with these ranges have been enduringly disconnected from each other and can preserve sediments up to 4000 m in thickness (Sosc, 1972; de Ureizita et al., 1996; Fisher et al., 2002). AFT data indicate that the frontal part of the Sierras Pampeanas (Sierra de Aconcagua) was rapidly exhumed about 5 Ma ago (Sobel and Strecke, 2003) and warping of erosional surfaces in the Northern Sierras Pampeanas indicates that deformation remains active in this area (e.g., Strecke et al., 1989). The Pampean Ranges accommodate only about 10–20 km of horizontal displacement relative to fixed South America (Fig. 4A, C and D) (Costa, 1992; Ramos et al., 2004).

3.4.5. Santa Barbara system

At about 26’S, the northern boundary of the Sierras Pampeanas is marked by an abrupt decrease in the wavelength of structures. There, the Santa Barbara system (Figs. 1 and 4A) accommodates deformation
in the foreland largely by inversion of rift-related normal faults (Kley and Monaldi, 2002). The thickness of Cretaceous strata varies significantly according to the spatial distribution of the rift depocenters (see Fig. 1 in Monaldi et al., 2008; Fig. 4F). The thickness of Cenozoic strata exposed within the Santa Barbara ranges varies between 2.5 and 5.5 km, the lower part being comprised of sediments reflecting the post-rifting thermal subsidence stage and the upper part being foreland basin strata related to the inversion of the rift (Kley and Monaldi, 2002). While the utilization of high angle structures has created significant topography, it has also allowed only modest amounts of shortening to be accommodated: total displacement relative to stable South America is estimated to be between 25 and 30 km (Fig. 4B, C, and D) (Kley and Monaldi, 2002). The timing of contractual deformation is still controversial, but appears to have mostly taken place during the Pliocene (Kley and Monaldi, 2002). The physiographic boundaries of the Santa Barbara system appear to be strongly controlled by the pre-Cenozoic paleogeography of the Andes. The southern extent of the Santa Barbara system lies at ~26°S, where the bedrock abruptly transitions from the greenschist-grade Neoproterozoic to early Cambrian Puncoviscana Formation to Mesos-Neoproterozoic migmatisites, gneisses, and granitoids (Allmendinger et al., 1983) and both the southern and northern boundaries of these ranges coincide with the termination/bifurcation of the rift system (Kley and Monaldi, 2002). Specifically, the bifurcation of the Lomas de Olmedo depocenter to the north, causes the strike of old Cretaceous normal faults to become roughly orthogonal with respect to the Cenozoic shortening direction (Kley et al., 1999), rendering their reactivation mechanically difficult, and together with the tapering of Mesozoic deposits, apparently leads to the termination of the Santa Barbara system.

3.4.6. Subandean ranges
In Bolivia, the eastern margin of the central Andes consists of the Interandean Zone and the Subandean fold-and-thrust belt, which we consider together in this discussion for simplicity, although we acknowledge that important differences between these two zones exist (e.g. Kley and Reinhardt, 1994). East of the Bolivian Eastern Cordillera, a thin-skinned fold-and-thrust belt has formed within the thick Paleozoic sedimentary section (Allmendinger et al., 1983). Deformation in this zone likely commenced during the early Miocene and has propagated eastward with time (Kley, 1996). AFT data also document the migration of exhumation from the EC around 30–20 Ma to the frontal part of the subandes in the late Miocene–Pliocene (~8–2 Ma) (Ege et al., 2007). This eastward propagation has contributed to the deposition of late Cenozoic foredeep sediments of up to ~5 km thick in front of the advancing tip of the fold-and-thrust belt that was, and continues to be cannibalized as the belt widens (e.g., Kley, 1996; Horton, 1997). Magnitude of relative displacement varies from north to south: in the north, it accommodates between 74 and 135 km of relative displacement relative to stable South America (Roeder, 1988; Baby et al., 1989, 1997; Roeder and Chamberlain, 1995), in its center ~163 km (McQuarrie, 2002), and in the south between 125 and 150 km (Kley, 1996; Baby et al., 1997) (Fig. 4B, C, and D). Mechani
cally weak horizons in the Silurian, Devonian, and Carboniferous part of the section accommodate the formation of many of the decollement surfaces of the fold-and-thrust belt (e.g., Baby et al., 1990; Dunn et al., 1995). The geometry of these detachment horizons also changes from north to south—in the north, the basal decollement of the fold-and-thrust belt is ~4°, whereas it decreases to ~2° in the south—this southward shoaling corresponds to a widening of the fold-and-thrust belt (Roeder, 1988; Roeder and Chamberlain, 1995; Kley, 1996; Horton, 1999). Likewise, some interpretations of GPS data (Norabuena et al., 1998) suggest that the distribution of active deformation within the fold-and-thrust belt changes along its length, with out-of-sequence thrusting characterizing the northern part of the wedge (Horton, 1999), and deformation localized at its eastern toe in the south. Finally, the fold-and-thrust belt disappears to the south due to the tapering of the Silurian and Devonian decollements in the Paleozoic basin (Belotti et al., 1995). This leaves a gap of ~60 km between the southern termination of the Subandean belt and the emergence of the Santa Barbara system to the south (Kley and Monaldi, 2002). Thus, the Subandean belt is closely associated with the presence of mechanically weak detachment horizons in pre-existing basin deposits (e.g., Allmendinger et al., 1983).

3.5. Current and Cenozoic climate
Today, the Andes constitute a major topographic barrier to atmospheric circulation in South America and separate the humid lowlands in the east from the arid Pacific margin in the west (see Fig. 1b in Bookhagen and Strecker, 2008). The semiarid to hyperarid climates observed on the western slope of the orogen and across the high Plateau (Fig. 1b; Strecker et al., 2007, and references therein), result from their proximity to upwelling of cold water along the Pacific coast due to the cold Humboldt current, which may have existed in some form since the early Cenozoic (Zachos et al., 2001), and the fact that these regions are located within a high-pressure sector with atmospheric subsidence (Houston and Hartley, 2003). Precipitation in the Central Andes is largely derived from the westward penetration of Atlantic moisture from the Chaco Plain and Argentine foreland (Rohmender, 1943; Halloy, 1982; Bianchi and Yañez, 1992; Garreau et al., 2003) via the Andean low-level jet (Nogués-Paegle and Mo, 1997). As moisture encounters the topography of the eastern Andes, precipitation is orographically enhanced (e.g., Bianchi and Yañez, 1992; Sobel et al., 2003; Strecker et al., 2007; Bookhagen and Strecker, 2008). Both the relief structure of the eastern flank of the central Andes and the source of incoming precipitation cause large latitudinal changes in the amount of rainfall reaching the eastern slope and interior of the orogen (Fig. 1; Strecker et al., 2007; Bookhagen and Strecker, 2008). In northern Bolivia at ~15–17°S, orographic rainfall peaks above 3.5 m/yr at a mean elevation of 1.3 km once topographic relief of about 1 km is achieved (Bookhagen and Strecker, 2008). South of 17°S, orographic precipitation decreases as the mountain belt widens (Montgomery et al., 2001) but is still on the order of 1.5 m/yr at 26°S (Bookhagen and Strecker, 2008). As the long, linear ranges of the Subandean fold-and-thrust belt give way to the largely discontinuous, steep-flanked and higher elevated ranges of the Santa Barbara system and Sierras Pampeanas, precipitation is strongly focused on the windward sides of these ranges (Strecker et al., 2007; Hilley and Strecker, 2005). Hence they shelter the eastern slope of the inner orogen from precipitation derived from the Atlantic (Fig. 1b). As a consequence, most areas of the Sierras Pampeanas, and the Principal, Frontal, and Pre-cordilleras receive little moisture from the east (Sobel and Strecker, 2003).

Since ~18 Ma, the latitudinal location of the central Andes has remained relatively stable (Scotese et al., 1988). In addition, the atmospheric and oceanic circulation that exists today appears to have been established since at least the early Cenozoic (Parrish et al., 1982), and perhaps the Mesozoic (Hartley et al., 1992). In the broadest sense, the persistent latitudinal position and stable atmospheric circulation has created a generally arid climate within the central Andes, with the western portions of the orogen characterized by the most intense aridity, while the eastern flanks generally receive far more moisture (Fig. 1; Strecker et al., 2007). While this overall pattern has persisted since the Miocene, paleoclimate proxies record significant climate changes throughout this period (Strecker et al., 2007, and references therein). Furthermore, along the western Andean margin, late Miocene to early Pliocene paleoclimate indicators are somewhat contradictory and vary latitudinally (Sáez et al., 1999; Gaupp et al., 1999; May et al.,...
2005); nonetheless, generally arid to semi-arid conditions with periods of greater moisture availability typify the western margin of the orogen since the mid-Miocene, and perhaps as early as the Eocene (Blanco et al., 2003). The termination of supergene alteration within the Atacama Desert and volcanic arc between 14 and 8 Ma (Alpers and Brimhall, 1988; Sillitoe et al., 1991) suggest that arid conditions commenced in these areas by mid-Miocene time. In addition, the deposition of thick halite and gypsum-bearing sedimentary units between 24 and 15 Ma within the Puna basins, indicate that by this time, the climate was arid (Alonso et al., 1991; Vandervoort et al., 1995).

In contrast, sediments found in areas east of the Puna and Altiplano show evidence for a more complex and variable climate history that appears strongly affected by the uplift of individual ranges and the resulting downwind aridification (Strecker et al., 2007). For example, intramontane basins in the Eastern Cordillera of northwestern Argentina preserve sediments that record a shift from semi-arid/arid conditions to humid environment in the late Miocene (Marshall and Patterson, 1981; Pascual et al., 1985; Pascual and Ortiz Jaureguizar, 1990; Nasif et al., 1997; Kleinert and Strecker, 2001; Starck and Anzótegui, 2001; Coutand et al., 2006; Strecker et al., 2007; Anzótegui, 2007; Hilley and Strecker, 2005). Further to the north in the Subandean ranges of Bolivia, climate also shifted towards more humid conditions in the late Miocene (Uba et al., 2005, 2006), a time during which megafans were developed in this area (Horton and DeCelles, 2001). In most of these marginal basins, the late Miocene change to more humid conditions was abruptly terminated as the uplift of upwind ranges prevented moisture from penetrating into these areas (Sobel et al., 2003; Sobel and Strecker, 2003; Coutand et al., 2006; Strecker et al., 1989, 2007, Hilley and Strecker, 2005). The timing of the onset of ensuing aridification varies from basin to basin, depending on the local tectonic history that uplifted the upwind orographic barriers (e.g., Strecker et al., 2007). This tectonic control on local precipitation patterns is observed in the modern rainfall distribution on the eastern slope of the Andes (Figs. 1 and 4). In summary, both past and current climate generally show an arid or-hyperarid western slope, an arid interior, and an eastern slope that varies from humid to arid, depending on local topography and latitudinal position.

4. Relationships between paleogeography, climate, and tectonics in the central Andes

The location of paleogeographic provinces, total shortening, downgoing slab geometry, mountain belt width, and climate correspond to varying degrees within the central Andes (Fig. 4). In summarizing the above discussion, between ~15° and 22.5°–23°S total shortening is large (40%–70%) and is concentrated within the Eastern Cordillera and the Subandean ranges (Fig. 4 C, D, and E). The subducted Nazca plate dips steeply eastward beneath continental South America (Fig. 4G) and the location of the foreland fold-and-thrust belt coincides with pre-existing thick Paleozoic deposits containing multiple stratigraphic horizons that serve as decollements (e.g., Allmendinger et al., 1983; Kley et al., 1999). However, from 17.5°S, the abrupt increase of the mountain belt’s width (Fig. 4A) spatially coincides with a decrease in mean annual precipitation (WMO, 1975; Montgomery et al., 2001) (Fig. 4B), the shallowing of the basal decollement angle of the orogenic wedge (Roeder, 1988; Roeder and Chamberlain, 1995; Kley, 1996; Horton, 1999) and the distribution of active deformation, as documented by GPS data (Norabuena et al., 1998; Horton, 1999). Between ~22.5°S and 26.5°S, the central Andes accommodates less shortening (30 to 40%) (Figs. 4C, D and E) and this corresponds spatially to the disappearance of the subandean ranges and Paleozoic foreland deposits with interlayered decollements. In the foreland, shortening becomes accommodated by inversion of Cretaceous normal faults in the thick-skinned Santa Barbara ranges (Jordan et al., 1983; Kley and Monaldi, 2002). Commensurate with these changes in total displacement, total shortening, and structural style, the mountain-belt width also decreases (Fig. 4H). At ~26°S, the Santa Barbara system becomes transitionally replaced by the high-angle reverse faulted Sierras Pampeanas to the south. Both these ranges focus Atlantic-derived precipitation on their eastern margin, resulting in large precipitation gradients between the humid flanks and arid interior of the orogen (Fig. 1B; Hoffmann, 1975; Sobel et al., 2003; Sobel and Strecker, 2003; Hilley and Strecker, 2005).

Finally, between 28°S and 34°S, pre-Cenozoic deposits are absent in the foreland belt, and upper crustal shortening is taken up by reactivated Paleozoic structures across the Sierras Pampeanas (Allmendinger et al., 1983; Schmidt et al., 1995; Ramos, 2000; Ramos et al., 2002) that accommodate only 10–20 km of horizontal displacement (Fig. 4D). The distributed nature of the deformation widens the mountain belt and consequently decreases the total shortening (Fig. 4E). The subduction angle is low and therefore active volcanism is absent (Fig. 4G). Previous geologic studies have suggested that tractions and/or thermal heating at the base of the South American lithosphere may have caused the eastward migration of deformation in this area (Jordan et al., 1983; Ramos et al., 2002) (Fig. 4H). In the arid westmost portion of this zone (Figs. 1 and 4), the re-emergence of a thick pre-Cenozoic sedimentary section corresponds with the development of the Principal, Frontal, and the presently active Precordillera fold-and-thrust belt (Fig. 4A, D, and F), where a larger amount of displacement is accommodated than in the Sierras Pampeanas (Fig. 4A and D).

In the Central Andes, the structural style (thin-skinned versus thick-skinned fold-and-thrust belt, basement-cored uplifts) of the external portion of the orogenic system mirrors the paleogeographic zones that existed prior to the Andean orogenic event (Allmendinger et al., 1983). The fact that the amount of horizontal displacement absorbed by the various structural provinces is markedly different suggests that latitudinal changes in deformation may be dominantly controlled by factors set in motion long before the construction of the modern mountain belt (Allmendinger et al., 1983). Thus, the paleogeographic heritage likely triggers marked dissimilarities in the topography of the eastern Andean margin. In contrast, the topography of the high Plateau itself is remarkably symmetric (Fig. 3; Gephart, 1994). However, the amount and style of deformation accommodated along the eastern boarder of the high plateau within the Eastern Cordillera appears to show marked asymmetry along its length. In Bolivia, the Plateau boundary is seated within a thick section of Paleozoic metasediments, and a large amount of shortening is accommodated within the Eastern Cordillera and its backthrust belt (McQuarrie and DeCelles, 2001; Elger et al., 2005). In contrast, along the Argentine Puna margin, the Paleozoic deposits generally tend to become thinner, high-angle reactivated structures accommodate deformation and total shortening decreases (Baby et al., 1997; Kley et al., 2005). Thus, while the location of the eastern margin of the Plateau appears to reflect the geometry of the downgoing Nazca plate, the nature of deformation and the amount of displacement accommodated along the margin appears controlled by the paleogeography of those areas.

The above observations allow us to draw some inferences about the roles that slab geometry, paleogeography, and climate may play in shaping the central Andes. First, we observe that the extent of the Plateau coincides with major changes in the downgoing Nazca plate geometry, suggesting a causal link between the two (Gephart, 1994). As an alternative hypothesis, others have suggested that the
paleogeography of the South American craton may have influenced the slab geometry itself: in the south where pre-existing high-angle structures are pervasive within the crust, the slab was flattened as it was pushed underneath a continent that could not yield to the degree that it could in the north, where the Paleozoic basin allowed large amounts of shortening to be accommodated within the foreland (Allmendinger et al., 1983; Allmendinger and Gubbels, 1996). Nonetheless, a combination of the paleogeography and subduction zone geometry appears responsible for the large-scale architecture of the interior part of the orogen, rather than climate playing a primary role as some have previously suggested (e.g., Montgomery et al., 2001; Sobel et al., 2003). In addition, the amount and nature of deformation accommodated within the interior and external parts of the orogen varies from north to south, leaving a strong imprint on the topography of the mountain range: (1) Total W–E shortening decreases stepwise to the south; (2) These variations are focused on boundaries of tectonic provinces characterized by different styles of deformation; and (3) The total width of the mountain belt changes within each of these provinces. Changes in the precipitation distribution that result from both regional air-mass circulations and local orographic effects broadly coincide with these tectonic and topographic boundaries. However, the spatial coincidence of these boundaries with factors established prior to the Cenozoic makes it unlikely that the current climate plays a dominant role in determining the large-scale architecture of the central Andes.

5. What role might climate play in the tectonics and topography of the central Andes?

If the paleogeography and plate interface geometry have left such a strong imprint on the morphology of the modern central Andes, then what role, if any, might climate have played in controlling different tectonic features? The coincidence of pre-Andean paleogeography with the current Andean tectonic, climatic, and topographic attributes argues that most of the orogen-scale features of the central Andes have been shaped in large part by this paleogeography. However, it is possible that at a more local scale, climate, erosion, and tectonics may be linked. To investigate this option, we further analyze three distinct morphotectonic provinces in the central Andes: (1) the internally drained Altiplano–Puna Plateau, (2) the foreland fold-and-thrust belts of Bolivia and Argentina, and (3) the Sierras Pampeanas basement-cored uplift province (Fig. 4; labeled region 1, 2, and 3 respectively). The Altiplano–Puna plateau typically receives <400 mm of annual precipitation (WMO, 1975; Sobel et al., 2003; Bookhagen and Strecker, 2008), and stratigraphic data indicate that the current western and eastern Altiplano margins were uplifted in the Eocene–Oligocene, and Oligocene–Miocene, respectively, leading to the formation of internal drainage in the area by ~25 Ma (Horton et al., 2002). Likewise, in the Puna, uplift of its present eastern margin and disparate basement uplifts within the plateau region may have led to the formation of internal drainage between Oligocene and Miocene time (e.g., Coutand et al., 2001, Vandervoort et al., 1995; Coutand et al., 2006). While the extent and symmetry of the Plateau appears to reflect the downwarping Nazca plate geometry, erosion may help to reinforce the nature of the topography and tectonics of this area. In particular, the establishment of internal drainage decouples the local base-level of streams draining the hinterland from their regional base-level, allowing large-scale aggradation to occur (Sobel et al., 2003; Hilley and Streeker, 2005). Integrated field studies and theoretical models indicate that the disconnection and reintegration of individual basins base-levels may occur over timescales of ~10^5 years (Allen, 2008; Hilley and Streeker, 2005). The fact that regional disconnection of the foreland and Plateau base-levels have persisted for much longer (e.g., Vandervoort et al., 1995) indicates that conditions that favor this disconnection have persisted over these longer timescales. This has two important effects on erosional–tectonic coupling: (1) Regional aggradation reduces relief in landscapes as channels aggrade and peaks degrade, and slope reduction decreases erosion in the orogen’s interior; (2) Internal drainage and uplifting upwarp orographic barriers prevent mass evacuation from this region, increase lithostatic loads, and reduce the isostatically driven rock uplift that would otherwise result from mass removal. Maintenance of lithostatic stresses by restriction of erosional mass export may conspire with geodynamic processes to cause deformation to migrate to the external parts of the orogen (Sobel et al., 2003). Together, geodynamic processes and an arid climate decouples the Plateau’s base-level from the foreland, allowing a broad, low-relief, tectonically quiescent Plateau to persist over geologic time in a manner similar to the hypothetical case when erosion is absent (e.g., Royden, 1996).

The foreland fold-and-thrust belts of Bolivia and Argentina (region 2 in Fig. 4A) are consistently associated with thick pre-existing sedimentary units that contain mechanically weak layers that served as decollement horizons. These regions are externally drained; thus, erosional mass efflux increases with mountain-belt width and/or with steeper mean orogenic slopes parallel to material transport direction (e.g., Dahlen and Suppe, 1988; Willett et al., 1993). In contrast to plateau environments, field (Hilley et al., 2004), experimental (e.g., Persson et al., 2004; Hoth et al., 2006) and theoretical (e.g., Dahlen, 1984) studies suggest that the voracity of erosional processes likely play a role in controlling the long-term deformation and topography. Piggy-back basins in fold-and-thrust belts may act as significant storage areas for sediment over 10^5 year timescales (Allen, 2008; Hilley and Streeker, 2005). However, the resistant bedrock exposed within the ranges of the fold-and-thrust belt likely limit their rate of erosion (Whipple et al., 1999). Slope adjustments within the fold-and-thrust belt generally take ~10^8–10^7 years to balance erosional mass efflux with the flux of tectonically accreted material (e.g., Stock and Montgomery, 1999; Hilley et al., 2004). Both the Bolivian and Argentine fold-and-thrust belts have been active since the Miocene, suggesting that erosion and deformation in these areas have at least partially adjusted to one another. This adjustment may be reflected by changes in the geometry of the fold-and-thrust belt in space and/or time. For example, in northern Bolivia, a southward decrease in mean annual precipitation coincides with a widening of the fold-and-thrust belt (Fig. 4A and B) and perhaps a transition from diffuse to localized deformation along the topographic front of the Bolivian fold-and-thrust belt (Horton, 1999). Theoretical considerations suggest that weaker erosion, presumably caused by decreased precipitation rates, results in widening of the orogen to evacuate tectonically accreted foreland sediments (e.g., Whipple and Meade, 2004). Also, studies of the Argentine Principal, Frontal, and Precordilleras (Fig. 4) indicate that changes in erosion rates, prompted by the exposure of different rock types, may result in changing fold-and-thrust belt geometry and topography over time (Hilley et al., 2004). Thus, theory predicts that externally drained areas presenting the proper geologic conditions for the formation of fold-and-thrust belts, may show a strong imprint of climatically–moderated erosional processes on their morphology (Montgomery et al., 2001). In the central Andes, segments of the mountain belt experiencing less erosion, either due to lower precipitation rates (Montgomery et al., 2001) or to the exposure of more resistant lithologies (Hilley et al., 2004), tend to be wider than more erosive areas. This is consistent with the view that the geometry of fold-and-thrust belts depends on the efficiency of erosion (e.g., Dahlen and Suppe, 1988). However, within the Bolivian fold-and-thrust belt, recent area balancing of structural sections found little difference in shortening within this area, which may suggest that erosion may exert less of a control on orogenic structure here than previously thought (McQuarrie et al., 2008b).

In contrast to fold-and-thrust belts, deformation along reacti- vated crustal-scale heterogeneities such as in the Sierras Pampeanas

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(region 3 in Fig. 4A) involve steep, pre-existing crustal-scale weaknesses that may undergo frictional failure without necessitating failure within the rest of the mountain range. An analysis of the stresses expected in this situation suggests that the development of erosionally moderated surface slopes should be decoupled from the reactivation of these structures (Hilley et al., 2005), except when mean slopes across the range exceed a threshold value. In this case, the large lithostatic loads produced by these slopes oppose horizontal stresses and favor failure of frictionally stronger structures without such loads. Because erosional processes moderate surface slopes, vigorous erosion may prevent the build-up of steep slopes, and hence the migration of deformation to other structures in low-lying areas. As in fold-and-thrust belts, bedrock erosion is likely the process that limits the timescale over which erosion responds to tectonic uplift. However, the time it takes to build topography significant enough to cross this threshold slope depends on rates of uplift produced by fault motion, the erosion efficiency, the value of the threshold slope, and the geometry of the underlying fault. This time-scale varies depending on local circumstances, but likely occurs over ranges of 10^5–10^6 years. These theoretical considerations shed light on the development of two mountain ranges within the northern Sierras Pampeanas that have undergone similar geologic histories but have vastly different kinematics. Those ranges in wet areas tend to have deformation focused on a limited number of structures with large amounts of displacement, while ranges in wet areas tend to have deformation focused on a limited number of structures with large amounts of displacement, while those in drier sectors display more widely distributed deformation.

Several important lessons may be gathered from interactions observed within the central Andes. Most importantly, the nature of coupling between erosion, climate, pre-orogenic geologic history, plate-boundary forces, and mantle processes is complex, and the impact of the initial state of all of these variables may not easily be erased over the life of the orogen. While our previous work has tried to view the orogenic structure in terms of simple interactions between these factors (Sobel et al., 2003; Hilley et al., 2004; Hilley and Strecke, 2004; Hilley et al., 2005) and the work of others have attempted to understand overall orogenic development in terms of simple erosional and geodynamic feedbacks (e.g., Willett et al., 1993; Whipple and Meade, 2004), these approaches may not be altogether appropriate for an orogen such as the central Andes that has been subjected to a long and rich tectonic history. They may fail to reveal the complex interactions between tectonic and erosional processes that characterize these types of orogens. However, because factors affecting the Earth’s mantle, crust, and surface at the onset of and during mountain building profoundly influence the development of the orogen throughout its life, we argue that the path to understanding these orogens requires an integrative approach that combines geodynamic, geologic, and geomorphologic observations and models.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at 10.1016/j.tecto.2009.06.017.

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