PERGAMON

Journal of South American Earth Sciences 15 (2002) 59–78

Journal of
South American
Earth Sciences

www.elsevier.com/locate/jsames

The Pampean flat-slab of the Central Andes

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Received 1 November 2001; accepted 1 November 2001

Abstract

Late Cenozoic Andean deformation in the Pampean segment of Argentina and Chile (27°00′–33°30′S) provides an exceptional opportunity to study the orogenic effects of flat subduction in an active convergent margin. At these latitudes, constraints derived from oceanic geomorphic structures indicate that collision of the Juan Fernandez Ridge in the Nazca plate with the Chilean trench propagated from north to south from ca. 18 to ca. 11 Ma. In this region, Cenozoic tectonics have resulted in the development of the Principal and Frontal Cordilleras, the Cuyo Precordillera, and the associated Sierras Pampeanas in the eastward foreland region.

The analyses of different structural transects across the Sierras Pampeanas between 27 and 33°S latitudes show a southward diachronic beginning of basement block uplift from 7.6 to 6 Ma in the Sierra de Aconquija (27°S), 4.5 to 4.19 Ma at Sierra de Famatina (29°S), 5.5 to 4.7 Ma at Sierra de Pocho (31°S), and ca. 2.6 Ma at Sierra de San Luis (33°S). Uplift times are constrained by the age of synorogenic deposits, fission-track data on apatite, and hydrogen isotopes in alunite, where available. These time constraints are compared with the migration and expansion of the magmatic activity into the foreland. The relation of these two data sets shows a striking coincidence, after a residence time of approximately 4–2.6 Ma, between magmatic activity and failure of the crust that resulted in basement block uplift. Thermal weakening of the crust associated with eastward migration of arc magmatism acted to elevate brittle–ductile subsurface décollments, thus leading to thick-skinned basement uplift of the Sierras Pampeanas.

The evolution of the Sierras Pampeanas is linked to that of the main Andes at 33°S latitude. The main deformation phases and uplift of the thin- and thick-skinned fold and thrust belts of Principal Cordillera occurred between 20 and 8.6 Ma. Arc volcanism migrated eastward at this latitude between 16 and 15.8 Ma. This expansion was accompanied by deformation and uplift with the development of a foreland basin. Soon after the shallowing of the subduction zone, orogenic shortening and basin formation were followed by a migration of the volcanic front to the foreland. Propagation rates of the thrust front from 2.5 mm/a in the early stage (15–9 Ma) accelerated to 13.3 and 13.7 mm/a during the middle (9–6 Ma) and late (5–2 Ma) stages of shallowing. Uplift of the Frontal Cordillera overlapped the last stages of Principal Cordillera deformation at 9 Ma. This time is also constrained by magnetostratigraphic studies of the foreland deposits that record a rapid increase in subsidence rates at the same time that arc magmatism expanded to the Sierras Pampeanas, between 9.5 and 6.4 Ma. This rapid propagation of the thrust front and migration of the arc magmatism occurred 2 m.yr later than the 11 Ma collision of the Juan Fernandez Ridge at this latitude. Uplift of the Precordillera by tectonic inversion of pre-existing normal faults and basement uplifts of the Sierras Pampeanas in the vicinity of 33°S occurred from 2.6 Ma to the present. Neotectonic studies, as well as GPS measurements, indicate active deformation in the thrust front of the Precordillera, as well as in the Sierras Pampeanas. Magmatic activity ended in the Principal Cordillera at approximately 8.6 Ma and in the Sierras Pampeanas at 1.9 Ma.

The last subduction-related volcanism occurred more than 750 km east of the trench. Analyses performed along four different transects show close relations between eastward propagation of the magmatic arc, formation of new brittle–ductile transition zones as a function of heat input, and orogenic deformation.

Uplift and deformation in Sierras Pampeanas, thus clearly follows the eastward propagation of arc magmatism. This relation implies that uplift and deformation are more likely related to thermal weakening and crustal anisotropy than to fluctuations in horizontal compressive stress. Eastward propagation of magmatism to the foreland over the last 6 m.yr plays an important role in the thermal crustal weakening and development of brittle–ductile transitions that allowed uplift of Sierras Pampeanas basement-cored blocks.

Keywords: Flat-slab segment; Basement; Migration; Magmatism; Uplift

1. Introduction

Several sectors of the Andes record the tectonic consequences of flat subduction of the Nazca plate in the late Cenozoic, including the Bucaramanga segment in the northern Andes (north of 5°N latitude) and the Peruvian (5–15°S

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PII: S0895-9811(02)00006-8

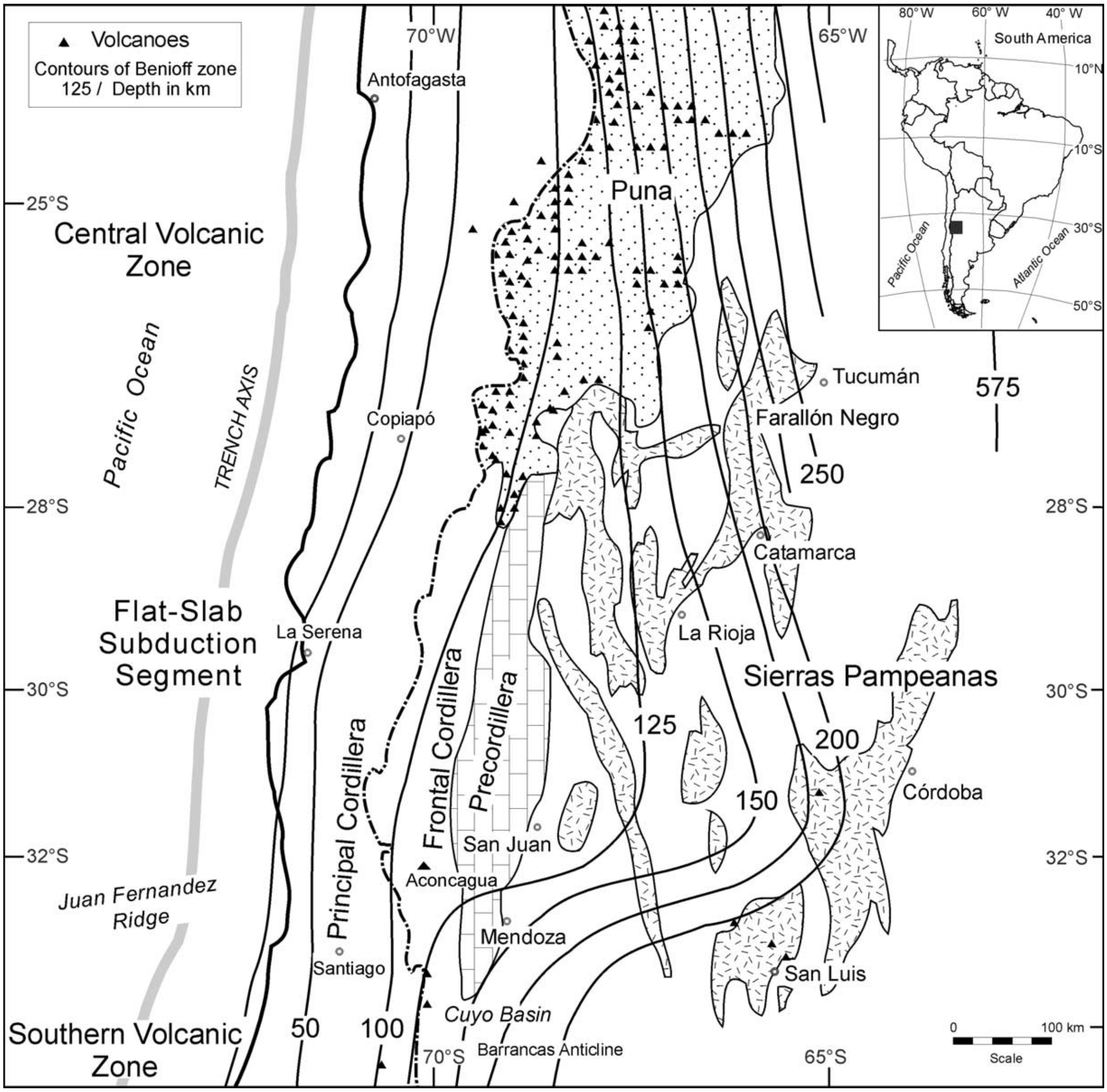


Fig. 1. Location of the Pampean flat-slab segment in the Central Andes of Argentina and Chile with contours of depth to the oceanic slab (after Cahill and Isacks (1992)) and outline of the major structural provinces and the basement blocks of the Sierras Pampeanas.

latitude) and Pampean (27°00′–33°30′S latitude) segments in the Central Andes (see recent review by Gutscher et al. (2000a)). However, the history of the geological processes associated with the shallowing of the subduction zone is better recorded in the Central Andes of Argentina and Chile (Fig. 1). This is due to the combination of a more advanced structural stage of evolution of this segment, in that flat subduction began here between 18 and 12 Ma, and an outstanding level of exposure resulting from the aridity of the eastern slope of the mountains at these latitudes. The region has been the focus of a continuous effort to study its magmatic, tectonic, and sedimentary history (Jordan et al., 1983a,b, 1993, 1997; Kay et al., 1988, 1991; Allmendinger et al., 1990; Kay and Abruzzi, 1996; Ramos et al., 1991, 1996a). The objective of this work is to analyze the structural, magmatic, and sedimentary response to late Cenozoic slab flattening in the Pampean flat-slab segment.

The Pampean flat-slab segment of the greater Andean orogenic system, defined in part by a gap of active arc volcanism in the main Andes, deformation, and late Cenozoic arc volcanism in Sierras Pampeanas, can be tied to the geometry and bathymetric relief of the subducting slab and specifically to the subduction of the Juan Fernandez Ridge. To analyze the relations among ridge subduction, slab flattening, and tectonics in the overriding plate, we focus specific attention on the timing and location of late Cenozoic deformation in four equally spaced structural transects. These are located in the northern Sierras Pampeanas at the latitude of Sierra de Aconquija (27°S) and Sierra de

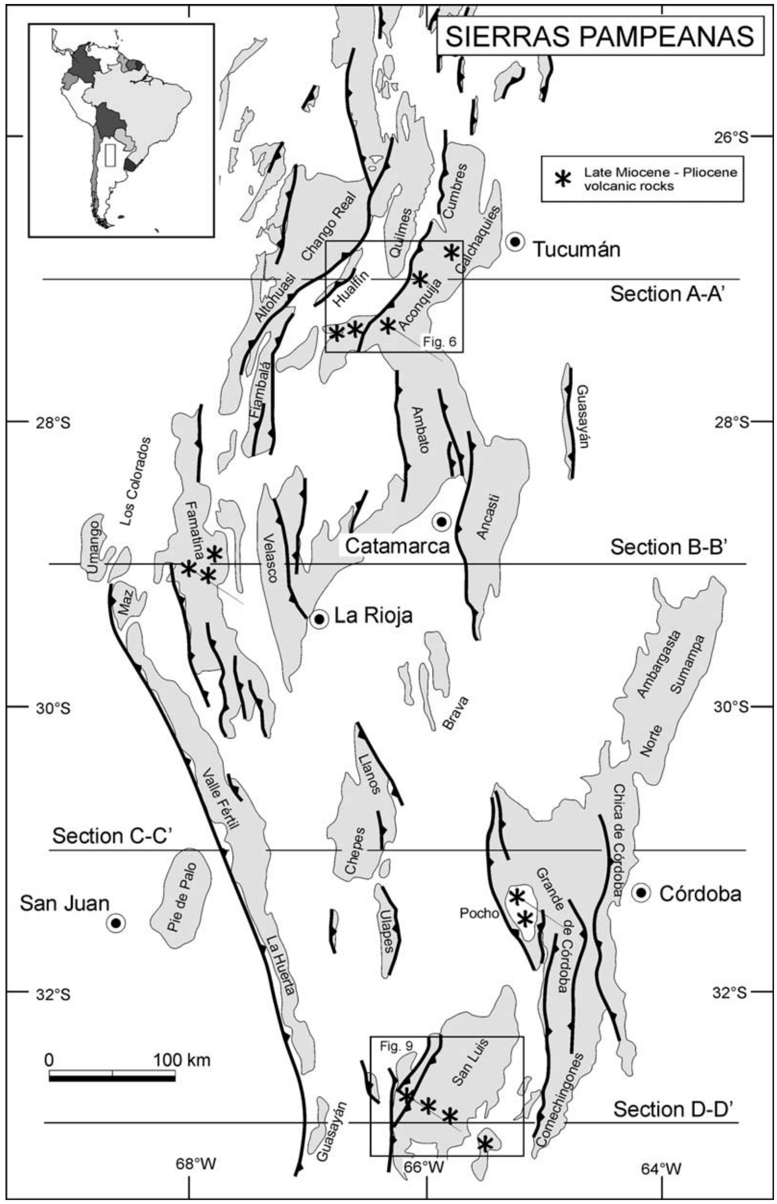


Fig. 2. Late Cenozoic Sierras Pampeanas mountain uplifts developed as a consequence of shallowing of the subduction zone. Major fault vergence based on González Bonorino (1950) and Jordan and Allmendinger (1986). Locations of crustal cross-sections of Fig. 4 and maps of Figs. 6 and 9 are indicated.

Famatina (29°S), the Central Sierras Pampeanas at the latitude of Sierras de Córdoba (31°S), and the southern Sierras Pampeanas at Sierras de San Luis (33°S). To integrate the evolution of the Sierras Pampeanas with Andean evolution as a whole, the results are incorporated into a transect across the entire Andean belt at 33°S, coincident with the southern reach of the present flat-slab sector.

1.1. Geologic setting

The present-day tectonic framework of the Pampean flat-slab segment is characterized by a gap in arc volcanism and the occurrence of a deformed and faulted foreland (i.e. basement blocks of Sierras Pampeanas; Fig. 2). The early work of Barazangi and Isacks (1976, 1979) recognized the presence of the Pampean segment of subhorizontal subduction between 27 and 33°30′S latitudes and confirmed a previous assumption by Stauder (1973) regarding the geometry of the subducted slab beneath Chile. Gaps in the volcanic arc led these authors to infer the absence of an asthenospheric wedge to interact with dehydration of the oceanic crust (Isacks and Barazangi, 1977).

The hypothesis advanced by Isacks et al. (1982), Jordan et al. (1983a)—that the flat-slab segment was linked to the formation of the Sierras Pampeanas—was corroborated by further studies (Jordan and Allmendinger, 1986; Ramos et al., 1991; Kay and Abbruzzi, 1996). In short, intraplate seismicity in the upper 50 km of the continental plate results

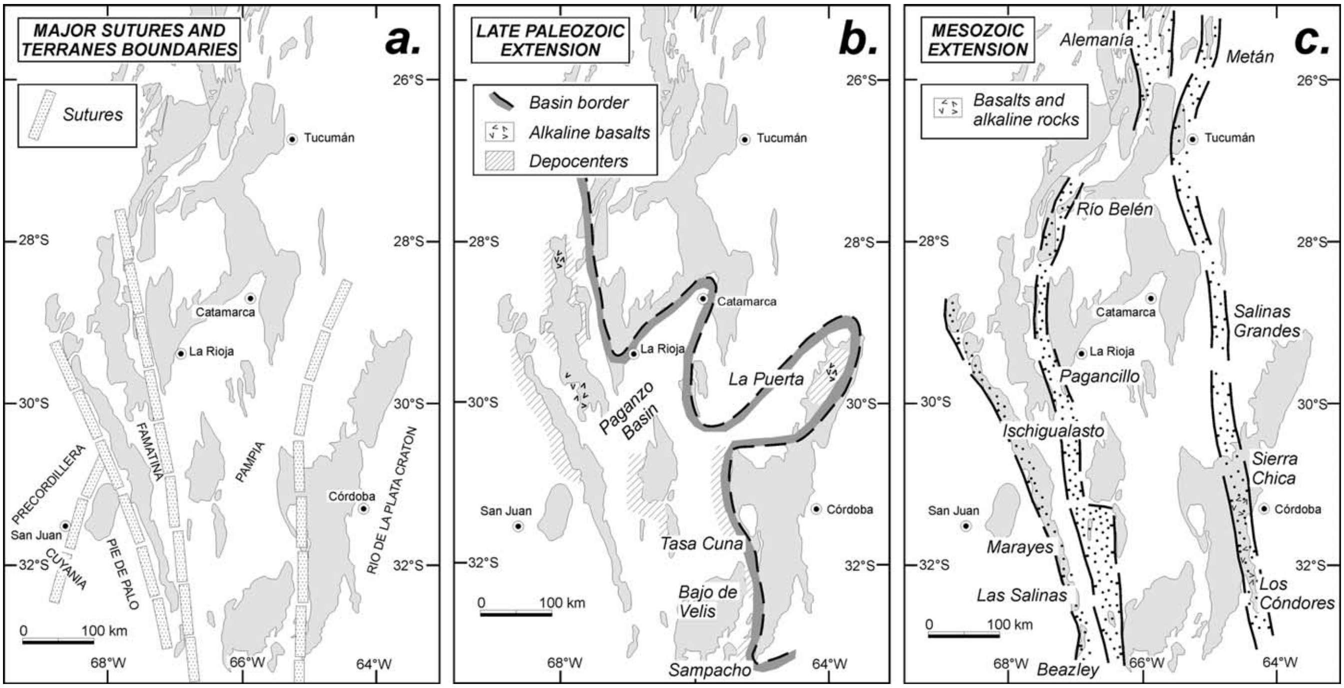


Fig. 3. Major discontinuities in the Sierras Pampeanas: (a) potential sutures and main terrane boundaries, (b) late Paleozoic Paganzo basin with main depocenters and associated alkaline basalt (modified from Salfity and Gorustovich (1984) and Koukharsky et al. (2002)), and (c) major rift systems of Triassic–early Jurassic and early Cretaceous ages and associated basalt and other alkaline rocks (based on Ramos (2000)).

in basement uplift and late Cenozoic shortening in the foreland region. Thus, the timing and spatial development of this deformed foreland gives indirect evidence of the processes related to the shallowing of the oceanic plate.

The older crustal discontinuities in the foreland played a strong role in the inception and geometry of master faults that differentially uplifted the blocks of Sierras Pampeanas (Ramos, 1994). These fault zones were mainly sutures and shear zones (Toselli et al., 1985) formed during the early Paleozoic consolidation of basement terranes. The master faults were reactivated by extension in late Paleozoic times during the orogenic collapse that formed the intracratonic Paganzo basin (Salfity and Gorustovich, 1984). A second episode of extensional deformation was associated with a generalized Triassic to early Jurassic rifting (Ramos and Kay, 1991) that predated the last phase of crustal extension related to the opening of the south Atlantic Ocean in Cretaceous times (Schmidt et al., 1994; Ramos, 2000).

1.2. Sierras Pampeanas

Uplift of the basement blocks of the Sierras Pampeanas was the result of Andean compression during late Cenozoic times. Tilting and rotation of the mountain blocks were controlled by décollments at the depth of the brittle–ductile transitions in the basement (as proposed by González Bonorino (1950)) located at different depths, as illustrated by Jordan and Allmendinger (1986), Introcaso et al. (1987). The relationships between older crustal discontinuities or weakness zones and present mountain uplifts of the Sierras Pampeanas are indicated in Fig. 3.

2. Basement control

2.1. Major crustal discontinuities

Although a generalized consensus does not exist regarding the structural outline of the basement terranes in the central region of Argentina and the timing of the collision, most authors recognize that this part of South America is a collage of parautochthonous and allochthonous terranes. A series of ophiolitic belts ranging in age from late Proterozoic to early Paleozoic indicates the location of potential sutures (for details, see Ramos et al. (2001)).

A major suture has been recognized between the Cuyania and Famatina terranes (Fig. 3(a)) on the basis of lithologic grounds (Vujovich and Kay, 1996; Ramos et al., 1996b), Bouguer gravity and magnetic anomalies (Giménez et al., 2000), sedimentary and volcanic history (Astini, 1998), and reprocessing of deep seismic industrial reflection lines (Zapata and Allmendinger, 1996). This suture juxtaposes Cuyania, a Laurentia-derived terrane that includes the Precordillera, against the protomargin of Gondwana (Astini and Thomas, 1999).

This major crustal discontinuity controlled the inception of the Valle Fértil and related faults, and in the late Cenozoic, it controlled almost the continuous basement uplift of the Sierra de Valle Fértil, La Huerta, Guasayán, and other ranges (Ramos, 1994) along more than 600 km (Fig. 2). This northwesterly trending fault in the Sierras Pampeanas is oblique to the major structures of the Andes. Detailed studies show that Cenozoic fault zone reactivation occurred along ductile shear zones formed during ancient collisional events (Schmidt et al., 1994).

Another major suture is located along the eastern side of the Sierra de Famatina, where mafic Ordovician rocks, as well as ultrabasic and basic assemblages, are exposed in the Paimán and Fiambalá areas (Pérez, 1991; Neugebauer and Miller, 1996; Grissom et al., 1998). This area has been interpreted as the suture between the Famatina and Pampia terranes, and these rocks are associated with important mylonitic zones formed during the early Paleozoic. The late Cenozoic Andean faults of Sierras de Fiambalá and Paimán were partially controlled by older ductile shear zones.

A third important suture, located between the Pampia cratonic terrane and the craton of Río de La Plata (Figs. 3(a) and 4) and outlined by Kramer et al. (1995) along the western side of Sierras de Córdoba, coincides with Cenozoic Andean faults. Precambrian structural fabrics within ductile shear zones control the western vergence of those Andean brittle faults. Even neotectonic movements occur along these fabrics (Massabie and Szlafsztein, 1991).

The important collisional episodes that occurred in late Proterozoic and early Paleozoic times in this part of southern South America (i.e. those that formed the western margin of Gondwana) also developed a penetrative deformation within the basement. This deformation is expressed by schistosity and foliation with growth of discrete ductile shear zones (Toselli et al., 1985; Von Gosen, 1998). These zones were significant in controlling the inception and geometry of later extensional and compressive deformation (Hongn and Seggiaro, 1998).

2.2. Late Paleozoic extension

Some of the most important sutures separating different amalgamated terranes were reactivated by extension during the orogenic collapse that followed early Paleozoic orogenic mountain building (Mpodozis and Ramos, 1989). The Paganzo basin, a late Paleozoic intracratonic basin, where thick sequences of taphrogenic sediment accumulated, has a series of depocenters controlled by these major crustal discontinuities (Fernández Seveso et al., 1993). The largest depocenters were associated with major crustal boundaries inherited from early Paleozoic orogenies and are considered typical episutural depocenters sensu Bally (1989). Right lateral strike-slip faulting along older structures associated with oblique subduction may have controlled pull-apart rifting during the late Paleozoic.

Some of the depocenters have alkaline basaltic flows, with typical intraplate signatures in the northern and southern margins of the Famatina terrane (Thompson and Mitchell, 1972; Valencio, 1980) and in the Sierra Norte region (Koukharsky et al., 2001). Paleomagnetic studies in these sequences confirm clockwise rotation of these blocks in several areas of the late Paleozoic basin (Rapalini and Vilas, 1991).

The region affected by the extension related to oblique subduction is confined to the southwestern sector of Sierras Pampeanas, as depicted in Fig. 3. Although the initial depocenters were linked to active faulting, the expansion of the sedimentation in the upper sections of the Paganzo basin was thermally controlled. Sag-phase facies were deposited directly on the basement over larger areas, as observed in the Sierra de la Huerta area.

Several Andean faults inverted the extensional late Paleozoic faults, as seen in the Sierra de Los Llanos and Chepes. The same paleogeographic control is seen in the Valle Fértil fault, where the isopachs of the Paganzo sequences thicken toward this major lineament. In the Famatina area, many older, normal faults are controlling the present reverse faults (Durand, 1996).

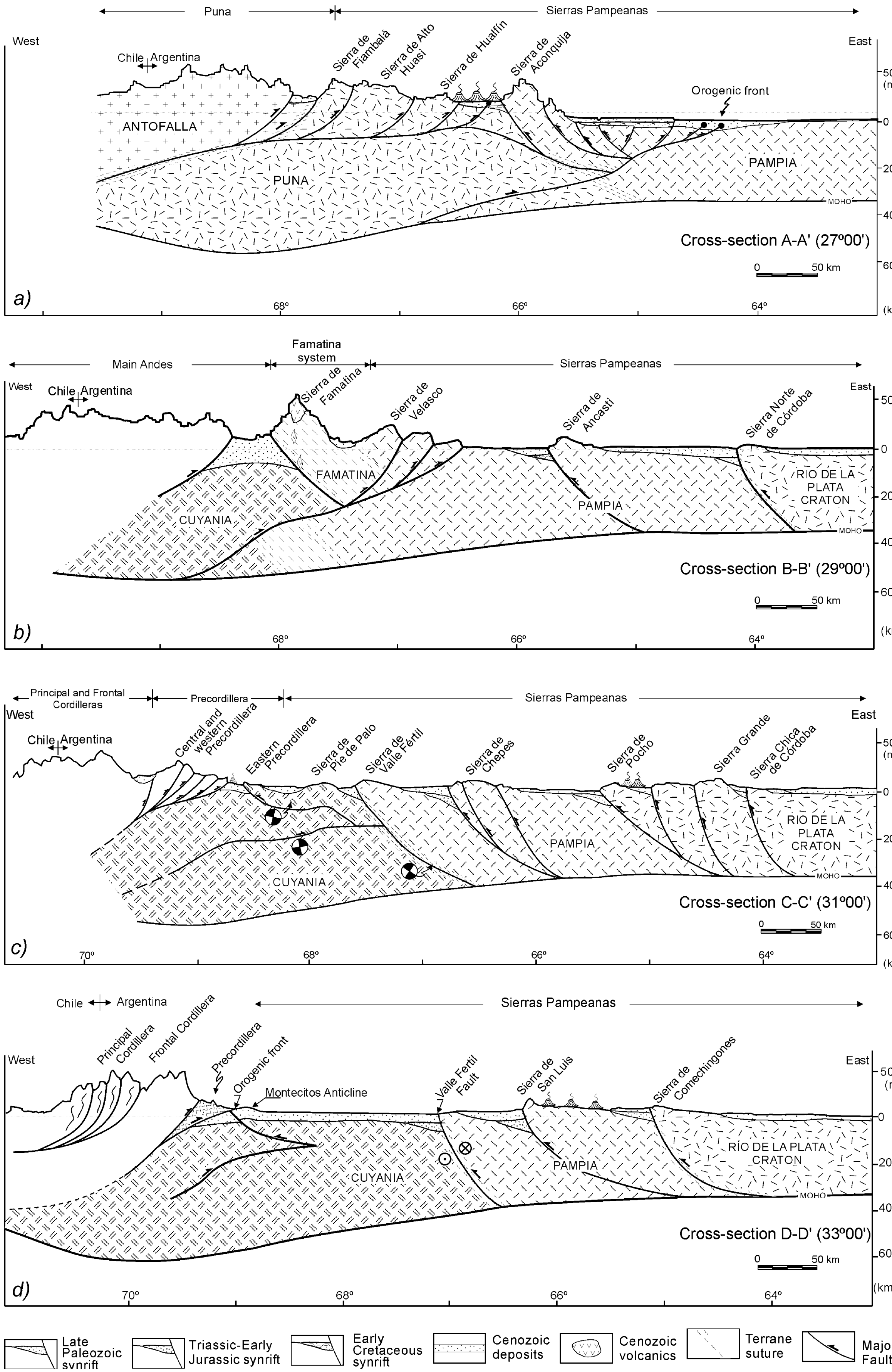
2.3. Late Triassic—early Jurassic extension

The Triassic master faults of the rift system formed on the hanging-wall of the sutures between the Cuyania and Pampia terranes, as well as between the Cuyania and Chilenia terranes. This last suture is located west of Sierras Pampeanas, along the present contact between the Frontal Cordillera and the Cuyo Precordillera (Ramos and Kay, 1991). Several depocenters ranging from 1500 to 2000 m in thickness developed along the Valle Fértil fault as the Ischigualasto, Marayes, Las Salinas, and Beazley basins. The synrift deposits of these basins are associated with alkaline basalts of intraplate signature. Time constraints in the stratigraphic fill indicate a late Triassic to early Jurassic age (Kokogián et al., 1993) with younger sedimentation toward the eastern areas.

Andean-age reverse faults in the Sierras de La Huerta, Valle Fértil, Chepes, and Ulapes reactivate the older, normal faults, though in some of these areas, new basement shortcuts are present (Chiaramonte et al., 2001). Most of these faults have a long history of deformation, as seen in the western slope of the Sierra de Chepes. Here, an early Paleozoic ductile fault zone was extensionally reactivated in the late Paleozoic and early Mesozoic times and inverted during Andean compression. This fault is still active today.

2.4. Early Cretaceous extension

Relative to the older extension, the locus of the early Cretaceous normal faulting shifted eastward. This episode is temporally and geographically linked with the opening of the south Atlantic at these latitudes (Rossello and Mozetic, 1999). Major narrow rifts developed along the western and eastern margins of the Pampia terrane with depocenters, as much as 5000 m thick. It is evident that the Cretaceous rifting was tectonically more active, areally extensive, and severe than the Triassic–early Jurassic rifting.


Puna
Sierras Pampeanas
West
East
Chile Argentina
Sierra de Fiambalá
Sierra de Alto Huasi
Sierra de Hualfín
Sierra de Aconquija
Orogenic front
ANTOFALLA
PUNA
PAMPIA
MOHO
5000 (m)
0
20
40
60
(km)
Cross-section A-A' (27º00')
0 50 km
a)
68º
66º
64º
Main Andes
Famatina system
Sierra de Famatina
Sierra de Velasco
Sierra de Ancasti
Sierra Norte de Córdoba
FAMATINA
CUYANIA
RIO DE LA PLATA CRATON
Cross-section B-B' (29º00')
b)
Principal and Frontal Cordilleras
Precordillera
Central and western Precordillera
Eastern Precordillera
Sierra de Pie de Palo
Sierra de Valle Fértil
Sierra de Chepes
Sierra de Pocho
Sierra Grande
Sierra Chica de Córdoba
Cross-section C-C' (31º00')
c)
70º
Principal Cordillera
Frontal Cordillera
Orogenic front
Montecitos Anticline
Valle Fertil Fault
Sierra de San Luis
Sierra de Comechingones
RÍO DE LA PLATA CRATON
Cross-section D-D' (33º00')
d)
Late Paleozoic synrift
Triassic-Early Jurassic synrift
Early Cretaceous synrift
Cenozoic deposits
Cenozoic volcanics
Terrane suture
Major Fault

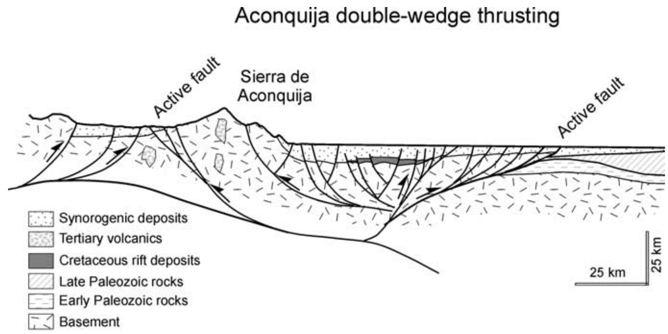


Fig. 5. Structural cross-section of Sierra de Aconquija based on reprocessed deep seismic reflection data (modified from Cristallini et al. (2001)).

The early Cretaceous extension observed within the Sierras Pampeanas propagated to the north within the Salta rift system, which is structurally bigger and deeper than the southern Cretaceous rift (Salfity and Marquillas, 2000). The northern rift has several linked depocenters that accumulated more than 6000 m of early and late Cretaceous synrift sediment (Comínguez and Ramos, 1994).

Alkaline basaltic flows, as well as basic alkaline intrusives, are abundant in the Sierra Chica de Córdoba (Kay and Ramos, 1996) and further north in the Río Belén area of Catamarca (Rossello et al., 1999).

The described normal fault systems controlled the uplift and inception of important regional faults in eastern Sierras Pampeanas. Based on the previous description, it is evident that the geometry and style of deformation, as well as the location of the Andean-age faults of the Sierras Pampeanas, were controlled by older crustal-scale structures. These faults, represented by early Paleozoic ductile shear zones and Mesozoic normal faults, were controlled by the basement fabrics and have a dominant dip to the east. As a result, Andean shortening in the Sierras Pampeanas has a dominant west vergence (Figs. 2 and 4), opposite the main vergence of the Principal and Frontal Cordilleras of the Andes to the west.

3. Timing of deformation

Most of the Sierras Pampeanas uplift was diachronically produced during the late Cenozoic. The areas involved were determined by the older structural fabrics, but the timing of deformation shows an important link with magmatic activity. To identify this relation, four equally spaced crustal transects across Sierras Pampeanas, at 27, 29, 31, and 33°S, were constructed and analyzed (see Fig. 4(a)–(d)).

3.1. Northernmost Sierras Pampeanas (27°S)

The crustal transect at 27°S encompasses the northernmost part of the Sierras Pampeanas, which constitutes a narrow belt of basement blocks. Most prominent among them is the Sierra de Aconquija (Fig. 4(a)), which is the highest range in the Sierras Pampeanas (5550 m a.s.l.). The abnormal northeast trend of the Sierra de Aconquija, when compared with other Sierras Pampeanas to the south, is the result of the Neogene interaction of the Aconquija, or Tucumán, Lineament (Mon, 1976; Ramos, 1977). The kinematics of this Tucumán transfer zone has been extensively studied by Urreiztieta et al. (1996), who interpreted it as a 100-km-wide dextral transpressive zone that accommodated the differences in Neogene shortening between the Puna–Subandean system and the Principal Cordillera–Precordillera–Sierras Pampeanas system.

This transect across the Sierra de Aconquija is controlled by deep seismic information to a depth of 50 km (Fig. 5; Cristallini et al., 2001). The seismic reflection data reveal double-wedge thrusting, which confirms the existence of this structural style, as proposed by Mon and Drozdzewski (1999) for this sector of the Sierras Pampeanas. Although both sides of this double-wedge thrusting record neotectonic faulting, the western side displays greater offset (Strecker et al., 1987; Cristallini et al., 1998; Hermanns and Strecker, 1999).

The corresponding basement uplift is surrounded by synorogenic deposits that fill a Neogene foreland basin, which is broken by the uplift of the Sierra de Aconquija. Detailed stratigraphic and magnetostratigraphic studies of these synorogenic deposits constrain the timing of the Aconquija uplift.

Deposits along the western slope of the Sierra de Aconquija are included in the Santa María Group, represented from base to top by the San José, Las Arcas, Chiquimil,

Fig. 4. Crustal cross-sections of the Andes at 27, 29, 31, and 33°S latitudes. Location of the sections in Fig. 2: (a) northernmost Sierras Pampeanas showing double-wedge thrusting of the Sierra de Aconquija, documented by deep seismic reflection data (Cristallini et al., 2001); (b) Famatina and northern Sierras Pampeanas section with interpreted double-wedge thrusting; (c) Central Sierras Pampeanas. Décollement levels in Pie de Palo are based on intraplate earthquake focal mechanisms (Regnier et al., 1992) and the Valle Fértil fault geometry constrained by deep seismic reflection data down to 15–20 km (Snyder et al., 1990); (d) crustal wedge beneath the orogenic front constrained by deep seismic reflection data at a depth of 30 km (based on Comínguez and Ramos (1991)).

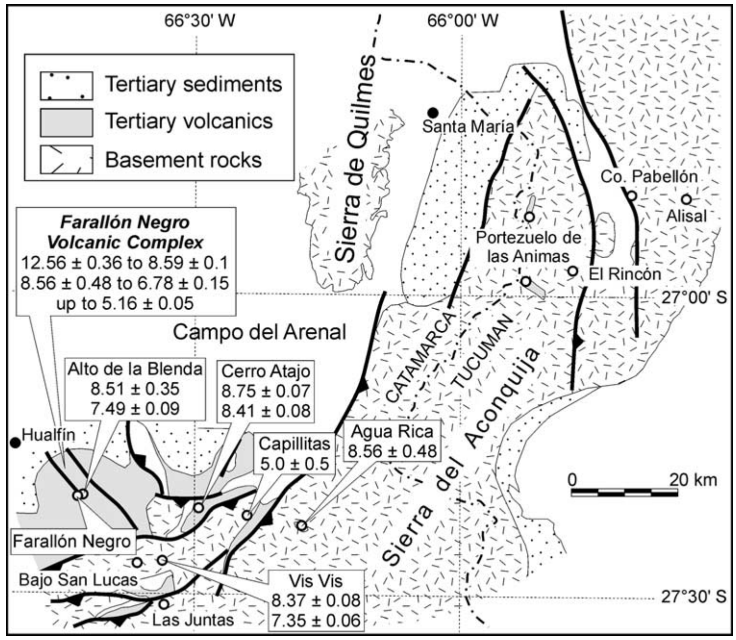


Fig. 6. Geologic map of Sierra de Aconquija region (based on González Bonorino (1951) and Ruiz Huidobro (1972)). Location of the volcanic rocks based on Jones (1997), González et al. (2000). Geochronological data are mainly Ar/Ar ages from Sasso and Clark (2000). The age and facies of synorogenic Tertiary sediments west of the Sierra de Aconquija constrain the uplift of the range.

Andalhualá, Corral Quemado, and Yasyamayo Formations (Ruiz Huidobro, 1972). The fine-grained gray sandstone and green siltstone of the San José Formation have middle Miocene forams of the Paraná marine transgression (Gavriloff and Bossi, 1992). The San José Formation is conformably overlain by red sandstone and conglomerate of Las Arcas Formation, which is covered by coarse to medium- to coarse-grained yellowish volcaniclastic sandstone of the Chiquimil Formation. This unit is subdivided in two members, a fine-grained lower member and a coarser upper member. The boundary between these two members has been dated at 6.68 Ma by magnetostratigraphy (Strecker, 1987). On the basis of an east-to-west paleocurrent direction, Strecker (1987) interpreted the upper member of the Chiquimil Formation as a response to an incipient uplift of the Sierra de Aconquija. Paleoclimatic considerations caused Strecker (1987) to claim that this block also remained as a low-relief positive block during sedimentation of medium to coarse sandstone and lensoid conglomerate of the Andalhualá Formation. This unit was deposited between 6.02 and 4.0–3.4 Ma and was covered by the coarse conglomerate of the Corral Quemado Formation. These conglomerate beds mark the final Neogene uplift of the Sierra de Aconquija (Strecker, 1987). The top of the Corral Quemado Formation has been dated at 2.97 ± 0.6 Ma. Very coarse and well-cemented conglomerate of the Yasyamayo Formation unconformably overlie the Santa María Group and represent the first pedimentation level of Quaternary age on the western slope of the Aconquija (Strecker et al., 1989). The synorogenic deposits constrain the main uplift of the Sierra de Aconquija between 6 and 4–3.4 Ma, uplift that was still active in Quaternary times (Strecker et al., 1989).

Apatite fission-track studies performed on the basement rocks of the Sierra de Aconquija provide, in contrast, an independent way to evaluate the uplift time of the range (Sobel et al., 1998; Coughlin et al., 1999). Results from 11 samples collected over a vertical interval of 1380 m from the hanging-wall of the western Aconquija fault indicate that the basement cooled rapidly from temperatures above 110 °C to less than 50 °C during the late Miocene to early Pliocene (7.6 ± 2.2–4.2 ± 1.1 Ma; Coughlin, 2000). Two other samples indicate rapid cooling after 3 Ma.

These data, when combined with the stratigraphic data provided by Strecker (1987), clearly indicate that the main uplift of the Aconquija can be bracketed between 7.6–6 and 4–3.4 Ma.

West of Sierra de Aconquija, a large volcanic field known as the Farallón Negro Volcanic complex was formed by a series of intrusive and extrusive volcanic retroarc rocks spread over an area of approximately 700 km^2 (Llambías, 1970). Beside this main volcanic field, a cluster of subvolcanic bodies was emplaced in the basement rocks west of the Aconquija fault, such as Capillitas, Cerro Atajo, Bajo San Lucas, Vis Vis, and Las Juntas (Jones, 1997; Sasso and Clark, 2000). Even the basement of Sierra de Aconquija has several minor bodies of volcanic rocks located in the hanging-wall of the western fault. These subvolcanic and volcanic rocks are represented by the andesite, dacite, and basaltic andesite of the Alisal Volcanic Complex in Cumbres Calchaquíes in the north (see Fig. 6); the Portezuelo de Las Ánimas andesite and andesitic breccia, which

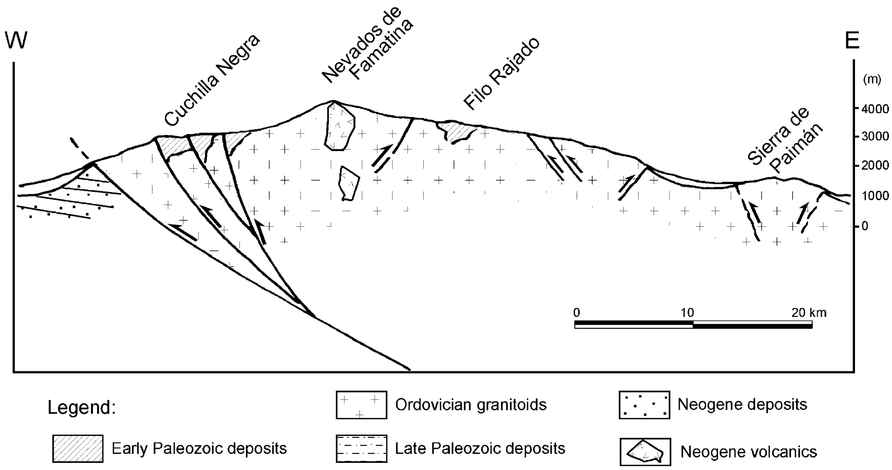


Fig. 7. Structure of Sierra de Famatina showing west vergence of main Andean thrusts and minor late Paleozoic extensional faults reactivated during Cenozoic deformation (modified from Durand (1996)).

cover an area of 50 km^2 in the northern Aconquija and have been interpreted as remnants of a large stratovolcano by González (1990); the El Rincón y Huertas Viejas andesitic lavas, pyroclastic rocks, and breccia in the eastern slope of the Aconquija (González et al., 1989); and El Pago volcanics (Alderete, 1974). In the southern part of the Aconquija, some other volcanic centers, such as Agua Rica, show a complex history of volcanic and intrusive rocks (Roco and Koukharsky, 2000).

All these exposures show that, probably during the late Miocene, most of the Sierras de Aconquija and Cumbres Calchaquíes were the loci of widespread volcanism. Many ages are available from the Farallón Volcanic Complex and the southern part of Sierra de Aconquija. These ages indicate three important periods of magmatic activity (Sasso and Clark, 2000). The first stage is related to the beginning of volcanism, with eruption of andesite, basalt, and dacite widely over the entire area between 12.56 ± 0.36 and 8.59 ± 0.1 Ma. The second stage relates to a large stratovolcano between 8.51 ± 0.35 and 6.78 ± 0.15 Ma, and the third, late stage is related to important hydrothermal alteration at approximately 5.16 ± 0.05 Ma (Ar/Ar dates after Sasso and Clark, 2000). The easternmost activity in Agua Rica has similar ages. Those volcanics close to the Aconquija fault, such as the rhyolite and dacite of Capillitas, have ages of 5 ± 0.5 Ma (Márquez-Zavalía, 2000). The volcanic rocks of Portezuelo de Las Animas have an age of 11.6 and 7.7 Ma, according to González et al. (2000).

Thus, the main uplift of the Sierra de Aconquija, as constrained by magnetostratigraphy and fission-track analyses, occurred after the expansion of the volcanic arc toward the east retroarc area. Uplift began between 7.6 and 6 Ma, immediately after a peak of volcanic activity at 6.78 Ma. The cessation of magmatism at about 5 Ma was succeeded by the longest uplift of the range after 4.27–4.0 Ma, as indicated by the reverse faults that cut most of the volcanic outcrops of Sierra de Aconquija (González, 1990).

3.2. Northern Sierras Pampeanas (29°S)

The transect of Sierras Pampeanas at 29°S crosses the Sierras de Famatina, Velasco, and Ancasti, as well as some minor hills related to the northern end of Sierras de Córdoba (Fig. 4(b)). The ranges encompassed in the Sierra de Famatina system, with elevations up to 6050 m, consist of a major basement uplift of crystalline basement rocks and a Paleozoic cover. Seismic data are not available to constrain the subsurface geometry. However, the surface geology indicates a double-wedge thrust system similar to the Sierra de Aconquija (Fig. 7).

The timing of the late Cenozoic uplift of the Sierra de Famatina can be constrained using the seismostratigraphy of the adjacent uplifted synorogenic depositional sequences of the Sierra de Los Colorados, a few kilometers west of Famatina (Fig. 2). This range exposes more than 10,000 m of late Cenozoic deposits and is one of the thickest Andean foreland basin sequences in the Central Andes (Ramos, 1970). The basin was a continuous and wide basin, until the uplift of the Sierra de Famatina, when it became a faulted foreland basin documenting rapid subsidence.

The base of the synorogenic sequences is composed of sandstone and shale of the Vinchina Formation, which consists of low-energy braided fluvial facies as thick as 5730 m that coarsen slightly to the north and south from a mud-rich axis in the central part. Two coarsening upward members of this unit represent distal synorogenic facies related to the Frontal Cordillera uplift further to the west. The Toro Negro Formation unconformably overlies the Vinchina Formation and consists of coarsening upward conglomerate and ash-fall tuff as thick as 5254 m. The conglomerate has abundant clasts of Cenozoic volcanic rocks and constitutes proximal facies that represent the beginning of the Sierra de Famatina uplift. Unconformably overlying the Toro Negro unit are several 100 m of fanglomerate deposits of the Santa Florentina Formation, which marked the deformation and cannibalization of the

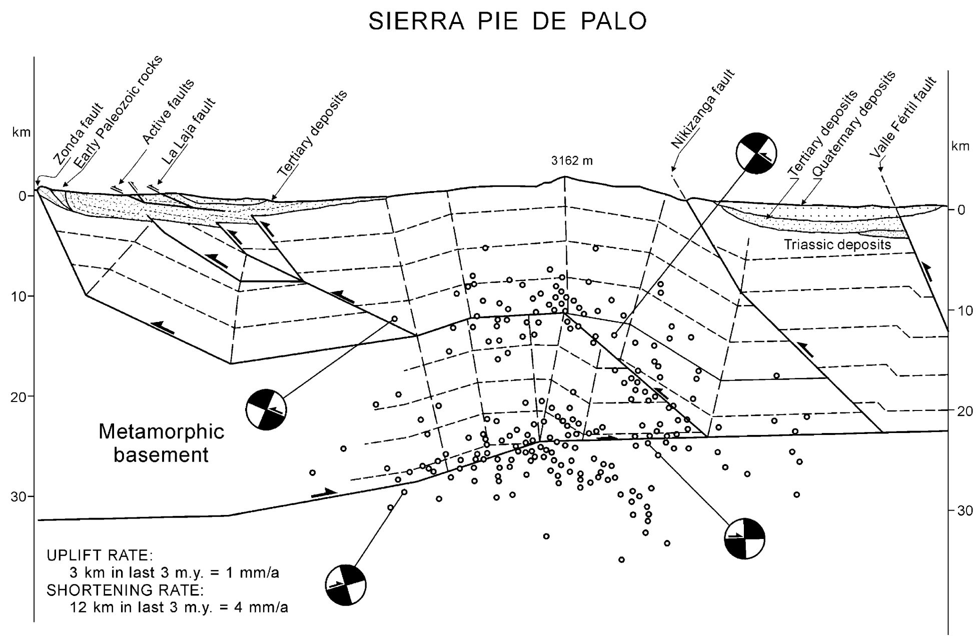


Fig. 8. Crustal cross-section of Sierra de Pie de Palo showing an east-vergent basement wedge constrained by field data and earthquake focal mechanisms provided by Regnier et al. (1992) (based on Ramos and Vujovich (2000)).

synorogenic basin during the final uplift of the Sierra de Famatina.

Time constraints from magnetostratigraphic studies indicate an age between 15 and 4.5 Ma for the Vinchina Formation (Limarino et al., 1999; Tripaldi et al., 2002), and therefore, a Pliocene age for the Toro Negro Formation. The Santa Florentina Formation is considered to be early Quaternary in age.

An independent way to constrain the uplift age is derived from isotopic studies from the Sierra de Famatina. Hydrogen isotope data for supergene alunite and hypogene sericite from epithermal precious metal deposits at Nevado de Famatina have a δ H_2O between -15 and -5%, indicating paleometeoric water at relatively low elevations at time of alunite formation (Taylor et al., 1997). These authors conclude that uplift of Sierra de Famatina should be younger than approximately 4 Ma, the age of alunite and sericite formation.

Migration of the main arc toward the foreland was responsible for the emplacement of several bodies of dacite and rhyodacite porphyries in the central part of the Sierra de Famatina. These masses are now exposed in the Mogote Río Blanco, Los Bayos, La Estrechura, Caballo Muerto, and many other areas of the Nevados de Famatina, widespread along 50 km (Brodtkorb et al., 1996). A compilation of available Ar/Ar ages indicates important magmatic activity from 6.38 ± 0.37 to 4.24 ± 0.11 Ma, as pointed out by Toselli (1996). Alunite and sericite associated with epithermal alteration and high sulphidation vein systems range in age from 4.84 ± 0.41 to 4.19 ± 0.27 Ma (Losada Calderón et al., 1994).

Magnetostratigraphic data and the isotope composition of paleometeoric waters indicate an age between 4.5 and 4.19 Ma for the beginning of uplift of Sierra de Famatina. Again, along this transect, large uplifts immediately postdate the migration and cessation of arc magmatism.

Toward the east, the Sierras de Velasco and Ancasti have no constraints to date their uplift age precisely. Synorogenic deposits in the Sierra de Ancasti, described as the Portillo Formation, consist of shale and gypsum that have been correlated with the Paraná marine transgression (Aceñolaza et al., 1983). This unit represents the most distal deposits of the Frontal Cordillera foreland basin. The conglomerate and tuff of the Las Cañas Formation conformably overlying the older unit are interpreted as Pliocene proximal deposits, linked with the uplift of neighboring ranges within Sierras Pampeanas. Main uplift occurred during Quaternary times, as indicated by the angular unconformity between coarse proximal fanglomerates and the Pliocene units and geomorphic evidence (Costa, 2000).

3.3. Central Sierras Pampeanas (31°S)

The central section at 31°S is one of the most complete exposed sections of the entire Sierras Pampeanas, a sector associated with important seismotectonic activity (Fig. 4(c)). Neotectonic structures and Quaternary active faulting are known along the western-bounding faults of the Sierras de Chepes and Valle Fértil and its prolongation in the Sierra de La Huerta. Present-day tectonics and uplift are concentrated further west in Sierra de Pie de Palo. This neotectonic activity matches the Eastern Precordillera uplift during late Quaternary times.

Geologic evidence in the Sierra de Pie de Palo indicates that the range was covered by late Pliocene distal synorogenic deposits derived from the Precordillera (Ramos and Vujovich, 2000). The predominance of thrust-type focal mechanisms alone shows that Pie de Palo is being actively compressed. On the basis of the location of intraplate focal mechanisms provided by Regnier et al. (1992), two décollement levels were proposed, one at about 12–15 km and another at 22–25 km depth. These décollments may explain the surface east-vergent deformation observed in the Sierra de Pie de Palo (see Fig. 8). The structure consists of an east-vergent, middle crustal wedge ramping beneath the Pie de Palo, which produced the arching of the dissected Gondwana peneplain exposed across the range. Jordan and Allmendinger (1986) notice that earthquake and aftershock hypocenters along west-dipping master faults that bounded the range to the east indicate motion between 25 and 17 km depth, but not at shallower levels. This is presently explained by a mid-crustal wedge dipping to the west, that is, passively transporting the Precordilleran block beneath Sierra de Pie de Palo. This crustal wedge coincides with the Precordillera terrane boundaries, as described on seismic grounds by Smalley et al. (1993).

The next range to the east is Valle Fértil, with a southern prolongation in Sierra de La Huerta. This range is bounded to the west by an active, east-dipping reverse fault, the Valle Fértil fault, which can be traced over 600 km. The fault plane coincides with hypocenters in the region of Las Chacras. Earthquakes indicate that reverse motion occurs on faults at 15–32 km depth that dip 60–30° (Chinn and Isacks, 1983; Jordan and Allmendinger, 1986). Deep seismic reflection data available that image the Valle Fértil fault extend the surface information down to 15–20 km depth (Snyder et al., 1990). Focal mechanism solutions show a hypocenter at approximately 32 km depth, consistent with this fault trace. This indicates that the Valle Fértil fault is a major crustal discontinuity cutting most of the crust and, further, that it coincides with an early Paleozoic suture between the Cuyania terrane and Gondwana protomargin (Zapata and Allmendinger, 1996; Ramos et al., 1996b).

The most recent active important fault bounds the Sierra de Chepes to the west. This fault coincides with an early Paleozoic mylonitic zone (Ramos, 1982) that indicates a brittle reactivation of a basement shear zone. This fault, south of Sierra de Chepes, is also the eastern boundary of the Jurassic–early Cretaceous rift.

The magmatic history along this transect reflects a rapid expansion at about 7 Ma, when volcanic arc rocks were erupted simultaneously in the eastern Precordillera and the eastern ranges of the Sierras Pampeanas (Kay and Abruzzi, 1996). The western side of the Sierras de Córdoba is the locus of the Pocho volcanic rocks, which, in spite of being 600 km away from the oceanic trench, have a geochemical signature of subduction-related magmas (Kay and Gordillo, 1994). Pocho volcanic rocks can be divided in two series. The older one erupted between 7.9 and 6 Ma, and the younger series formed between 5.5 and 4.7 Ma (Gordillo and Linares, 1981). The oldest, more basic rocks are in the western side of the volcanic region and are deeply eroded, whereas the younger and more acidic series is associated with preserved volcanic edifices. The older volcanic rocks are cut by faults, indicating they were formed prior to the tectonic shortening that uplifted this sector of the Sierras de Córdoba.

Age evidence from sedimentary rocks is scarce, because the synorogenic deposits in the footwall of the east-dipping western fault of Sierras de Córdoba have not been dated. However, in the hanging-wall of the Sierra de Chepes block (Fig. 4(c)), the synorogenic deposits of the Los Llanos Formation were dated as Pliocene, an age partially corroborated by mammal fossils found in travertine deposits (Kraglievich and Reig, 1954) associated with either the late hydrothermal activity of the volcanic center or the tectonic activity of faults.

On the basis of these data, the timing of uplift of this sector of the Sierras de Córdoba is constrained by the ages of both volcanic series and can be bracketed between 6 and 5.5 Ma, probably in the uppermost Miocene. This fault has no Quaternary activity as is seen in the western region. Deformation continues toward the west, first in the Sierra de Chepes and Valle Fértil and presently in the Pie de Palo range. Deformation in Pie de Palo started at approximately 3 Ma and continues to the present with a shortening rate of about 4 mm/a and an uplift rate of 1 mm/a (Ramos and Vujovich, 2000).

3.4. Southern Sierras Pampeanas (33°S)

The transect at 33°S crosses the youngest uplifted sector of the Sierras Pampeanas. Neotectonic studies show widespread Quaternary activity along the major bounding faults of the Sierra de San Luis and Comechingones (Costa, 1992). These faults juxtapose crystalline basement over alluvial fan and piedmont deposits. Large earthquakes have been recorded along the transect, such as the San Martín event (May 22, 1936, 6.4M_L) in the northern Sierra de San Luis. Holocene slip, as much as 2.1 m has been recorded and dated in western-bounding faults of the Sierra de Comechingones (Costa and Vita-Finzi, 1996). These authors measured rakes between 87 and 71° in the fault plane and computed a slip rate of 2 mm per year in a 45° east-dipping fault for the last 1300 years. These figures imply a shortening rate of less of 1.3 mm per year in the late Holocene, almost one order of magnitude smaller than that of the active thrust front. Although generalized Quaternary activity is well documented, constraints to determine the beginning of the deformation are not well established. Most of the Neogene deposits that fill the adjacent basins are distal synorogenic deposits (Pascual and Bondesio, 1981) related to the interaction of the main Andean thrust fronts. This relation implies that the youngest proximal facies are mainly Quaternary or, at the earliest, late Pliocene in age.

The magmatic activity in the Sierra de San Luis consists

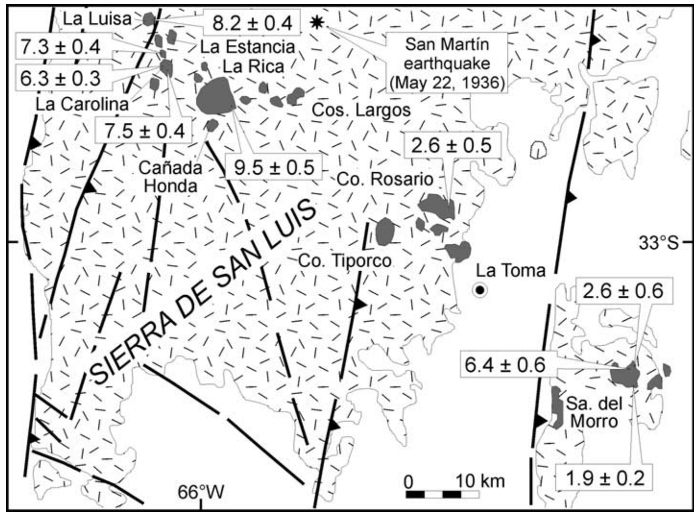


Fig. 9. Volcanic activity in Sierras de San Luis and adjacent areas (K/Ar ages, Ramos et al., 1991; Ar/Ar ages, Urbina et al., 1997). The epicenter of San Martín earthquake is after Costa and Vita-Finzi (1996).

of a series of volcanoes and subvolcanic bodies of andesitic, trachyandesitic, and trachytic composition and is well documented along a northwesterly trending belt from La Carolina to El Morro (Llambías and Brogioni, 1981). Their geochemical characteristics show a typical subduction signature (Kay et al., 1991), though the eruptions are located between 650 and 750 km east of the trench. Geochronologic data show that the volcanic activity began in the western sector of La Carolina between 9.5 ± 0.5 and 8.2 ± 0.4 Ma and rapidly expanded to the entire range between 6.3 ± 0.3 and 6.4 ± 0.6 Ma (Ramos et al., 1991; Urbina et al., 1997). Volcanic activity in the middle sector of the Cerros del Rosario and in the easternmost sector of El Morro lasted until 2.6 ± 0.6 Ma, when it ceased in the western sector, and in El Morro, until 1.9 ± 0.2 Ma (Fig. 9). This volcanic activity is the youngest record in the whole Sierras Pampeanas.

The comparison between time of uplift in the Sierra de San Luis, poorly constrained by the proximal synorogenic deposits of the late Pliocene and Quaternary, and the residence time of volcanism in this range, until 2.6 Ma show that uplift occurred around this time. The Sierra del Morro further east may have been uplifted at an earlier time.

4. Timing of late Cenozoic deformation in the Andes

Recent studies on the stratigraphy of the synorogenic deposits along the southern Central Andes have shown that deformation and uplift of the different ranges were not coeval, and when precise data are available, they fail to confirm previously presumed orogenic phases, such as Quechua, Diaguitic, and others (Reynolds et al., 1997; Jordan et al., 1997). The preexisting Paleozoic and Mesozoic structures controlled an uneven eastward propagation of the deformation toward the foreland (Coutland et al., 2001).

To illustrate the relation between the shallowing of the Nazca plate beneath the Andean system and the Sierras Pampeanas, a structural section exposed across 33°S has been chosen. This section at the southern end of the flat-slab is one of the youngest that formed during the shallowing of the subducting slab, yet it has a reasonable amount of geologic data to reconstruct the geometry and timing of the structures. The description of the different processes is presented in four stages.

4.1. Initial conditions (early Miocene ≈ 20 Ma)

A general agreement exists that there is a close relationship between aseismic ridge collision and flat-slab subduction (Barazangi and Isacks, 1976; Pilger, 1981, 1984; Sacks, 1983; Kirby et al., 1996; Gutscher et al., 2000a,b). The effects of ridge collision in the forearc and adjacent trench areas, as seen in submarine relief associated with the Juan Fernandez aseismic ridge, indicate that collision is moving southward (Pilger, 1981, 1984; Von Huene et al., 1997). On the basis of the convergence rates and the reconstructed shape of this ridge, Yañez et al. (2001) concluded that convergent parallel collision began at approximately 18 Ma in the northern latitudes and arrived at the 33°S latitude at about 11 Ma, which is in good agreement with geologic data.

On the basis of the preceding premises, it is reasonable to assume that, at about 20 Ma, the tectonic setting of the Principal Cordillera was not yet affected by the collision of the Juan Fernandez Ridge. At this time, the volcanic front was active along the Chilean slope of the Principal Cordillera. The Farellones volcanic arc was active, and thick sequences of andesite, dacite, and rhyolite lavas and pyroclastic rocks were unconformably deposited over large areas. Volcanic activity is bracketed between 20.4 ± 0.5 and 16.6 Ma in the segment between 32 and 33° (Rivano

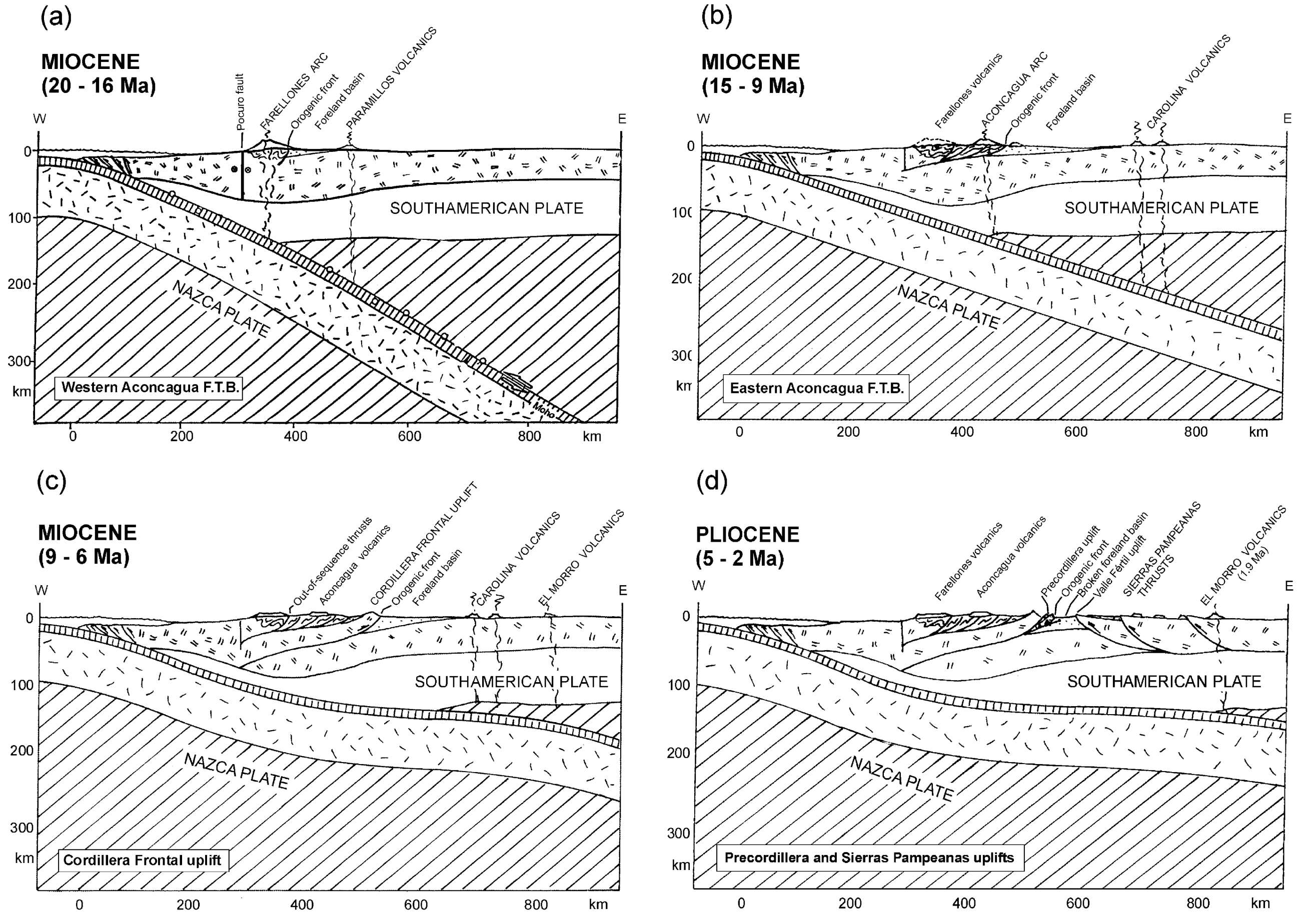


Fig. 10. Late Cenozoic stages of Andean evolution associated with the shallowing of Nazca plate subduction. Geometries of the Wadati–Benioff zone are constrained by the location of the volcanic front. (a) Normal subduction prior to Juan Fernandez Ridge collision; (b) initial slab flattening related to the beginning of ridge collision at 11 Ma; (c) maximum deformation, thrust front propagation, and arc migration during shallowing of the subduction zone; and (d) last arc magmatic activity prior to present-day cessation of magmatism. See text for discussion.

et al., 1990). This early Miocene arc was the result of the Farallon plate breakup into the Nazca and Cocos plates and the beginning of the present period of nearly orthogonal convergence (Cande and Leslie, 1986). The present-day direction of underthrusting is N76°E at 85 mm/a (Yañez et al., 2001; Laursen et al., 2002).

Deformation at this time was concentrated along the axis of the cordillera and the inner and westernmost area of the Aconcagua fold and thrust belt. The retroarc areas were not yet deformed, and a series of andesitic domes and subvolcanic bodies were emplaced in the Precordillera at the Paramillos region. K/Ar dating of unaltered rocks vary from 18.9 ± 0.7 to 16.0 ± 0.8 Ma (Ramos et al., 1991). Recent dating of the alteration indicates a range between 17.8 ± 0.9 and 15.2 ± 0.51 Ma, timing, when the magmatism ceased in the retroarc. These rocks have a geochemical signature typical of a retroarc setting (Kay et al., 1991). This activity was coeval with the main Andean arc and was possibly related to a second slab dehydration front in the subduction zone (Fig. 10(a)).

Synorogenic deposits were partially preserved in the depression located between the Principal and Frontal Cordilleras. Almost no record of this sedimentation span exists in the present foothills, and most of the synorogenic deposits were subsequently cannibalized and resedimented further to the east.

4.2. Early stage (middle Miocene 15–9 Ma)

The first migration of the volcanic arc eastward toward the Sierras Pampeanas is recorded in the volcanics of the Aconcagua region. Huge amounts of andesitic and dacitic rocks were erupted at approximately 15.8 ± 0.4 and 8.9 ± 0.5 Ma in the Aconcagua massif. This area constitutes the new volcanic front 50 km east of the Farellones arc. The retroarc magmatism of Paramillos (Fig. 10(a) and (b)) was shut off at 15.2 Ma, almost concurrent with arc migration. Retroarc magmatism is known in the Sierra de San Luis from 9.5 to 8.2 Ma. The Carolina volcanics were partially coeval with the last volcanic episodes in the Aconcagua arc (Fig. 10(b)).

This shifting of the magmatism was preceded by a pulse of deformation in the western half of the Aconcagua fold and thrust belt (Ramos et al., 1996a). The shortening of the thin-skinned Aconcagua fold and thrust belt, which was detached in Jurassic evaporites, ceased at approximately 8.6 Ma. As a result, the orogenic front migrated through a deeper detachment level, from the Principal Cordillera (Las Cuevas) to the Frontal Cordillera (Punta de Vacas), about 25 km toward the east. This gives a propagation rate of 2.5 mm/a, which is a low rate, when compared with data presented by DeCelles and DeCelles (2001). The shortening rate in the Principal Cordillera at this time was approximately 5.5–5.75 mm/a, according to balanced cross-section data (Cegarra and Ramos, 1996).

Synorogenic deposits were preserved in isolated exposures between the Principal and Frontal Cordilleras, in the Uspallata–Calingasta depression, and in the present foreland basin in the foothills around Mendoza (Fig. 1). Magnetostratigraphic studies performed in the foothills show that sedimentation started at 15.7 Ma in the Mariño Formation with distal fluvial and eolian deposits that record a low sedimentation rate of 0.22 mm/a. This unit was followed by conglomerate and sandstone of the La Pilona Formation at 11.7 Ma, which exhibit a marked increase in accumulation rate with time. The base of the sequence started at 0.17 mm/a and increased progressively to 0.33, 0.77, and 0.95 mm/a (Irigoyen, 1997; Irigoyen et al., 2000), which shows the increase of flexural bending as the thrust front migrated to the east toward the basin.

4.3. Middle stage (late Miocene, 9–6 Ma)

Magmatic arc activity is concentrated in the Sierras Pampeanas. Volcanic arc rocks are widespread between 8 and 6 Ma, all across Sierra de San Luis. No magmatic activity occurs in the Principal and Frontal Cordilleras (Fig. 10(c)).

This stage records the cessation of thin-skinned contraction in the Principal Cordillera at about 8.6 Ma and the uplift of the Frontal Cordillera ca. 9 Ma. Frontal Cordillera uplift was controlled by thick-skinned thrusts detached at much deeper levels than the Aconcagua fold and thrust belt. The décollement is located within the basement at 16–20 km depth (Kozlowski et al., 1993). The thrust front shifted from the Frontal Cordillera (Punta de Vacas) to the western Precordillera (Uspallata depression), about 40 km, which implies a propagation rate of 13.3 mm/a to the east.

The foreland basin in the foothills records the cannibalization of the inner sections and thick accumulation of tuffs and conglomerates derived from the Principal and Frontal Cordilleras uplifts. Sedimentation rates are poorly constrained at ca. 0.41 mm/a in the Tobas Angostura and the Río de Los Pozos Formations, deposited between 9 and 5.5 Ma (Irigoyen, 1997).

4.4. Late stage (Pliocene, 5–2 Ma)

The only volcanic activity in this region in the Pliocene is restricted to the eastern part of the Sierras Pampeanas before the final extinction of magmatism in the flat-slab segment at Sierra del Morro (1.9 Ma).

This stage coincides with the beginning of the uplift of the Precordillera. Striking differences in timing of deformation of the Precordillera occurred from north to south, as documented by Jordan et al. (1997). The inception of Precordillera thrust sheets at Las Juntas (29°S) was at approximately 16 Ma, in Río Azul (30°30′S) at 10 Ma, along Río San Juan (31°30′S) at 8 Ma (Vergés et al., 2001), and at less than 5 Ma at about 33°S latitude. The inception of thrusting in the Precordillera was about 3.5 Ma at 33°30′S, the estimated age of the base of the Mogotes Formation (Irigoyen, 1997). This proximal synorogenic deposit is related to the uplift of the Precordillera. The eastward migration of the thrust front 55 km to the present foothills implies a propagation rate between 13.7 and 9.1 mm/a.

The Sierras Pampeanas of Sierra de San Luis, El Morro, and Comechingones were uplifted during late Pleistocene–early Quaternary times (Fig. 10(d)), as indicated by a marked angular unconformity between distal foreland facies derived from the Andes and proximal facies derived from Sierra de San Luis (Costa, 1992).

4.5. Present setting

The cessation of magmatism at 1.9 Ma coincides with the formation of the present structure. Active faulting is concentrated at the orogenic front of the Precordillera and its southern extension in the Cuyo basin (Fig. 1). This area has gentle, large-scale structures, such as the basement controlled the Barrancas anticline (Fig. 1), whose growing strata records a steady contraction during most of the Quaternary. This structure concentrates important seismic activity, such as big earthquakes and aftershocks (Chiaramonte et al., 2001). The Precordillera has an active thrust front, where several hypocenters of large earthquakes are located. The city of Mendoza was destroyed by one of these large earthquakes in 1864 (Bastías et al., 1993). On the eastern side of the Andean thrust front, most of the Sierras Pampeanas record active Holocene faulting and intense seismicity. Deformation is mainly partitioned by the western faults of the Sierras de San Luis and Comechingones (Costa and Vita-Finzi, 1996).

5. Discussion

5.1. Effects of slab flattening in the main Andes

The structural evolution of the Andes at 33°S at the southern edge of the present flat-slab shows, in the last 20 Ma, not only changes in the style of deformation, but also important variations in the eastward propagation rate of the thrust front. Deformation began with a slow eastward propagation rate of 2.5 mm/a in the Principal Cordillera in the Aconcagua thin-skinned belt, with a shortening rate of

5.5–5.75 mm/a. This fold and thrust belt has a décollement in the Jurassic evaporites, which is 4–5 km deep (Cegarra and Ramos, 1996).

The migration of the thrust front to the east of the Frontal Cordillera, which implies a deeper décollement during basement-involved thrusting, shows a rapid increase in propagation rate of the thrust front from 2.5 to 13.3 mm/a. This rapid increase in propagation rate is associated with a rapid shifting of the magmatic activity toward the foreland. At about 8.6 Ma, the arc volcanism ceased in the Principal Cordillera, almost the time, when magmatism began in the Sierras de San Luis to the east (Fig. 4(d)).

This magmatic shifting of the arc linked to the rapid eastward retreat of the asthenospheric wedge is interpreted as a clear indication of the shallowing of the subduction zone. On the basis of the angular unconformity that separates the volcanic pile from deformed Mesozoic rocks in the Aconcagua region (Fig. 1; Ramos et al., 1996a), it is evident that deformation occurred first, followed by an eastward shift of arc magmatism. The early synorogenic deposits predate the volcanic activity, but when the basin is overfilled at the last stage, the late synorogenic deposits interfinger with volcanic and pyroclastic eruptions.

The rapid eastward shifting of deformation to the Precordillera thrust front with propagation rates of 13.7–9.1 mm/a and the sedimentary response to the deformation were not accompanied by magmatism in the Andes to the west. A small shift in magmatism is recorded between western and eastern Sierra de San Luis. Shallowing of the subduction zone thus appears to have decoupled the asthenospheric wedge from the main Andes.

The timing and geometry of the eastward wave of deformation among the Principal and Frontal Cordilleras and the Precordillera fit a model in which the Nazca plate shallowed after the arrival of the bend of the Juan Fernández aseismic ridge at the Chile margin at ca. 11 Ma (Yañez et al., 2001). We consider that the rapid propagation of the thrust front is a response to the beginning of the Nazca plate shallowing. If this is accepted, the beginning of the flattening is associated with the cessation of arc magmatism in the Principal Cordillera and the uplift of the Frontal Cordillera at about 9 Ma on the Argentine side.

The rapid increase of deformation in late Miocene also could be triggered by collapse and basin inversion of the Oligocene–early Miocene extensional basin, as proposed by Godoy et al. (1999), on the Chilean side.

This rapid increase of deformation marks a change in the propagation rate of the thrust front from 2.5 to 13.1 mm/a at about 9 Ma. Furthermore, an increase in subsidence in the foreland basin, as detected at this critical time by Irigoyen (1997) and Irigoyen et al. (1998), is remarkable. Accumulation rates show an increase from 0.33 to 0.77 and 0.95 mm/a at this time, indicating a rapid eastward migration of the thrust front.

The tectonic evolution of the greater Andean mountain system along the 33°S can be compared to the evolution depicted at 31°S by Kay and Abruzzi (1996). Although the deformation at 31°S migrates toward the foreland at the same time as the magmatism, remarkable differences occurred in timing and geometry of the structure. Structural styles in the Principal Cordillera changed in the north to a thick-skinned belt, with minor shortening (Ramos et al., 1996c), and the timing of deformation in the Precordillera is older to the north (Kay et al., 1991; Jordan et al., 1997). Although part of these differences can be related to variations in crustal rheology controlled by the older rock fabrics and history, such as the existence of Mesozoic rift systems or different basement terranes, the difference in timing may indicate an earlier shallowing at 31°S. The relation between geochemical characteristics and the age of the Miocene magmas and the increase in contamination related to thickening of the crust as a result of thin-skinned shortening in Precordillera at 31°S were described by Kay and Abruzzi (1996). Deformation of the Precordillera at this latitude began at an earlier time and had a history of expansion of magmatic activity different from the southern part. This general youthening trend to the south was recognized in the entire Central Precordillera by Vergés et al. (2001).

The transition from the flat-slab segment at 33°30′ to a steeper subduction segment coincides with the southern end of the Precordillera fold and thrust belt (Jordan et al., 1983b). Because the early Ordovician carbonate platform that controls the thin-skinned fold and thrust belt continues several 100 m to the south of this end, the truncation of the Precordillera is not related to a paleogeographic change. In addition, north and south of this limit, there is a Triassic rift (Fig. 1) that was partially inverted and preserved in the Cuyo basin immediately south of the Precordillera, but was severely inverted further north within the Precordillera. Other features of this transition are discussed by Giambiagi and Ramos (2002).

5.2. *Effects of slab flattening in the Sierras Pampeanas*

The analysis of the four different transects shows a striking relation between the timing of magmatism and the uplift of the Sierras Pampeanas basement. These relations were also described in the Principal Cordillera (Ramos et al., 1996a) and the Precordillera fold and thrust belts (Kay and Abruzzi, 1996). However, relationships in these belts are not as clear, because thin-skinned belts in both areas were detached within a sedimentary sequence, obliterating the behavior of the basement.

In the Sierras Pampeanas, the eastward shifting of the magmatism from 12 to 8.5 Ma and the peak of volcanic activity at 6.78 Ma in the western foothills of the Sierra de Aconquija may have increased the heat flow in this area considerably. The heat flow produced by the existence of an astenospheric wedge results in a surface heat flow up to 70 mW/m^2, according to the Andean thermal model proposed by Springer (1999). Numerical modeling has shown the importance of increase in heat flow in positioning the depth of the brittle–ductile transition in the crust (Kusznir and Park, 1986). Although no heat flow studies have been conducted at these latitudes, if we take into consideration the gradients in present surface heat flow measured by Springer and Förster (1998) further north, it is reasonable to assume a heat flow of about 50 mW/m^2 for the Sierras Pampeanas, prior to the shifting of the magmatism, as indicated in present times by Hamza and Muñoz (1996). At the present heat flow regime of the crust, these conditions will not result in a brittle–ductile transition in the extra-Andean foreland. With an increase to 60–80 mW/m^2, as is seen further north, brittle–ductile transitions could form between 15 and 20 km depth in a typically thick continental crust (Allmendinger, 1987). This will not require an increase of the regional stress to obtain brittle failure at these depths, as strain rates increase with temperature (Park, 1988).

When all the crustal sections are compared (Fig. 4(a), (d)), it is remarkable that the double-wedge thrusting developed in the northern sections at about 15–20 km depth between the upper and lower crust at the thrust front of the Sierras Pampeanas. Farther south, a basement wedge occurred in the Andean thrust front at approximately the same depths. Some of the décollments surfaces are constrained by deep seismic reflection data, as in the Aconquija (Fig. 4(a); Cristallini et al., 2001) and the Cerro Salinas (Fig. 4(d); Comínguez and Ramos, 1991). This last locality provides data indicating a fossil brittle–ductile transition at ca. 14 km that dips gently to the west from the thrust front. This décollement level deepens westward beneath the Precordillera. This kind of brittle–ductile transition can be correlated with the low velocity zones described by Yuan et al. (2000) between 20 and 24°S beneath the Andes on the basis of receiver function images obtained by converted seismic phases. The low velocity zones can be interpreted as brittle–ductile transitions that serve as intracrustal décollments to decouple upper crust brittle structures from ductile lower crust deformation. These lower velocity zones are enhanced by magmatic activity, as detected in the receiver function images over huge Andean calderas (Chmielowski et al., 1999) or in some regions of the Puna–Altiplano (Yuan et al., 2000).

On the basis of this information, it is evident that the depth to the brittle–ductile transitions beneath the Andes appears to be correlated with magmatic activity and the location of crustal wedges in the thrust front of the Sierras Pampeanas. Therefore, the timing of failure of the foreland basement is mainly controlled by the inception of magmatism and subsequent increase of heat flow in the crust. In the different transects, a residence time of the magmatism is required prior to the initiation of crustal failure. In the Aconquija area, failure occurred at about 4 Ma; in the Famatina area, 2.2 Ma; in the Pocho area, 2.4 Ma; and in the Sierra de San Luis, between 6.9 and 3.7 Ma. Brittle–ductile failure occurred during the peak or immediately after the cessation of volcanic activity, consistent with a model that shows that thermal effects associated with concentrated volcanic activity vanish within 1 Ma (Springer, 1999).

From north to south, the beginning of tectonic uplift within each of the analyzed sections varies from 7.6 to 6 Ma in the Aconquija, 4.5 to 4.19 Ma in the Famatina, 5.5 to 4.7 Ma in Pocho, and ca. 2.6 Ma in San Luis. These dates indicate a rough decrease in age to the south, if it is taken into consideration that these areas are not equally distant from the trench.

These data confirm the assumption of James and Sacks (1999), who suggest that an increase of compressional stress is not always necessary to produce Andean deformation, as the latter can occur in response to local lithospheric thermal weakening. This circumstance can explain why, at different localities, regional deformation occurs as soon as the magmatic activity weakens the lithosphere and allows a brittle–ductile transition zone to form in the crust.

These observations rule out the need for pulses of deformation within episodic orogenic phases related to increases in the stress field. Time periods of more rapid convergence are longer than those associated with local deformation pulses, and therefore, fluctuations in plate velocities alone cannot adequately explain the observed deformation patterns. When precise dating of deformation is available, the apparent synchronicity of deformation vanishes, as seen in the different transects of Sierras Pampeanas.

The major control on the uplift of the Sierras Pampeanas is the anisotropy of previous weakness zones, such as normal faults, sutures, or other discontinuities, that will be enhanced by the increase of thermal flow. Heating will increase as magmatism migrates toward the foreland during shallowing of the subduction zone. Because this migration is controlled by the timing and geometry of the subducting Nazca plate, it is evident that the wave of deformation will move eastward into the foreland. A similar trend of southward migration also follows the shallowing of the Wadati–Benioff zone, which can be correlated with the southward shifting of the Juan Fernández ridge collision along the trench (Von Huene et al., 1997; Yañez et al., 2001).

6. Conclusions

The structural evolution, magmatic activity, and sedimentation of synorogenic deposits of the main Andes and associated Sierras Pampeanas structural province to the east match the timing and location of the shallowing subducting Nazca plate. However, at different latitudes, the tectonic responses to this shallowing vary according to the rheologic characteristics of the crust at each segment. The Andes as a whole has different structural styles and behaviors, mainly controlled by the previous crustal history and distribution and timing of the magmatic activity.

The relation between migration of the magmatic arc to the foreland, the associated increase in heat flow, and the development of brittle–ductile transitions within the crust seem to control the timing and location of Sierras Pampeanas uplifts. Thus, the diachronism of local uplifts can be explained better by these factors than by stress variations induced by changes in plate boundary conditions.

Acknowledgements

The authors acknowledge funding received from PICT 6729 of Agencia Nacional de Promoción Científica y Tecnológica to support this research. Special thanks to Dr Suzanne M. Kay (Cornell University, USA) and Estanislao Godoy (Sernageomin, Chile) for many fruitful discussions on different aspects of the tectonic evolution of the Pampean flat-slab segment of Argentina and Chile. The reviews and constructive comments by David James, Brendan McNulty, and Dave Scholl are appreciated.

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