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THE EVOLUTION OF THE ALTIPLANO-PUNA PLATEAU OF THE CENTRAL ANDES

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ABSTRACT

The enigma of continental plateaus formed in the absence of continental collision is embodied by the Altiplano-Puna, which stretches for 1800 km along the Central Andes and attains a width of 350–400 km. The plateau correlates spatially and temporally with Andean arc magmatism, but it was uplifted primarily because of crustal thickening produced by horizontal shortening of a thermally softened lithosphere. Nonetheless, known shortening at the surface accounts for only 70–80% of the observed crustal thickening, suggesting that magmatic addition and other processes such as lithospheric thinning, upper mantle hydration, or tectonic underplating may contribute significantly to thickening. Uplift in the region of the Altiplano began around 25 Ma, coincident with increased convergence rate and inferred shallowing of subduction; uplift in the Puna commenced 5–10 million years later.

INTRODUCTION

The Altiplano-Puna plateau of the Central Andes (Figure 1) is the highest plateau in the world associated with abundant arc magmatism, and it is second only to Tibet in height and extent. Yet, this remarkable feature was uplifted in the absence of continental collision or terrane accretion; in fact, material has been removed from the continental margin during and prior to plateau uplift. Because of its obvious association with Andean magmatism, the plateau was originally thought to be a product of magmatic processes (James 1971b,

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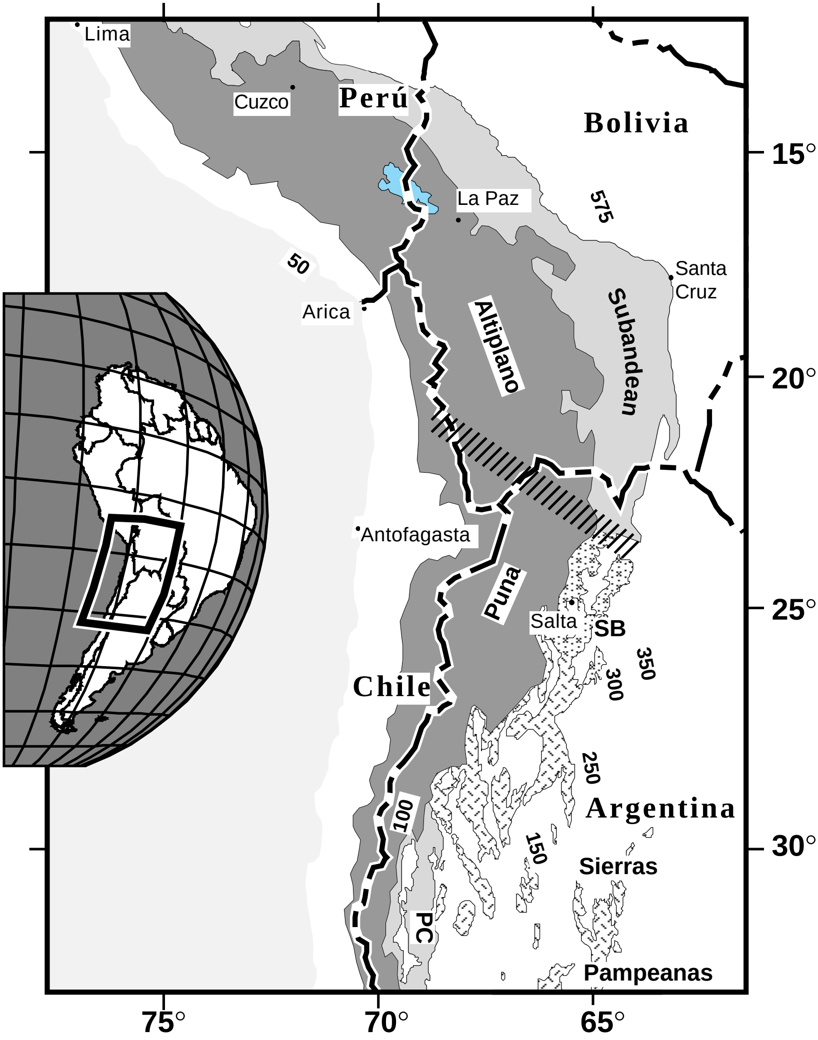


Figure 1 Location map showing the extent of the high plateau of the Central Andes. Dark gray shows area above 3 km elevation; the plateau is defined by the wide area above 3 km between 13 and 27°S. The light gray provinces east of the high topography are thin-skinned thrust belts in the Subandean ranges of Bolivia and the Precordillera (PC) in Argentina. The Sierras Pampeanas and Santa Bárbara System (SB) are thick-skinned foreland provinces. Thin curves are contours of depth to the Wadati-Benioff zone in kilometers from Cahill & Isacks (1992). The hachured zone trending NW-SE across the Argentine-Bolivian border corresponds to a variety of lateral change in Andean and pre-Andean features and is taken here to be the boundary between the Altiplano and Puna.

Reymer & Schubert 1984, Thorpe et al 1981). However, analyses of the plateau topography and structures on the eastern flank of the plateau carried out during the 1980s resulted in the conclusion that crustal shortening could produce most, if not all, of the required crustal thickening and that thickening, combined with lithospheric thinning, could account for the plateau elevations (Isacks 1988, Roeder 1988, Roeder & Chamberlain 1995, Sheffels 1990). Here, we review these arguments, as well as more recent results that appear to show that shortening may not be able to account for all of the crustal thickening.

The central Andean plateau must be viewed not just in terms of volumes and magnitudes, but also in light of its evolution. In this review, we focus on the temporal and spatial evolution of the plateau: when it began to lift up and how it varies laterally, as well as the relative importance of magmatism, crustal shortening, and lithospheric thinning. The plateau is composed of two distinct parts: the Altiplano of Bolivia and the Puna of northwest Argentina and adjoining parts of Chile. These areas differ in topography, magmatism, and lithospheric structure, and illustrate the range of conditions under which a continental plateau can develop in a noncollisional orogen.

The data that we review here supports and refines Isacks' (1988) two-stage model for the development of the plateau. Stage 1 uplift began around 25 Ma in the Altiplano segment and between 15 and 20 Ma in the Puna segment, when an episode of low-angle to, locally, nearly flat subduction (Coira et al 1993, Kay et al 1995) thinned and thermally softened the lithosphere underlying the area that was to become the plateau. Shortening ceased in the Altiplano and shifted eastward (Stage 2) beginning between 12 and 6 Ma, but shortening continued in the Puna until 1–2 Ma.

PHYSICAL DESCRIPTION OF THE PLATEAU AND RELATED FEATURES

A convenient definition of the high plateau of the Central Andes is provided by the notable broadening of the area above the 3-km elevation contour (Figure 1). Defined this way, the high plateau of the Central Andes stretches 1800 km along the backbone of the range, from southern Peru to northern Argentina, and varies between 350 and 400 km in width. This definition of the plateau, which follows that of Isacks (1988), is considerably broader than the more common association of the plateau with the internally draining basins of the Altiplano and Puna.

Plate Geometry

The geometry of the Nazca Plate beneath South America is well known (Barazangi & Isacks 1976, Bevis & Isacks 1984, Cahill & Isacks 1992, Hasegawa & Sacks 1981, Stauder 1975). Currently, the plateau correlates with a 30°–east-dipping segment of the subducted Nazca Plate (Figure 1). To the north and south, where the mountain belt narrows considerably, the subducted plate shallows and is nearly horizontal. Post-Pliocene volcanism follows this correlation: It is absent where the plate is nearly flat and well developed in the plateau where the plate is steeper. The distribution of Neogene volcanism is virtually identical to the spatial extent of the plateau, both latitudinally and longitudinally.

The subducted plate geometry differs markedly beneath the northern and southern ends of the plateau (Figure 1). Beneath southern Peru, there is a marked bend in the subducted plate. To the south beneath the Puna, however, the subducted plate gradually shoals between 24 and 30°S. In this zone of shoaling, there is a notable gap in Wadati-Benioff zone earthquakes between 25 and 27°S (Cahill & Isacks 1992). This gap could be an artifact of the short sampling interval of the instrument record, or it could reflect first order, lithospheric scale processes. Contours of depth to the Wadati-Benioff zone project smoothly across the gap, and ray-path modeling and studies of seismic wave attenuation (Whitman et al 1992) indicate that the subducted Nazca plate is present across this earthquake gap.

Morphology

The availability of regionally consistent topographic data incorporated into digital elevation models has revolutionized the study of modern mountain belts and provides considerable insight into the tectonics of the Central Andes (Figure 2). In this largely arid region, the effects of late Cenozoic tectonics and magmatism on topography have not been obliterated by erosion. Isacks (1988) showed that the average elevation of the plateau between 13 and 29°S is 3.65 km, and he interpreted the 250–300 km wide area of internal drainage in the plateau between 15 and 27°S as evidence of a young age of uplift. The smooth western flank of the Central Andes contrasts strongly with the rough topography on the eastern flank (Isacks 1988). The Puna has an average elevation nearly a kilometer higher than the Altiplano (Figure 3), which has been attributed to greater thinning of the lithosphere beneath the Puna (Whitman et al 1996).

The intimate connection between plate motions, mountain belt topography, and the geometry of the subducted Nazca Plate was demonstrated clearly by Gephart's (1994) analysis of the Isacks topographic data set. He showed that the topography of the Central Andes and the underlying Wadati-Benioff zone is remarkably symmetric about a vertical plane (approximately at the Arica bend) whose pole is oriented about 63°N 113°W. The symmetry axis coincides with the Nazca–South America finite pole of rotation for the period between 36 and 20 Ma; the symmetry plane is closely coincident with the Euler equator

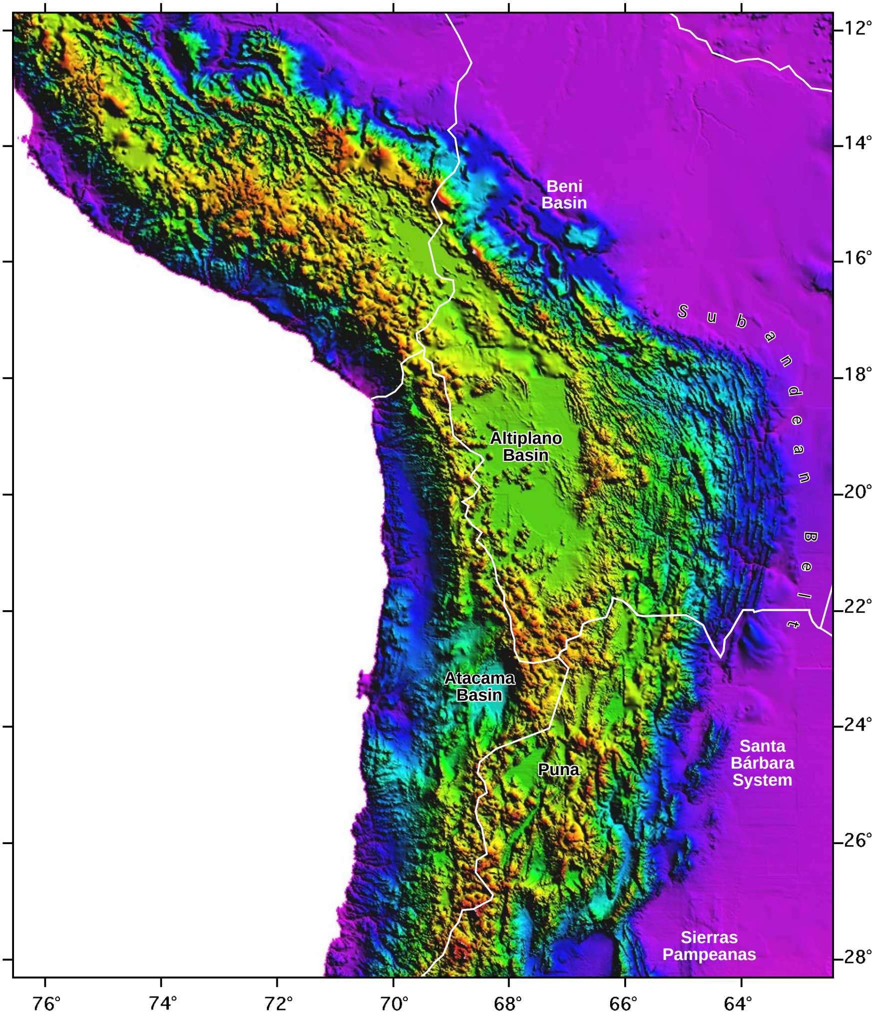


Figure 2 Shaded relief map showing the topography of the Central Andes, based on the 1-km DEM of the Defense Mapping Agency. The Altiplano basin is the extremely flat area in the center of the image between 17 and 21°S. The image highlights the differences between the Altiplano and Puna.

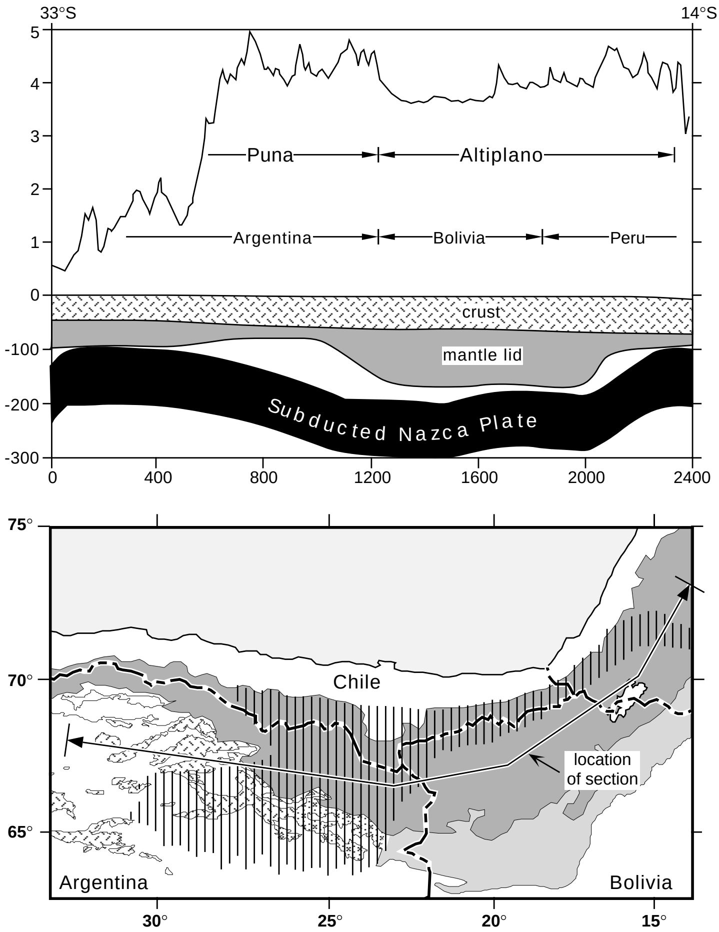


Figure 3 The along-strike variation, in the Central Andes, of lithospheric thickness and corresponding changes in topography, highlighting the differences between the Altiplano and Puna (modified from Whitman et al 1992, 1996). In the cross section at the top, the white area above the "Subducted Nazca Plate" is the asthenosphere beneath South America; the white area beneath it is the asthenosphere and deeper mantle beneath the Nazca Plate. In the map at the bottom, the vertical (east-west) rule overlay shows the area of high seismic-wave attenuation. The other patterns are the same as in Figure 1.

of relative motion since the mid-Tertiary (Gephart 1994). The topographic symmetry exists despite the substantial lateral geologic variations, implying that the continent yields in whatever way necessary to fill a prescribed volume.

The physiography and high rates of orographic precipitation along the northeastern flanks of the Andes in Bolivia and southern Peru indicate high erosion rates, which have removed significant amounts of material from the plateau during the Late Cenozoic (Isacks 1988). In the region of the Beni Basin (Figure 2), as in Himalayan mountain belts, high rates of erosion may dominate the morphology of the active thrust belt (Masek et al 1994). In contrast, along the eastern flank in southern Bolivia, where rates of precipitation and erosion are considerably reduced, the tectonic signal remains dominant in the morphology (Gubbels et al 1993, Masek et al 1994). This difference in morphology and erosion is accompanied by differences in wedge taper and magnitude of shortening in the flanking Subandean thrust belt (see below).

Crustal Thickness, Rheology, and Isostatic Support

Information on crustal thicknesses in the Andes comes from several sources: refraction experiments, broadband passive recording of earthquakes in the subducted plate, and modeling of the gravity field. One of the earliest comprehensive studies of crustal thickness was that of James (1971a), who estimated maximum thickness in excess of 70 km beneath the Western Cordillera based on interpretations of surface waves. Refraction experiments have defined the thickness and velocity structure on the margins of the plateau but commonly do not detect Moho at its deepest point, owing to the highly attenuating nature of the crust (Ocola & Meyer 1972, Wigger et al 1994). Broadband recording of earthquake sources, such as that carried out by the recently completed BANJO (Broad Band Andean Joint) and SEDA (Seismic Exploration of the Deep Altiplano) experiments, minimizes the attenuation problems, thus enabling crustal thickness estimates across the plateau (Beck et al 1996). We show the broadband and refraction results as a new contour map of depth to Moho (Figure 4). Near the triple point where Argentina, Bolivia, and Chile come together, Zandt et al (1994) concluded that the crust could be as thick as 80 km. Most seismic studies have concluded that the average velocity of the crust beneath the Altiplano is low ($V_P \approx 6$ km/s), as is the Poisson's ratio of 0.25 (Beck et al 1996, Wigger et al 1994, Zandt et al 1996), which imply a felsic composition (Zandt et al 1996).

Regional gravity studies have been carried out at the northern margin of the plateau in Peru by Fukao et al (1989) and Kono et al (1989) and in the central and southern plateau area by Götze et al (1994). Modeling of regional gravity measurements between 20 and 26°S resulted in the conclusion that the crust beneath the Altiplano and Puna was less than 70 km thick (Götze et al 1994), in

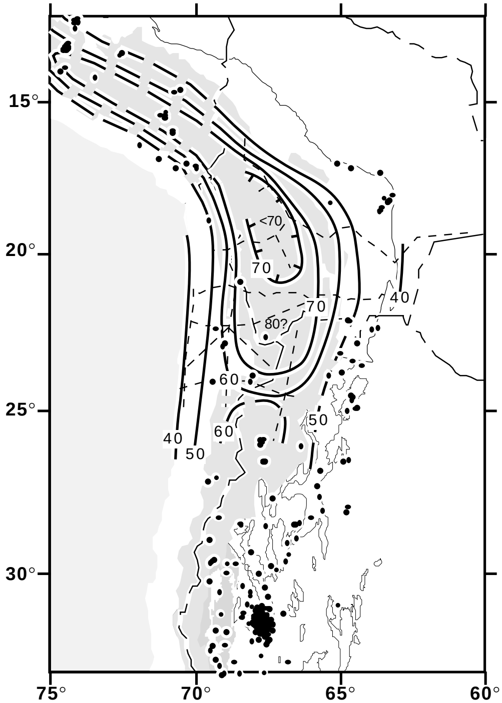


Figure 4 Crustal seismicity at less than 60-km depth in the Central Andes east of the 75-km Wadati-Benioff zone contour. From ISC and PDE catalogs for events during 1964–1988 reported by more than 20 stations. Note the lack of seismicity beneath the plateau, except at its northern and southern ends. Moho contours represent a compilation of data of different types and qualities. Solid contours are based on published interpretations of refraction data (Wigger et al 1994) and broadband data (Beck et al 1996, Zandt et al 1994), with locations of these surveys shown with short dashed lines. Dashed contours in northernmost Bolivia and Peru are from James (1971a), and the dashed 60-km contour in the southern Puna is from Götze et al (1994).

contrast to the refraction and broadband results. The differences in interpreted depth to Moho may indicate that the lower crust is composed of material that is crust by a velocity definition but mantle by density (and petrologic) definitions or vice versa. If so, this could help explain the apparent dearth of shortening necessary to fill the lower crust of the plateau.

The isostatic residual gravity anomaly can be used to assess the state of isostatic compensation in the Central Andes. Several workers have used regional gravity data sets to show that the eastern flank of the Andes in Bolivia is flexurally supported (Lyon-Caen et al 1985), although the degree of support, or flexural rigidity, varies along strike (Watts et al 1995). Whitman et al (1996) showed that south of $\sim$24°, the eastern flank of the Andes is locally, not flexurally, compensated, which has important implications for the style of shortening along the margins of the plateau.

Current State of Stress in the Plateau

The current state of stress in the plateau can be inferred from a number of different observations, including distribution of crustal seismicity, studies of young fault populations, and, locally, the presence of young mafic volcanic rocks. In general, the northern and southern margins of the plateau appear to be the loci of seismic and neotectonic activity—characterized by approximately north-south horizontal extension—whereas the central part is little deformed (Figures 4, 5), which suggests that far-field compression is in balance with the weight of the uplifted plateau (Froidevaux & Isacks 1984, Molnar & Lyon-Caen 1988).

A single earthquake at about 11-km depth has been recorded beneath the southern Puna (Chinn & Isacks 1983). This event, an oblique-thrust mechanism, is probably related to a regionally important fault zone that governed the location of the Antofalla Salar (salt pan); surface features record Quaternary strike-slip along the fault (Allmendinger et al 1989, Marrett et al 1994). Much of the Puna has been dominated for the last 1–2 Ma by strike-slip and extensional faulting, in contrast to a protracted earlier history of thrust faulting (Cladouhos et al 1994, Marrett et al 1994). These faults are commonly associated with and may have acted as conduits for young, volumetrically minor mafic lavas (Allmendinger et al 1989, Fielding 1989, Kay et al 1994a). It is unlikely that these dense magmas could have traversed $\sim$70 km of continental crust under anything but a nearly neutral to extensional stress regime (Marrett & Emerman 1992).

At the northern end of the plateau in southern Peru, crustal seismicity and young faulting also suggest approximately north-south extension (Grange et al 1984, Lavenu 1982, Sébrier et al 1985, Suárez et al 1983). At both the northern and southern ends of the plateau, the horizontal extension is oriented at 55–60° to the local trend of the mountain belt. Because the extension is not perpendicular to the orogen, the margins of the plateau are not collapsing but instead are being deformed by left-lateral strike-slip in the north and right-lateral strike-slip in the south. This could result from "continental escape" but is more likely a kinematic consequence of diminished shortening north and south of the plateau.

Lithospheric Thickness

Based on mapping of seismic wave attenuation beneath the plateau, modeling of seismic wave attenuation (Q) in the mantle, and other geophysical data from across the Puna-Altiplano plateau, Whitman et al (1992, 1996) suggested that the modern lithospheric thickness is roughly 150 km beneath the Altiplano and

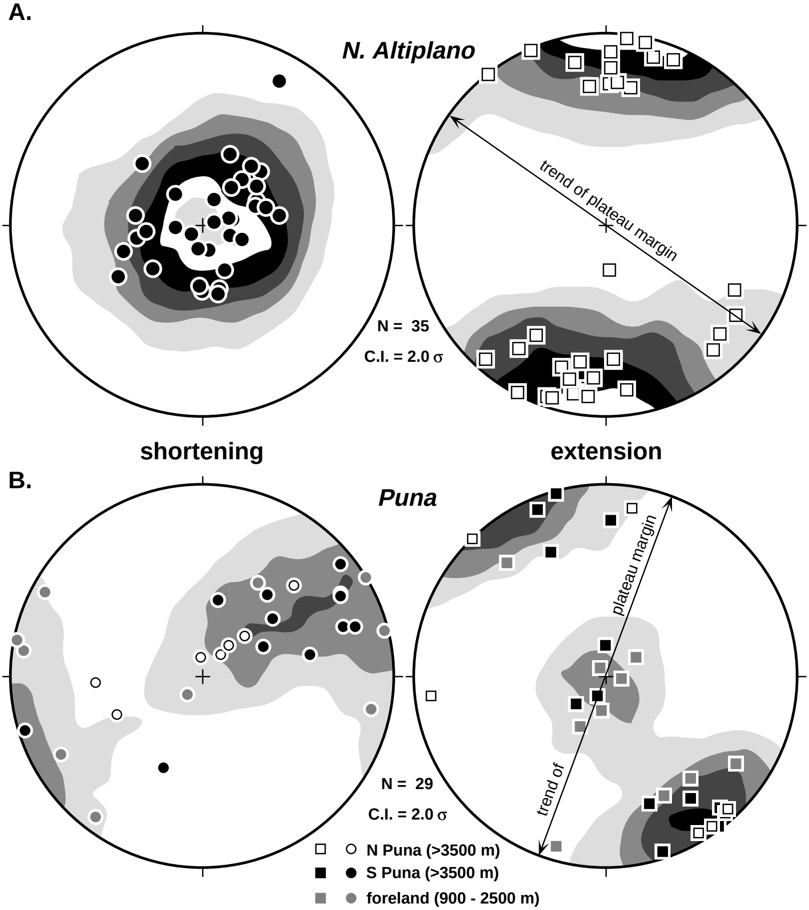


Figure 5 Summary of fault-slip analyses of Quaternary deformation at the northern (Sébrier et al 1985) and southern (Allmendinger et al 1989, Cladouhos et al 1994, Marrett et al 1994) margins of the plateau, plotted as composite P (*dots*) and T (*boxes*) axes. Each dot/box represents anywhere from <10 to >100 individual fault analyses at a particular geographic site. The double-headed arrows show the local trend of the plateau topography at the northern and southern termini. The Puna data are further categorized in terms of elevation to show that horizontal extension is not restricted to high elevations, nor is horizontal shortening restricted to low elevations.

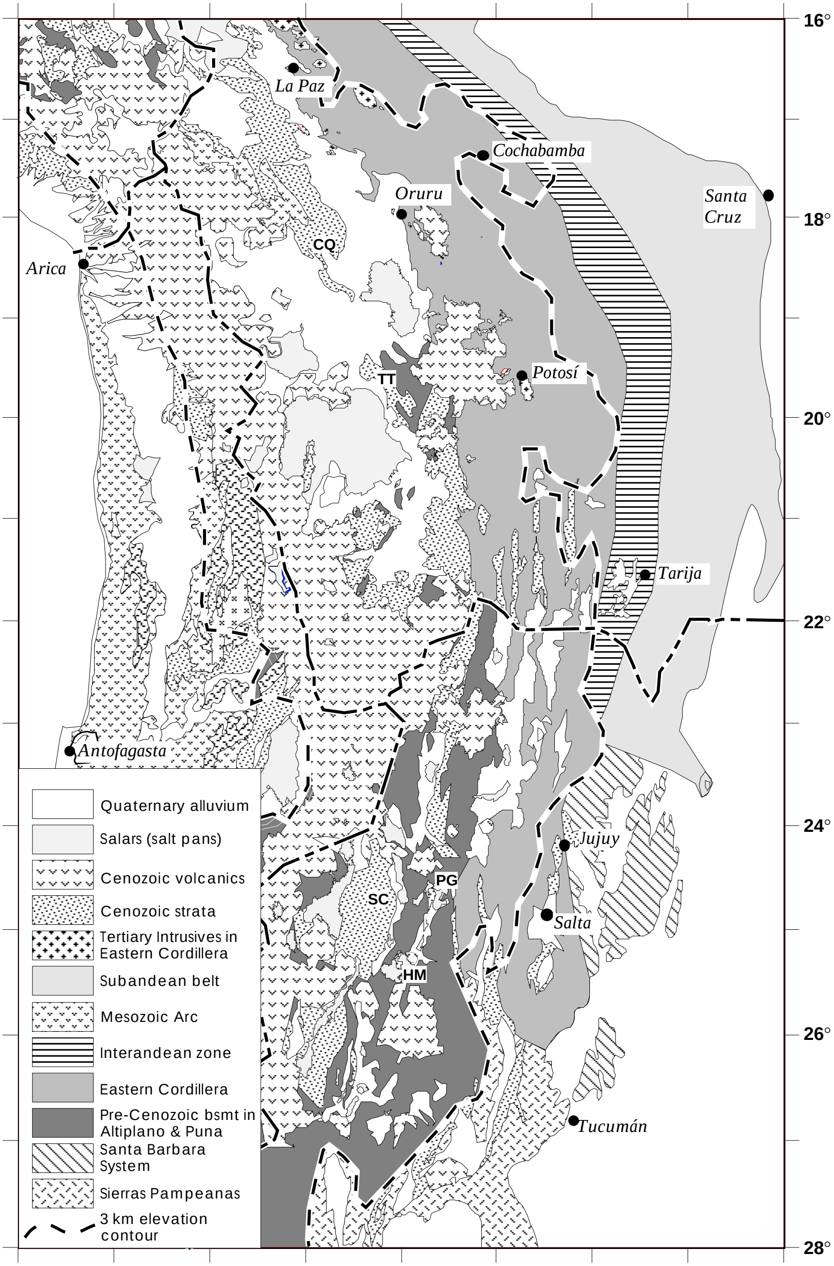
is significantly thinner only in a narrow band beneath the Western Cordillera (Figure 3). Data on the distribution and chemistry of young mafic (shoshonitic) magmas over the same region (Davidson & de Silva 1995, Kay et al 1994a, Soler et al 1992) imply that young mafic back-arc magmas are largely derived from small degrees of melting of enriched continental lithosphere. Extrapolation of He-isotopic data from hot springs farther north (Hoke et al 1994) confirms that mantle magmas are in the lithosphere, but puts few real constraints on lithospheric thicknesses. Farther south, Whitman et al (1992) concluded that the lithosphere was substantially thinned beneath the entire Puna, by as much as 50 km with respect to the Altiplano (Figure 3). This interpretation provides explanation of the higher topography of the Puna and is supported by studies of young mafic magmatism in the southern Puna between 24 and 27°S (Kay et al 1994a, Whitman et al 1996).

TECTONIC OVERVIEW

The high plateau of the Central Andes must be considered within the context of the entire orogen. Here we describe the salient differences between the Altiplano and Puna segments of the Plateau in terms of the structures across two key transects.

Altiplano Transect (North of 22°S)

North of 22°S, a transect of the Andes crosses (from east to west; Figure 6) the down-flexed but otherwise undeformed crust of the Chaco foreland basin, the Subandean thin-skinned fold and thrust belt, the Eastern Cordillera (Cordillera Oriental), the Altiplano, the active magmatic arc in the Western Cordillera (Cordillera Occidental), the Chilean Precordillera, the Longitudinal Valley of northern Chile, the Coastal Cordillera, and the Peru-Chile trench. The Chaco is a foreland basin that stretches 600 km across Central Bolivia to the Precambrian shield in the eastern part of the country. The age of the fill is poorly known but is generally considered to be Neogene and Quaternary. The Subandean belt is a classic thin-skinned fold and thrust belt (Baby et al 1992, Baby et al 1995, Dunn et al 1995, Kley & Reinhardt 1994, Mingramm et al 1979, Roeder 1988, Roeder & Chamberlain 1995, Sheffels 1990). The western limit of the Subandean belt is marked by the "principal frontal thrust" (Cabalgamiento Frontal Principal or CFP). West of the CFP, Silurian rocks are exposed in a narrow belt known as the Interandean zone (Figure 6). Across a complex structural zone called the Principal Andean thrust or Main Andean thrust (Cabalgamiento Andino Principal or CANP), Ordovician and, locally, older rocks in the Eastern Cordillera dominate the outcrop. The eastern limit of the plateau is marked by the high topography of the Eastern Cordillera, the eastern limits of the Late Oligocene


16°
18°
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70°
68°
66°
64°
La Paz
Cochabamba
Oruru
Santa Cruz
Arica
CQ
TT
Potosí
Tarija
Antofagasta
Jujuy
PG
SC
Salta
HM
Tucumán
Quaternary alluvium
Salars (salt pans)
Cenozoic volcanics
Cenozoic strata
Tertiary Intrusives in Eastern Cordillera
Subandean belt
Mesozoic Arc
Interandean zone
Eastern Cordillera
Pre-Cenozoic bsmt in Altiplano & Puna
Santa Barbara System
Sierras Pampeanas
3 km elevation contour

to Late Miocene magmatic arc (Figures 2, 6) and the eastern limits of remnants of the high-level geomorphic surfaces described by Gubbels et al (1993).

The Altiplano surface is covered by several large salars, Quaternary fill, and, locally, Late Oligocene to Recent volcanic rocks, including immense Late Miocene to Pliocene ignimbrite centers at the southern end of the plateau. Sparse exposures of the underlying basement consist of Ordovician and Cretaceous rocks. There are widely divergent opinions about the importance of Cenozoic strike-slip faulting in the Altiplano and Eastern Cordillera (e.g. compare Hérail et al 1996 and Horton 1996).

The Western Cordillera, the modern magmatic front, is marked by a line of stratovolcanoes overlying older ignimbrite sheets. In the Chilean Precordillera, a belt of intense Incaic ($\sim$38 Ma) shortening involves rocks of the early Tertiary and Mesozoic magmatic arc as well as pre-Andean igneous and basement rocks (Scheuber et al 1994). The Longitudinal Valley is a forearc depression filled with Quaternary to Miocene strata, and the Coastal Cordillera is dominated by the Mesozoic Andean magmatic arc. The lack of Mesozoic forearc rocks indicates that considerable tectonic erosion has truncated the leading edge of South America since the Late Jurassic (Rutland 1971, von Huene & Scholl 1991).

Transition from Altiplano to Puna

A major lateral transition occurs along a NW-SE zone, running from 23–24° at the eastern margin of the Andes to 20–21° along the main magmatic arc (Figure 1). A number of fundamental changes occur across this transition zone that are variably thought to reflect Precambrian to Mesozoic paleogeography and changes in subduction zone geometry and lithospheric thicknesses (e.g. Allmendinger & Gubbels 1996, Allmendinger et al 1983, Coira et al 1993, Whitman et al 1996).

East of the modern arc, an early Paleozoic sedimentary wedge overlies an old Precambrian basement north of $\sim$21°. To the south, an early Paleozoic submarine arc and associated back-arc sedimentary sequence were constructed upon Precambrian basement that is younger than that to the north. This paleogeographic change is reflected in the chemistry of Tertiary magmatic rocks and in the restriction of important Ag-Sn deposits to north of 22°S and their complete absence south of 24°S. The most significant north-south structural change in the back arc is the southward termination of the thin-skinned Subandean belt near

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Figure 6 Simplified geologic-tectonic map of the Central Andean plateau in Bolivia and northern Argentina. Shows locations of stratigraphic section discussed in the text and illustrated in Figure 7.

23 and 24°S (Figures 1, 6). This change correlates with a southward pinch-out of the wedge-shaped Paleozoic basins and the superposition of Late Cretaceous rift basins in the foreland south of 24°S (Allmendinger et al 1983). In the arc, a chemical transition in young lavas from 22–20°S has been attributed to thrusting of older basement southward over younger basement (Wörner et al 1992). Between 22.5 and 24°, the arc is displaced eastward by the Atacama basin (Figures 2, 6), an unexplained first-order anomaly in the Andean forearc, which dates at least to the late Paleozoic (Flint et al 1993).

Puna Transect (South of 24°S)

In the foreland east of the Puna, the thick-skinned Santa Bárbara System and the northern Sierras Pampeanas (Figures 1, 6) replace the thin-skinned Subandean belt. Crustal seismicity in these commonly west-verging structures is as deep as 30 km, more than twice as deep as sparse Subandean belt seismicity to the north (Cahill et al 1992, Chinn & Isacks 1983). The Eastern Cordillera is dominated by outcrops of the Precambrian rocks with minor Late Cretaceous deposits. Southward, the Precambrian is increasingly strongly metamorphosed and intruded by Paleozoic and Precambrian plutons (Willner et al 1987). Internally, the Puna is broken up into numerous contractional "basins and ranges," in contrast to the broad flat Altiplano basin (Figures 2, 6). Most of the Puna ranges are composed of Paleozoic rocks that exhibit several phases of deformation. Cutting across the Puna are several northwest-trending fingers of Miocene and younger volcanic centers (Figure 6), which may have been controlled by old, northwest-trending zones of lithospheric weakness (Allmendinger et al 1983, Alonso et al 1984, Coira et al 1982, Salfity et al 1984). One of these fingers terminates at the eastern edge of the Puna in the Cerro Galán caldera, one of the youngest ignimbrite centers in the entire plateau (Sparks et al 1985, Francis et al 1989).

VARIATION IN SHORTENING ALONG THE PLATEAU MARGIN

Subandean Shortening

Shortening in the Subandean belt, measured from the deformation front to the Main Subandean thrust (CFP), varies along strike (Table 1). Displacement is greatest in Bolivia north of the bend at 18°S and averages about 135 km (Baby et al 1995, Roeder & Chamberlain 1995). This amount of shortening is distributed across a narrow belt (~70 km), which results in a steep wedge taper with a topographic slope of 3.5° and a decollement dip of about 4°. This large amount of shortening correlates with the greatest rainfall and the highest erosion rates in the Central Andes (along the Beni basin, Masek et al 1994).

Table 1 Summary of amount of shortening in the Central Andes over the past 100 million years

Location	Shortening (km) A[1]	B[2]	C[3]	D[4]	Wedge taper	Annual precipitation (mm)	Reference
N. Bolivia	74	—	177	191	7°	1600–2400	Baby et al (1996)
	135	—	—	—	7°	1600–2400	Baby et al (1995)
	—	137	—	230	7°	1600–2400	Roeder (1988)
	132	156	—	279	7°	1600–2400	Roeder & Chamberlain (1995)
Bend at 18°S	—	—	210	320?	—	400–1400	Sheffels (1990)
S. Bolivia	33[5]	—	—	—	2.5°	600–1000	Baby et al (1992)
	100	159	—	—	2.5°	600–800	Dunn et al (1995)
	—	140–150	195–230	215–250	3°	800–1200	Kley (1993), Kley et al (1996), Schmitz & Kley (1996)
	—	—	—	320	3°	—	Schmitz (1994)
	86	125	211	230	3°	800–1200	Baby et al (1996)
N. Argentina	60	75	—	—	5°	800–1400	Mingramm (1979), Allmendinger et al (1983)

[1]Subandean belt only (footwall of the CFP).
[2]Subandean + Interandean zone (footwall of CANP).
[3]Subandean + Interandean + Eastern Cordillera.
[4]Total shortening east of the current arc in the western Cordillera.
[5]Shortening estimate includes only the eastern part of the Subandean belt.

In southern Bolivia at 21°S, the Subandean belt shortening is about 100 km back to the CFP (a present width of 100–125 km) (Dunn et al 1995). The decollement dip is shallower (2°W) and the topographic slope is just 0.5–1.0°, resulting in a very gentle wedge taper (Table 1). Precipitation is 2–4 times lower than in the Subandean belt north of 18°S. Shortening diminishes farther south as the Subandean belt dies out in northern Argentina (Allmendinger et al 1983, Mingramm et al 1979).

Shortening in the Eastern Cordillera and Altiplano

The Eastern Cordillera is the site of an important change in vergence, from predominantly east-verging in the Subandean belt and eastern part of the Eastern Cordillera to west-verging structures that form the eastern boundary of the Altiplano (Kley et al 1996, Roeder 1988). The magnitude of Andean shortening within the Eastern Cordillera has proven difficult to determine for several reasons: (*a*) The rocks exposed are a monotonous, featureless sequence of Ordovician strata; (*b*) those strata were deformed prior to the Andean Orogeny (i.e. prior to the Cretaceous); (*c*) the area was uplifted and eroded prior to deposition of Cretaceous rift-related strata; and (*d*) outcrops of Tertiary strata are scarce, which makes the recognition of Andean-age structures difficult.

Nonetheless, crude estimates of total Andean shortening (including both Subandean and Eastern Cordillera) have been made both north and south of the bend at 18°S (Table 1). In general, Eastern Cordillera and Altiplano shortening is thought to be considerably less than Subandean/Interandean shortening. The best constrained estimates (but still based solely on surface geology) come from the Eastern Cordillera in southern Bolivia at about 21°S, where greater preservation of Tertiary strata facilitates the identification of Andean structures and where crustal scale refraction and magnetotelluric data provide some additional constraints (Hérail et al 1992, Hérail et al 1996, Kley et al 1996, Kley & Reinhardt 1994, Schmitz 1994).

South of 24°S, the amount of Andean shortening is yet more poorly constrained, even in the foreland. At 25°30′S, Grier et al (1991) calculated about 70 km of shortening in the Santa Bárbara System and Eastern Cordillera (64–66°15′W longitude), based on surface geology alone. Shortening within the Puna farther west is almost completely unknown. The greater internal relief and more abundant exposure of pre-Cenozoic basement suggest either larger magnitude or younger shortening within the Puna than in the Altiplano.

SEDIMENTARY BASIN VARIATION WITHIN THE HIGH PLATEAU

The history of basin subsidence can provide clues to the timing of deformation, mechanisms of vertical movement, and the emergence and erosion of source areas. The histories and scales of basins in the Puna and Altiplano segments point to different times of deformation and different controls on subsidence in the two areas. Middle and late Cenozoic strata comprise four principal stratigraphic intervals, of which the first two are found in both the Puna and Altiplano, whereas the latter two, of Miocene to Pliocene age, differ between those provinces and suggest a divergence in basin-forming processes.

The most regionally extensive unit is the oldest: a suite of redbeds that reaches a 5-km thickness and spans the late Paleocene through Oligocene (Figure 7, Stage 1) (Alonso et al 1991, Evernden et al 1977, Kennan et al 1995, Pascual et al 1978, Sempere et al 1997, Vandervoort 1993). These units may have accumulated in a foreland basin (Sempere et al 1990a) that is associated first with Incaic deformation west of the basin (∼38 Ma), and later with an early phase of deformation to the east (Kennan et al 1995, Sempere et al 1997).

The second unit is thought to be laterally extensive but is very poorly described; it comprises earliest Miocene red sandstones and mudstones associated with basalts and dacitic tuffs (dates from about 23–21 Ma). These strata may

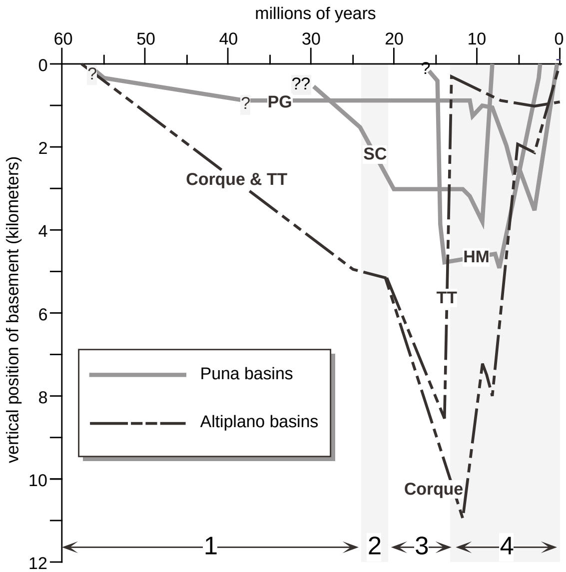


Figure 7 Subsidence histories of sedimentary basins in the central Altiplano (17–20°S) and southern Puna (24–26°S). Line segments inclined down-to-the-right indicate subsidence and sediment accumulation; line segments inclined up-to-the-right indicate uplift and erosion associated with local folding or faulting. Locations of inflections in curves reflect resolution of available data and not necessarily times of true shifts in rates of vertical motion. Zero on the vertical axis is the position of the earth's surface at the time that locally preserved strata began to accumulate. CQ—Corque syncline, TT—Tambo Tambillo, PG—Pastos Grandes, SC—Siete Curvas/Salar de Pocitos, HM—Catal Island in Salar Hombre Muerto.

be less than a kilometer thick, but they are recognized from at least the southern Altiplano to south of the Puna (Hérail et al 1993, Kennan et al 1995, Vandervoort et al 1995). The chemistry of the associated basalts suggests lithospheric extension (Hérail et al 1993, Soler & Jimenez 1993), and thus we speculate that the basin may be of thermal sag origin, with local modifications where preexisting faults were reactivated.

Strata overlying the earliest Miocene basalts and redbeds are spatially quite variable. In the southern Altiplano, thick piles of early and middle Miocene clastics are widespread. In the southern Puna, there was a hiatus until about 15 Ma, when accumulation of strongly evaporitic strata began.

Two depocenters in the Central Altiplano (Corque syncline and Tambo Tambillo, Figure 6) contain 3–6 km of clastics, with only minor gypsum, that span the early and part of the middle Miocene (Kennan et al 1995, MacFadden et al 1995) (Figure 7, Stage 3). The Corque basin must have had a surface area of at least 10,000–20,000 km^2. The structural nature of these basins is not well defined. Because the Eastern Cordillera and easternmost Altiplano was a domain of west-verging thrusting during this time interval (Horton 1996, Kennan et al 1995, Kley et al 1996, Tawackoli et al 1996), the Altiplano may have behaved as a foreland basin (Baby et al 1990, Gubbels et al 1993), which is consistent with the broad area and thickness of the early and middle Miocene strata. Existing rock descriptions suggest that low-gradient streams and shallow lakes were common, but conditions were not appropriate for creating evaporites.

After about 13 Ma, accumulation of strata on the Altiplano was characterized by thinner units (totaling less than 2000 m, Evernden et al 1977) that are spatially variable. Progressive tilting of these units indicates a complicated history of local deformations as well (Figure 7, Stage 4). Whereas partial folding of the Corque syncline area occurred between 15 and 9 Ma, the principal folding occurred between 9 and 5 Ma, and continued after 5 Ma (Kennan et al 1995). In the Tambillo area, principal deformation apparently was before 13 Ma, and units younger than 13 Ma are gently rotated (Kennan et al 1995, MacFadden et al 1995). If the middle-Miocene Altiplano basins formed as foreland basins in response to thrusting in the Eastern Cordillera, the fact that deformation in the Eastern Cordillera had largely ceased before about 10 Ma (Gubbels et al 1993) may explain the apparent demise of basin subsidence in the Altiplano during the late Miocene and Pliocene.

Most of what, in the Altiplano, constitutes the third stratigraphic stage (Figure 7) is apparently an unconformity in the Puna basins. Nevertheless, late Cenozoic sedimentary basins in the southern Puna are noteworthy for their great thicknesses (up to 5 km), small spatial dimensions, economical evaporite concentrations, and continuation of the basin-forming conditions to the present (e.g. Alonso et al 1991, Vandervoort 1993). Because the sections exposed in now-separate valleys are highly diachronous (Figure 7, Stages 3 and 4) (Alonso et al 1991, Vandervoort et al 1992), the strata probably also formed in separate basins. The late Cenozoic basins were an order of magnitude smaller in area than those of the Altiplano (Vandervoort 1993). Ranges flanking these basins were undergoing thrusting and folding prior to and contemporaneous with this middle and late Miocene subsidence. In regions between and adjacent to the basins illustrated in Figure 7 (Pastos Grandes, Hombre Muerto, and Siete Curvas), Marrett et al (1994) showed that thrusting and folding were active during the middle Miocene, as well as continuing through the late Miocene and Pliocene.

The spatially discontinuous and diachronous deposits of the Puna basins, and their small size, suggest that subsidence was controlled by local structural relationships. The most plausible basin-forming mechanisms are thought to be block rotation in footwalls of range-bounding reverse faults and synclinal subsidence, with perhaps little or no flexural subsidence. In addition, drainage blockage caused by volcanic activity and anticlinal growth contributed to ponding and sediment accumulation.

CHANGING MAGMATIC PATTERNS ACROSS THE PLATEAU

The distribution of magmatic centers records where mantle-derived magmas were added, which led to new crustal growth and contributed to crustal thickening; records where crustal melting occurred, which reflects high crustal geothermal gradients; and provides clues as to how crustal and lithospheric thickness varied in space and time. The three principal types of centers, and the implications of their temporal and spatial distributions, are briefly summarized below.

Stratovolcano complexes constitute the main volcanic chains. These complexes are composed of thick sequences of andesitic to dacitic lavas associated with pyroclastic flows, dacitic to rhyodacitic domes, and hot avalanche deposits. Some have had huge catastrophic debris avalanches (up to 100 km^2) caused by partial collapse of the central edifice (Francis et al 1985). Eruption at high elevations, 5000- to 17,000-m high eruptive columns, and west-to-east stratospheric winds with speeds over 150 km/hr have strongly influenced the distribution of airfall deposits. Coarse grained, proximal pyroclastic fall deposits occur near the centers; intermediate distance deposits are sparse; and fine-grained distal deposits are concentrated in the Eastern Cordillera, Subandean belt, and Chaco Plain (see Glase et al 1989). The Puna Ojos del Salado (27.1°S, 6887 m high) and Lullaillaco (24.7°S, 6723 m) complexes are the highest active volcanic centers on Earth.

Caldera complexes erupted the voluminous silicic andesitic to dacitic (63–68% SiO_2) back-arc ignimbrite sheets that are the dominant late Miocene–Pliocene volcanic deposits on the plateau. These deposits cover more than 500,000 km^2, which makes the plateau the largest young ignimbrite province on Earth. Most of these eruptions occurred from huge calderas parallel to the main arc or in the transverse volcanic chains that cross the plateau (Figure 6). Their size is such that many of their vents were only recognized after the advent of satellite imagery (e.g. Baker 1981, Gardeweg & Ramírez 1987, Ort 1993, Sparks et al 1985). The large phenocryst-rich (up to 40–50% crystals) ignimbrites with relatively homogeneous compositions probably erupted from homogeneous magma chambers. Smaller ignimbrites, with more variable compositions, are considered to have erupted from smaller, zoned magma chambers (de Silva 1991, Hawkesworth et al 1982). In general, the ignimbrites probably resulted from massive amounts of crustal melting induced by introduction of mantle-derived magmas into the thickened crust (Coira et al 1993, de Silva 1989, Francis et al 1989).

Small back-arc mafic monogenetic cones and fissure flows are primarily of latest Oligocene to early Miocene or latest Miocene to Recent age. These basaltic to mafic andesitic flows are dominantly mantle-derived. Young centers are most voluminous in the southern Puna where they are generally associated with extensional or strike-slip NNW-SSE, NE-SW, and N-S trending faults (see Marrett et al 1994). Some of these flows have calc-alkaline and others intraplate-like chemistry, whereas smaller flows in the northern Puna and Altiplano have shoshonitic chemistry (see summary in Kay et al 1994a).

The spatial and temporal distribution of these Puna magmatic types are discussed with respect to four time windows shown in Figure 8. As described in a following section (Evolution of the Central Andean Lithosphere), the pattern of magmatism is consistent with a subducting slab beneath the northern Puna-southern Altiplano that steepened through time, flanked by progressively shallowing segments to the north and south (Coira et al 1993, Kay et al 1995).

Activity during the late Oligocene to early Miocene (26–16.5 Ma, Figure 8*a*) was concentrated from 24–21 Ma, with a relative lull occurring from 20–16 Ma. Centers to the south of 25°S are mostly restricted to the Western Cordillera, whereas those near 25–24°S and north of 22°S extend across the plateau into the Eastern Cordillera. A virtual gap in magmatism occurred from about 22–24°S. Subsequent volcanic sequences overlie the "Chayanta erosional surface" (Sempere et al 1990b), an unconformity that extends from the Puna into the Altiplano.

During the middle Miocene (about 16–12 Ma, Figure 8*b*), a number of important changes took place in the back arc as local andesitic to dacitic eruptions from long-lived centers spread across the southern Puna, and small stocks and extrusive domes in the northern Puna began to erupt at about 13 Ma in the region of the magmatic gap (see summary in Coira et al 1993). North of 22°S, back-arc eruptions from major centers with ages from 16–12 Ma continued on the eastern margin of the Altiplano (Coira et al 1993, Richter et al 1992, Soler & Jimenez 1993). Back-arc mafic volcanism ceased across the entire region. In the Western Cordillera arc, stratovolcanic complexes continued to erupt south of 25°S (Kay et al 1994b, Mpodozis et al 1995, Naranjo & Cornejo 1992), whereas ignimbritic eruptions dominated north of 21°S (Baker 1981, Jordan & Gardeweg 1989). The extension of magmatism and basin formation into the Puna back arc is consistent with important plateau uplift in the south just before and during this time.

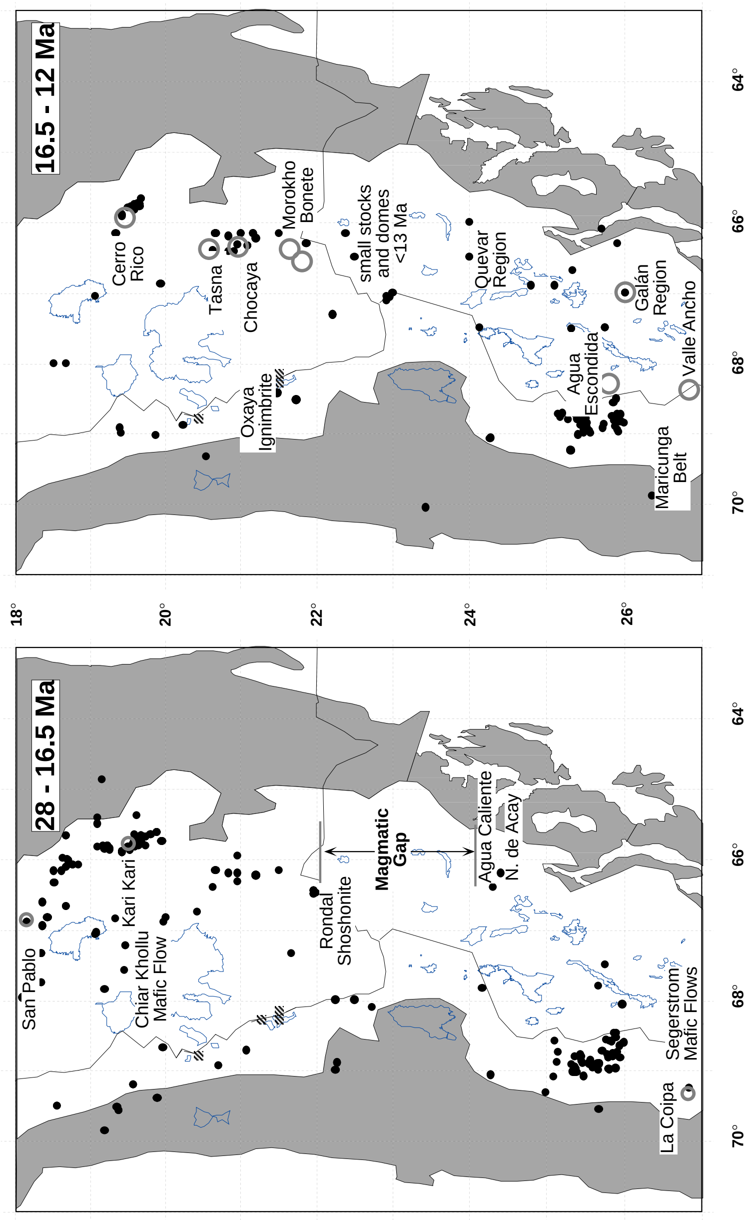
The late Miocene (12–5 Ma) marks the onset of an intense and voluminous period of ignimbritic eruptions that lasted until the late Pliocene (3–2 Ma) (Figure 8*c*). Huge ignimbrites erupted from centers just behind the arc front and along the transverse NW-SE–trending chains crossing the plateau. Northern Puna-Altiplano back-arc flows overlie the widely recognized San Juan de Oro surface, which postdates Miocene deformation in the eastern part of the plateau (Sempere et al 1990b). Particularly spectacular are the gigantic centers between 21.5 and 23°S that extended across the plateau over the early-Miocene volcanically quiescent region (Coira et al 1993, de Silva 1989, de Silva & Francis 1991, Mobarec & Heuschmidt 1994, Ort 1993, Seggiaro 1994). De Silva (1989) assigned these centers to the so-called Altiplano-Puna Volcanic Complex (APVC). Kay et al (1995) have suggested that the eruption of these centers correlate with a marked steepening event of the subduction zone in the northern Puna and southern Altiplano, analogous to the "ignimbrite flare-up" of the western United States (Dickinson & Snyder 1978). Magmatic addition associated with such intense volcanism in this region could help explain the extreme crustal thicknesses implied by the geophysical studies of Zandt et al (1994, 1996). Giant late Miocene–Pliocene ignimbrites also erupted outside of the APVC. Most important were the 8–6.5 Ma eruptions from the eastern Altiplano–western Cordillera Oriental and early eruptions of the Cerro Galán caldera (Sparks et al 1985) in the southern Puna back arc near 26°S. Back-arc stratovolcanic-caldera complexes also erupted during this time (Coira et al 1993).

The youngest period of plateau magmatism (0–3 Ma, Figure 8*d*) is dominated by andesitic to dacitic composite stratovolcanic-dome complexes and minor rhyodacitic ash-flow tuffs in the Western Cordillera arc (active centers catalogued by de Silva & Francis 1991), as well as small mafic monogenetic cones and fissure flows in the back arc. The largest of the mafic flows, which have intraplate-like chemistry, are concentrated above the modern seismic gap in the down-going slab, whereas intermediate-size high-K calc-alkaline flows principally occur between 26 and 27°S and from about 25–23°S. Small shoshonitic flows occur near the El Toro lineament at 24°S (Déruelle 1991, Kay et al 1994a, Knox et al 1989) and in the Altiplano (Davidson & de Silva 1995, Soler et al 1992). The only major back-arc Quaternary stratovolcano is the dacitic to basaltic andesitic Cerro Tuzgle in the easternmost Puna at 24°S (Coira & Kay 1993), and the only major ignimbrite is the 1000-km^3 late Pliocene Galán Ignimbrite in the southern Puna (Sparks et al 1985). The volume of Pleistocene-Quaternary volcanic material is much less than that of the late Miocene–Pliocene centers.

BALANCE OF MAGMATIC ADDITION AND CRUSTAL SHORTENING

Role of Magmatism in Crustal Thickening

Early models for explaining crustal thicknesses in the Central Andes appealed to subduction-related subcrustally derived magmas as the principal cause (Reymer


16.5 - 12 Ma
Cerro Rico
Tasna
Chocaya
Morokho
Bonete
small stocks and domes <13 Ma
Quevar Region
Galán Region
Valle Ancho
Agua Escondida
Oxaya Ignimbrite
Maricunga Belt
28 - 16.5 Ma
San Pablo
Kari Kari
Chiar Khollu Mafic Flow
Rondal Shoshonite
Magmatic Gap
Agua Caliente
N. de Acay
Segerstrom Mafic Flows
La Coipa
18°
20°
22°
24°
26°
64°
66°
68°
70°

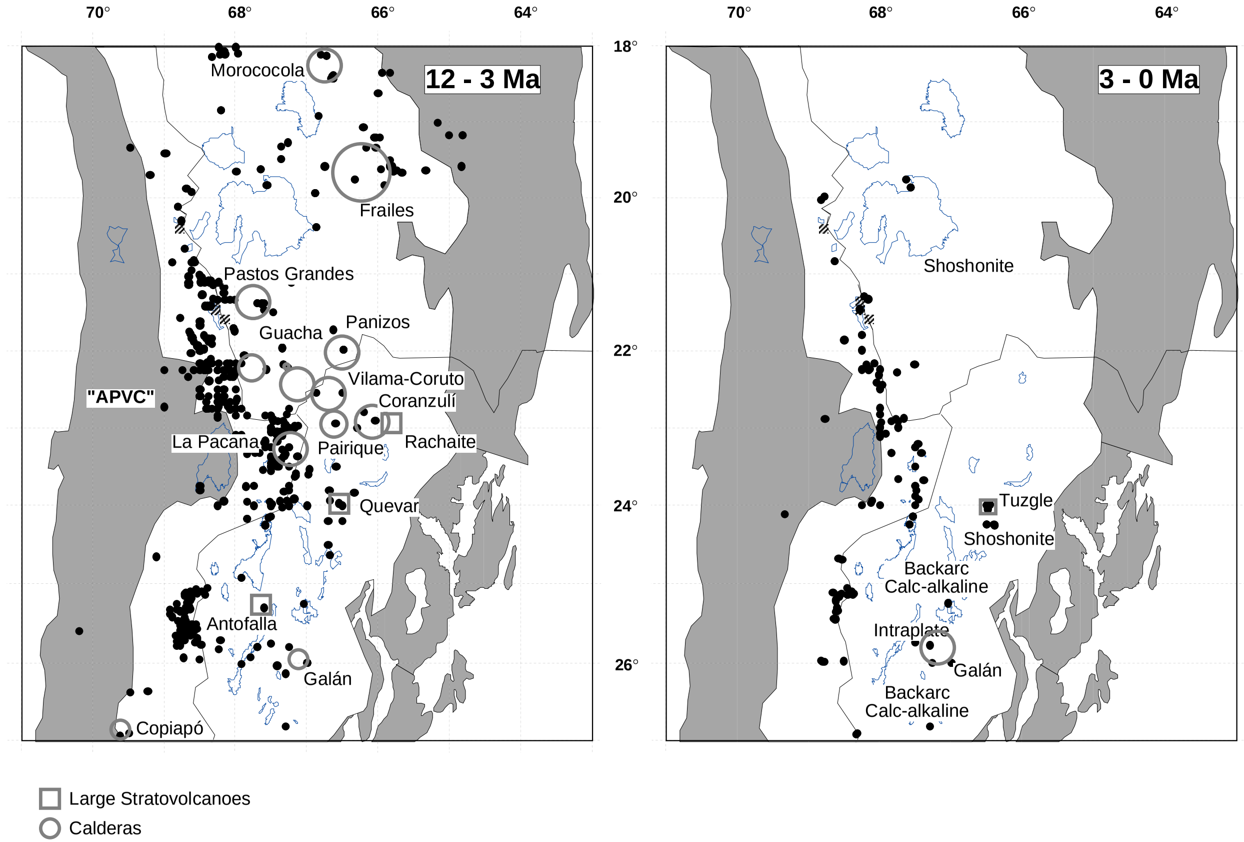


Figure 8 Maps of the Bolivian Altiplano and Argentine Puna plateau showing the distribution of dated magmatic rocks in the four time intervals discussed in the text. *Dots* show locations of many but not all dated centers. Locations of major and representative smaller caldera systems (symbol size indicates relative volume), regions of mafic volcanism labeled with type, and some important stratovolanic centers are shown. Principal references to centers are cited in text and can particularly be found in Coira et al 1993 (includes most pre-1993 references), Kay et al (1994a, 1994b), Soler & Jimenz (1993), and Richter et al (1995). "APVC" refers to Altiplano-Puna Volcanic Complex of de Silva (1989). A complete reference set can be found at ⟨http://www.geo.cornell.edu/geology/cap/CAP_WWW.html⟩.

& Schubert 1984, Thorpe et al 1981). Subsequent appraisals of magmatic addition rates above subduction zones and estimates of crustal volumes in the Central Andes show the difficulty with this model. The principal problem is that average magmatic addition rates above Mesozoic to Recent arcs, which are on the order of 20–40 km^3/km/my (see Reymer & Schubert 1984), cannot produce the crustal volume required unless Andean crustal thickening has been ongoing since the Jurassic. This time frame is contrary to the common view that crustal thickening and uplift of the Andean plateau has occurred in the last 15–25 Ma (e.g. Isacks 1988). The addition rate would need to be $\sim$500–800 km^3/km/my for thickening to occur in 15 million years.

The problem with pure magmatic thickening models is further emphasized by calculating the volumes of Central Andean late Oligocene to Recent magmas by comparing volcanic rock distributions on satellite images with topographic data (Isacks 1988, Isacks et al 1986). Important observations are that (*a*) volcanic rocks are spread above the deformed, eroded, and beveled plateau surface; (*b*) almost all high peaks are volcanic edifices; and (*c*) a major break at 3.65 km in a hypsometric plot represents the elevation of the plateau surface upon which the volcanics are superimposed. Making the assumption that all material above 3.65 km is from new volcanic addition, the added volume is 340,000 km^3. Spreading this material over a generalized plateau 2000 km long and 400 km wide gives only 0.2 km of added crustal thickness. Given that crustal-thickness increases during this time are considered to be on the order of 20 km, even large errors make little difference. The big unknown is the intrusive/extrusive ratio. Even a very generous and almost surely unreasonable 1:40 ratio yields only 8 km of thickness. Francis & Hawkesworth (1994) reached a similar conclusion by summing volumes of individual Central Andean centers (particularly between 21 and 22°S). They concluded that magmatic addition over the last 15 million years could account for only 1.5% of the required crustal volume. A further problem is that geochemical studies show that plateau magmatic rocks contain remelted crust, so that not all erupted material is new crust. A rule of thumb is that an equal amount of new mantle material must be added for each volume of melted crust.

The existing data essentially rule out models that attribute crustal thickening and plateau uplift largely to magmatic addition. To appraise the real role of magmatic thickening, better constraints on starting conditions, the time frame of thickening, and intrusive/extrusive ratios are needed.

Even if magmatic addition is relatively minor in accounting for crustal thickening, magmatism is still fundamental to understanding the thickening process, as magmas and the heat they transport exert a fundamental control on rheology and the mechanical behavior of the crust. Magmas may also have a more local control on the level, extent, and yielding along midcrustal decollements beneath the plateau (Hollister & Crawford 1986). Whereas rheological control may be important across the plateau, magmatic addition has certainly been more important in the frontal arc and along the major volcanic back-arc chains than across the plateau in general.

Crustal Shortening and Thickening

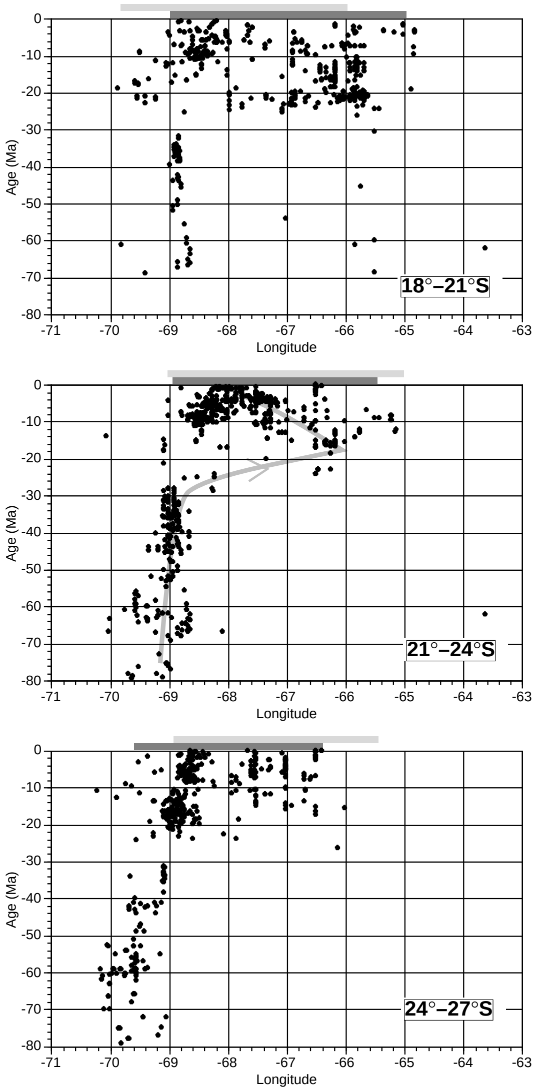
Crustal shortening (Table 1) is the single most important mechanism for thickening the crust of the Central Andes (Isacks 1988, Roeder 1988, Sheffels 1990). However, the known shortening may not be adequate to fill the cross-sectional area. Roeder (1988) suggested that 10% of the area on his line of section could be filled by magmatic addition or other mechanisms. Sheffels' (1990) known shortening accounted for only two thirds of the total area, but she stated that the unaccounted-for shortening in the Altiplano could well fill the remaining area. At 21°S, both Schmitz (1994) and Baby et al (1996) also conclude that shortening is insufficient. Schmitz suggested that 20% of the area must be accounted for by such processes as underplating of tectonically eroded forearc material and magmatic addition.

The key unknowns that must be addressed by future studies are, in order of increasing uncertainty, (*a*) amount and geometry of shortening within and beneath the Eastern Cordillera; (*b*) the magnitude of Cenozoic shortening, particularly in and beneath the Altiplano Basin and on the western slope of the Andes (i.e. Incaic shortening); (*c*) the role of Cenozoic strike-slip faulting and three-dimensional flow of the lower crust in voiding the assumptions inherent in two-dimensional balancing; and (*d*) the initial crustal thicknesses and the time of initiation of thickening.

Initial Conditions and Timing of Uplift

To evaluate the contributions of magmatism and shortening to the crust of the Altiplano, there must be an accurate description of the crustal thickness not only today, but at a time in the past that serves as an initial condition. One reasonable choice of the initial condition is the late Oligocene (~25 Ma) reorganization of plates in the Pacific region. This is a sensible choice because the volcanic arc magmatism spread markedly to the east, to span the area that is now the high plateau, at that time (Figure 9). However, it is an imperfect choice because the crust over at least part of the region was already somewhat thickened by thrusting (see below). At the southern end of the plateau, the chemistry of late Oligocene–early Miocene lavas suggests that the crust was ~50 km thick beneath what is now the western edge of the plateau (Kay et al 1994a). But data elsewhere in the plateau are inadequate for comparable interpretations.

The timing of uplift of the plateau is known only indirectly. Ideally, one would like to have paleo-altimetry data, but those do not exist. The times of



18°–21°S
21°–24°S
24°–27°S
Age (Ma)
Longitude
0
-10
-20
-30
-40
-50
-60
-70
-80
-71
-70
-69
-68
-67
-66
-65
-64
-63

annealing of fission tracks in apatite and zircon have been used to infer times of denudation and uplift but are subject to numerous uncertainties (Benjamin et al 1987, Crough 1983). Masek et al (1994) reanalyzed Benjamin and coworkers' results to show increased denudation rates after 10–15 Ma.

Two substitute approaches have been used. One possibility is to determine the age(s) at which the internal drainage of the plateau was established. Vandervoort et al (1995) showed that the age of initiation of internal drainage in the southern Puna, under climate conditions similar to today's, was approximately 15 Ma, a time that corresponds relatively well with the magmatic history. The sedimentary units of the Altiplano basins have not yet been scrutinized for evidence of the drainage history, although the western slope of the Andes has received more attention (Guest 1969, Hollingworth & Rutland 1968, Mortimer 1973, Mpodozis et al 1995, Tosdal et al 1984).

Alternatively, one can assume that crustal thickening is the primary cause of uplift and that crustal shortening and/or magmatism caused the thickening, in which case a determination of the ages of shortening and magmatic activity provides the temporal history of uplift. Given the conclusion that magmatic contribution must be a minor component of thickening, one naturally focuses on the considerable shallow crustal shortening in the Eastern Cordillera and Subandean belt.

In general, thrusting and crustal shortening in what is now the high plateau progressed from west to east (Kley et al 1996, Sempere et al 1990b). In Paleocene to early Oligocene time (∼60–30 Ma), the region east of today's magmatic arc functioned mostly as the foreland basin to a zone of shortening in Chile (Sempere et al 1997), but some shortening occurred in the Eastern Cordillera and easternmost Altiplano (Kennan et al 1995, Sempere et al 1997). The locus of deformation shifted strongly eastward beginning at about 27 Ma (Marshall & Sempere 1991, Marshall et al 1993, Sempere et al 1990b).

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Figure 9 Geochronology vs longitude plots for different latitudinal swaths across the high plateau of the Central Andes. Most ages are for volcanic and intrusive igneous rocks; tuffs in sedimentary sequences are not shown. The vast majority of ages were determined by the K/Ar or Ar/Ar method, although ages determined with other methods are also included. The gray line and arrowhead highlight the eastward sweep of magmatism across the plateau during the Miocene and subsequent retreat of the arc to its current position in the Western Cordillera. All three graphs show that Miocene and younger magmatism is spatially coincident with the current aerial extent of the plateau. Comparison of the top (Altiplano) and bottom (Puna) graphs shows that magmatism spread across the Altiplano at 25 Ma but did not spread across the Puna until 15–20 Ma; by inference, the Altiplano was uplifted before the Puna. See text for discussion. For references, see URL given in Figure 8 caption.

The long-term kinematic history of the Eastern Cordillera is relatively well known for the period ∼27–8 Ma. Study of several extensive and largely intact Tertiary basins preserved in the interior of the belt indicate that these basins developed in response to both forelandward and hinterlandward thrusting in late Oligocene to late Miocene time (Hérail et al 1996, Horton 1996, Tawackoli et al 1996). The end of deformation in the Eastern Cordillera is generally placed at ∼9–10 Ma, on the basis of the distribution and undeformed nature of the San Juan del Oro erosional surface and related local deposits (Gubbels et al 1993). Comparable knowledge of deformation in the Interandean Zone is lacking.

Although the thrust front is widely interpreted not to have entered the Subandean zone until after 10 Ma, and perhaps 6 Ma (Baby 1995, Baby et al 1990, Gubbels et al 1993, Kley et al 1996, Moretti et al 1996, Sempere et al 1990b), this would imply that Subandean thrusting could not have contributed to thickening the Altiplano crust until the latest Miocene. These conclusions, based on very sparse chronological data from the foreland basin units, are called into question by new extensive chronological data for the Subandean belt near the Bolivia-Argentina border: Hernández et al (1996) suggest that (*a*) 16–8.5 Ma foreland basin strata predate local deformation and (*b*) units spanning 8.5–0 Ma accumulated between growing neighboring anticlines.

In summary, Eastern Cordillera shortening apparently thickened the high plateau crust throughout the early and middle Miocene (∼24–10 Ma). Thrusting in the Subandean belt contributed to thickening throughout the time from the late Miocene to the present (since ∼9 Ma). Thus, uplift of Altiplano segment of the high plateau may have been progressive through the Neogene. In contrast, shortening did not begin until 15–20 Ma in the Puna segment, and it continued until the late Pliocene (1–2 Ma).

EVOLUTION OF THE CENTRAL ANDEAN LITHOSPHERE

The one-to-one spatial correlation of Neogene magmatism and the current extent of the 3-km elevation contour (Figures 8, 9) suggests that the lithosphere has been thermally softened. Because this correlation is true of both the Altiplano and Puna segments—despite their differing basements, timing, and modes of shortening—we interpret that to indicate that lithospheric softening has been a key condition for plateau development in the Andes.

Prior models

The spread of magmatism across the plateau at 25 Ma (or earlier) in the central and northern Altiplano has been linked by several workers to shallowing of the angle of subduction of the Nazca Plate beneath the Central Andes (Coira et al 1993, Isacks 1988, Kay et al 1995, Pilger 1981, Pilger 1984). Pilger (1984) related this shallowing to the impingement and subduction of the Juan Fernandez Ridge, noting that the southward migration of the ridge along the margin correlates with a space-time gap in magmatic activity. However, the Juan Fernandez Ridge did not begin to subduct until after 20 Ma, whereas magmatism spread across the central Altiplano about 5 million years earlier.

More importantly, as pointed out by Pilger (1984) and numerous subsequent workers (e.g. Pardo-Casas & Molnar 1987, Scheuber et al 1994), 26–27 Ma is the time of marked increase in trench–normal convergence rate, perhaps producing a lower angle of subduction as a result of overriding of the subducted plate by the leading edge of South America. Given the restriction of the currently active magmatic arc to the Western Cordillera, the angle of early mid-Miocene subduction was probably shallower than it is today.

Isacks (1988) argued that the physiography of the plateau, in combination with available data on the late Cenozoic structural and magmatic history, supported a two-stage model for uplift of the plateau by crustal shortening and thickening. An initial stage of shortening distributed across the width of the plateau was replaced by the current system of shortening, in which the foreland underthrusts the plateau and continued shortening and thickening are confined to the lower crust beneath the plateau. The surface of the plateau has uplifted in the second stage as a relatively low relief, internally drained, and little deformed geomorphic "surface." The two-stage model remains viable for the Altiplano (Gubbels et al 1993); Stage 1 appears to have begun at $\sim$25 Ma and ended around 10 Ma. The structural history for the Puna has been found to be more complex; Stage 1 started between 15 and 20 Ma but has continued locally to 1–2 Ma, and there is little evidence for underthrusting of South American craton beneath the Puna (Allmendinger & Gubbels 1996, Whitman et al 1996).

Isacks (1988) suggested that the topographic data could be explained by crustal thickening due to shortening, combined with the thermal uplift that corresponds to lithospheric thinning by about 70 km. Though the rugged eastern flanks of the plateau reflect the ongoing crustal-scale faulting, the smooth western flank of the Central Andes was interpreted as an upper crustal "monocline" responding to the western boundary of lower crustal thickening beneath the plateau. This feature would also mark the western boundary of lithospheric heating and thinning that is coincident with the western tip of the asthenospheric wedge beneath the South American plate (Isacks 1988).

Recent Modifications

By analogy with the modern Andean setting, the lack of magmatic rocks between 17 and 28 Ma age in the region between 22 and 24°S (Figure 8) has been interpreted as evidence for an episode of flat subduction (Coira et al 1993, Kay et al 1995). If correct, the magmatic centers to the north of the gap (which spread to the eastern edge of the current plateau) overlay a shallowly dipping segment of the subducted plate; to the south of the gap, magmatic centers remain restricted to the Western Cordillera, indicating steeper subduction. The advent of volcanism in the northern Puna in the 16.5- to 12-Ma time frame is consistent with steepening of the subducting slab in this region, whereas the eastward spread of magmatism farther south in the Puna is consistent with shallowing in that region.

Late Pliocene–Recent plateau magmatism can be explained by modern plate geometry and lithospheric thickness. The frontal arc stratovolcanic complexes correlate with a 30°–east-dipping subduction zone beneath the plateau. The concentration of the more voluminous intraplate-like and calc-alkaline back-arc mafic flows in the southern Puna and the small-volume shoshonitic flows in the northern Puna and Altiplano is consistent with geophysical evidence for a thinner lithosphere beneath the southern Puna than under the Altiplano (Kay et al 1994a, Whitman et al 1996). Kay et al (1994a) suggested that the southern Puna lithosphere was thinned during a late Pliocene episode of lithospheric delamination, triggered by instability of over-thickened dense continental crust (Kay & Kay 1993). In contrast, the lithosphere beneath the Altiplano and northern Puna would have been thickened in the late Miocene in association with underthrusting of the Brazilian shield (Gubbels et al 1993) and steepening of the subduction zone, which would lead to a virtual cessation of back-arc magmatism (Kay et al 1995).

CONCLUSIONS

Although the first-order morphologic characteristics of the Central Andean plateau span the Altiplano and Puna segments, their evolutionary paths to their present states differed. The timing of deformation, sedimentary basin subsidence, and age distribution patterns of Cenozoic magmatism suggest that the Central Altiplano region began its principal phase of uplift about 25 Ma, although some uplift could have begun as early as the Eocene (53–34 Ma). The Puna segment of the Central Andean Plateau probably began to rise somewhat later, between 15 and 20 Ma. Differentiation of the plateau as a tectonic unit was made possible by thermal softening of the lithosphere due to high convergence rate and relatively low-angle subduction (Stage 1 of Isacks 1988). The difference in timing between the Altiplano and Puna must reflect the late Cenozoic history of subduction, but it also correlates with first-order differences in the lithospheric character of the two regions. These differences have resulted in contrasting styles and timing of shortening within and along the flanks of the plateau, as well as magmatic variations. Shortening is clearly responsible for the majority of crustal thickening during the time of uplift of the plateau. However, a not-insignificant minority of thickening (10–30%) must be due either to shortening on as-yet-unrecognized structures, incorrect assessment of initial crustal thickness, magmatic addition, conversion of upper mantle rocks to lower crustal velocities by hydration processes, or local tectonic underplating. In addition to crustal thickening, some of the current topography is supported by lithospheric thinning.

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