Paleozoic to early Cenozoic cooling and exhumation of the basement underlying the eastern Puna plateau margin prior to plateau growth

N. Insel, M. Grove, M. Haschke, J. B. Barnes, A. K. Schmitt, and M. R. Strecker

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[1] Constraining the pre-Neogene history of the Puna plateau is crucial for establishing the initial conditions that attended the early stage evolution of the southern extent of the Andean plateau. We apply high- to low-temperature thermochronology data from plutonic rocks in northwestern Argentina to quantify the Paleozoic, Mesozoic and early Tertiary cooling history of the Andean crust. U-Pb crystallization ages of zircons indicate that pluton intrusion occurred during the early mid-Ordovician (490–470 Ma) and the late Jurassic (160–150 Ma). Lower-temperature cooling histories from \(^{40}\text{Ar}/^{39}\text{Ar}\) analyses of K-feldspar vary substantially. Basement rocks underlying the western Puna resided at temperatures below 200°C (<6 km depth) since the Devonian (~400 Ma). In contrast, basement rocks underlying the southeastern Puna were hotter (~200–300°C) throughout the Paleozoic and Jurassic and cooled to temperatures of <200°C by ~120 Ma. The southeastern Puna basement records a rapid cooling phase coeval with active extension of the Cretaceous Salta rift at ~160–100 Ma that we associate with tectonic faulting and lithospheric thinning. The northeastern Puna experienced protracted cooling until the late Cretaceous with temperatures <200°C during the Paleocene. Higher cooling rates between 78 and 55 Ma are associated with thermal subsidence during the postrift stage of the Salta rift and/or shortening-related flexural subsidence. Accelerated cooling and deformation during the Eocene was focused within a narrow zone along the eastern Puna/Eastern Cordillera transition that coincides with Paleozoic/Mesozoic structural and thermal boundaries. Our results constrain regional erosion-induced cooling throughout the Cenozoic to have been less than ~150°C, which implies total Cenozoic denudation of <6–4 km.

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1. Introduction

[2] Orogenic plateaus are some of the most prominent and enigmatic features on Earth [e.g., Isacks, 1983; Fielding et al., 1994; Royden, 1996; Allmendinger et al., 1997; Hatzfeld and Molnar, 2010; Schildgen et al., 2012]. One of the least constrained, yet crucial parameters required for understanding orogenic plateau development is the thermal and mechanical state of the crust both prior to and during the earliest phases of plateau formation [Vanderhaeghe, 2012]. Previous studies in the central Andes have argued that the onset, spatial pattern, and style of lithospheric deformation was controlled by lithospheric strength and temperature [Whitman et al., 1996; Babeyko and Sobolev, 2005] as well as preexisting crustal heterogeneities [e.g., Allmendinger et al., 1983; Grier et al., 1991; Allmendinger et al., 1997; Kley et al., 1999; Carrera et al., 2006; Oncken et al., 2006; Hongn et al., 2007]. In particular, lithospheric weakening through magmatism and shortening is considered to have been a controlling factor for explaining variations in Altiplano-Puna plateau deformation during the early to mid-Cenozoic [Isacks, 1983; Wdowinski and Bock, 1994]. Changes in Cenozoic exhumation rates have been documented throughout the central Andes [e.g., Elger et al., 2005; Ege et al., 2007; Carrapa and DeCelles, 2008;
What is known is that the present boundary between the Puna plateau and the Eastern Cordillera has a long history of regional tectonism and emplacement of granitoid magmas (Figure 1). Subduction along the Pacific coast of South America has triggered episodes of magmatism, basin formation, and tectonics in the central Andes since the early mid Ordovician (~490–470 Ma) [Coira et al., 1982]. Previous studies have mostly concentrated on detailed petrological-geochemical analyses of single plutons to understand their genesis [e.g., Coira et al., 1999; Poma et al., 2004; Kirschbaum et al., 2006; Hongn and Riller, 2007] and established the geologic evolution of the central Andes during the early Paleozoic [e.g., Coira et al., 1999]. The genesis, ascent and emplacement of Ordovician magmas along the present-day Puna and Eastern Cordillera boundary occurred in a complex back-arc setting that was associated with thermal anomalies and lithospheric thinning, and characterized by transpressional shortening episodes in an overall extensional regime [Coira et al., 1999; Kirschbaum et al., 2006; Coira et al., 2009b]. Host rocks consisted of different tectonometamorphic domains with distinct metamorphic grades, rheologic properties [Coira et al., 1982; Coira et al., 1999], and linked ductile deformation zones that could have facilitated the ascent of magmas at deeper crustal levels [Hongn and Mon, 1999; Hongn and Riller, 2007].

Recent studies of the Puna plateau, its eastern flanks, and the adjacent broken foreland suggest an irregular Cenozoic deformation pattern controlled by inherited crustal discontinuities [e.g., Hongn et al., 2007]. Thus, the preexisting Paleozoic structural and stratigraphic heterogeneities may have influenced magmatism and subsequent deformation phases [e.g., Mon and Hongn, 1991; Hongn and Riller, 2007; Carrera and Muñoz, 2008; Wegmann et al., 2008; Hongn et al., 2010; Pearson et al., 2012]. For example, the trend of Cretaceous intracratonic rift structures in northern Argentina is similar to the preexisting Paleozoic crustal anisotropies [e.g., Galliski and Viramonte, 1988; Grier et al., 1991]. Moreover, many Mesozoic rift structures were reactivated and inverted during the Cenozoic Andean orogeny [Grier et al., 1991; Kley et al., 2005; Mon et al., 2005]. In the Puna plateau region Mio-Pliocene volcanic

Figure 1. Geologic map of part of the Puna plateau margin in northwest Argentina showing the distribution of Ordovician and temporally related magmatic and sedimentary units (modified from Rapela et al. [1992] and Coira et al. [1999]). Map shows principle faults, salars (white areas) and cities (open circles). Generalized boundaries between the Western Puna Eruptive Belt, Eastern Puna Eruptive Belt, and Eastern Cordillera are shown as heavy dashed lines [after Coira et al., 1999]. Black squares are sample locations of plutonic rocks (MAC = Sierra de Macón, CO = Complejo Oire, VL = Valle Luracatato, COL = Colomé, CD = Cerro Durazno, GA = Aguilar granite, GF = Fundiciones granite, COB = Cobres granite, AC = Nevado de Acay). OETL, Olacapato–El Toro lineament. Inset is a shaded 90 m SRTM DEM of the central Andes showing major morphological units (WC = Western Cordillera, AP = Altiplano, EC = Eastern Cordillera, IA = Interandean zone, SA = Subandes) [e.g., Troëng et al., 1993; Kley, 1996] and location of study area (black box).
centers are concentrated along inherited NW oriented shear zones and magmas have exploited these zones of weakness to reach the surface [e.g., Kay et al., 1994; Riller and Onccken, 2003].

[5] Unraveling the lower Paleozoic magmatism and thermal effects of successive tectonic phases throughout the Phanerozoic is key for understanding the geological and thermal history of the basement rocks that constitute the Puna plateau. In this study, we combine new U-Pb geochronology and \(^{40}Ar/^{39}Ar\) thermochronology to constrain the Paleozoic and Mesozoic cooling histories of basement rocks along the eastern Puna/Eastern Cordillera boundary. We provide new age constraints on Paleozoic, Mesozoic and Tertiary plutons in the northern part of the Puna/Eastern Cordillera transition zone. By integrating our results with previously reported low-temperatureapatite fission track (AFT) results [Deeken et al., 2006] we (1) provide quantitative constraints on sample time-temperature histories from crystallization (~850°C) to closure of the K-Ar system (~350–150°C); (2) use our results to interpret differences in the amount of pre-Cenozoic denudation of the crust across different domains; and (3) explore the influence of inherited zones of crustal weakness on subsequent deformation phases. In particular, we present new age and uplift constraints for the Paleozoic, Mesozoic and early Cenozoic that influenced the locus of Cenozoic crustal deformation processes and the initial thermomechanical conditions for orogenic plateau growth.

2. Geological Setting

[6] The Altiplano-Puna plateau covers a region of ~500,000 km\(^2\) with an average elevation of ~4 km across the central Andes [Isacks, 1988]. In this study, we focus on the structural transition from the eastern Puna margin to the Eastern Cordillera between 23\(^\circ\)S and 26\(^\circ\)S (Figure 1). At the present-day boundary between the Puna and the Eastern Cordillera, magmatic rocks comprise two NNW-SSE striking magmatic belts known as the Faja Eruptiva de la Puna Occidental in the west [Palma et al., 1986] and Faja Eruptiva de la Puna Oriental in the east [Méndez et al., 1973]. We refer to these two zones as the western and eastern Puna eruptive belts [e.g., Coira et al., 2009b; Coira et al., 2009a]. In addition, lower Paleozoic magmatic units are exposed farther east in the Eastern Cordillera (Figure 1). Most of the magmatism that characterizes the basement rocks of these geological provinces occurred in the lower to mid Ordovician (~490–460 Ma [e.g., Coira et al., 2009b, and references therein].

[7] Lower Paleozoic magmatic, sedimentary and metamorphic rocks are widely distributed throughout the study area. The western Puna eruptive belt includes granite-granodiorite-monzodiorite plutons that record three main magmatic episodes between 502 and 476 Ma, 450–440 Ma, and 420–425 Ma [Mpozos et al., 1983; Palma et al., 1986; Koukharsky et al., 2002; Kleine et al., 2004]. The northern part of the eastern Puna eruptive belt is defined by greenschist-grade metasediments and the magmatic-sedimentary Cochinoca-Escaya Complex (Figure 1) [Coira et al., 1999] with granitoids comprising ages of ~480–470 Ma [Lork and Bahlburg, 1993; Kirschbaum et al., 2006]. To the south, the Cochinoca-Escaya Complex is cut by undifferentiated higher-grade metamorphic assemblages [Coira et al., 1982; Damm et al., 1990]. Extensive magmatic units are exposed with ages of 500–462 Ma, 450–440 Ma, and 420 Ma [Linares, 1979; Lork and Bahlburg, 1993; Lucassen et al., 2000]. Paleozoic pressure-temperature records suggest transient high temperatures (>650°C) and pervasive partial melting of the crust at ~15–25 km depths at that time [Lucassen and Franz, 2005]. Farther east, Ordovician granitoids in the Eastern Cordillera mainly intruded into upper Precambrian to Cambrian turbiditic sequences of the Puncoviscana Formation [Mon and Salfity, 1995; Coira et al., 1999]. Ordovician units were emplaced syntectonically and characterized by linked NW-SE and N-S striking ductile shear zones and NW dipping metamorphic foliations [Büttner et al., 2005; Hongn and Riller, 2007]. Deformation processes continue into the Silurian [Turner and Méndez, 1975; Bahlburg, 1990; Wegmann et al., 2008] as part of mountain building processes at the western margin of Gondwana that included complex contraction and extension associated with left-lateral shearing [Mon and Hongn, 1991; Bahlburg and Herve, 1997; Zimmermann and Bahlburg, 2003; Hongn and Riller, 2007; Coira et al., 2009b].

[8] Posttectonic Devonian-Carboniferous alkaline granitoids have been reported in the eruptive belts [Rapela et al., 1992]. In the Carboniferous, anorogenic magmatism affected a large region of the present-day Argentine Sierras Pampeanas south of the study area [Dahlquist et al., 2010]. Most of these granitoids were emplaced at high crustal levels along prominent older shear zones [Höckenreiner et al., 2003; Dahlquist et al., 2006] and intruded into host rocks of lower Paleozoic age [e.g., Rapela et al., 1990; Sims et al., 1998; Pankhurst et al., 2000; Grosse et al., 2009; Dahlquist et al., 2010]. The generation of Carboniferous magmas is consistent with a regional reheating of the crust at ~350–335 Ma [Grissom et al., 1998; Höckenreiner et al., 2003; Dahlquist et al., 2006; Alasino et al., 2012]. This period was followed by cooling and uplift in the late Carboniferous [Höckenreiner et al., 2003; Dahlquist et al., 2010]. Although Carboniferous magmatism is not evident in our study area, late Paleozoic transient heating of rocks in our study area associated with the anorogenic event is a possibility.

[9] An intracratonic rift system developed in northern Argentina during Cretaceous-Paleogene time (~160–60 Ma) [Salfity, 1982; Galliski and Viramonte, 1988; Salftity and Marquillas, 1994; Kley et al., 2005]. Rift-related igneous rocks are mainly hornblende-biotite granodiorites/granites that intrude early Cambrian (Puncoviscana Formation) or Ordovician sediments. The Salta rift was characterized by a series of NW to NE striking normal faults that bound horsts and grabens [e.g., Galliski and Viramonte, 1988; Grier et al., 1991]. Many of the present-day faults that bound the igneous basement ranges investigated here are reactivated structures inherited from the Salta rift system [Grier et al., 1991; Mon et al., 2005]. Thus, Cretaceous plutons often occupy the hanging walls of thrust faults. Previous studies place the onset of rift inversion in the Puna region to Eo-Oligocene time (40–30 Ma) [Andriessen and Reutter, 1994; Coutard et al., 2001; Kley and Monaldi, 2002], hence, prior to the superseeding major phases of Andean deformation, uplift and exhumation that occurred during the Miocene in the eastern Puna [e.g., Carrapa et al., 2005; Coutard et al., 2006]. Deformation in the Puna remains active within the Plio-Quaternary [Allmendinger et al., 1989; Cladouhos et al., 1994; Marrett et al., 1994; Schoenbohm and
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Table 1. Sample Locations, U-Pb, and $^{40}$Ar/$^{39}$Ar Analysesa

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Latitude (°)</th>
<th>Longitude (°W)</th>
<th>Altitude (m)</th>
<th>Age ± 1σ SD (Ma)</th>
<th>MSWDb</th>
<th>Number of Grains</th>
<th>J Factorc</th>
<th>Total Gas Age ± 1σ SD (Ma)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MAC</td>
<td>Sierra de Macond</td>
<td>24°35.687</td>
<td>67°19.585</td>
<td>4000</td>
<td>473.73 ± 4.33</td>
<td>0.96</td>
<td>7</td>
<td>0.003538</td>
<td>420.34 ± 6.20</td>
</tr>
<tr>
<td>CO</td>
<td>Complejo Oired</td>
<td>25°00.345</td>
<td>66°41.941</td>
<td>3990</td>
<td>481.90 ± 4.19</td>
<td>0.80</td>
<td>5</td>
<td>0.003541</td>
<td>317.51 ± 4.63</td>
</tr>
<tr>
<td>VL</td>
<td>Valle Lucacats</td>
<td>2°38.095</td>
<td>66°29.540</td>
<td>3150</td>
<td>491.20 ± 4.35</td>
<td>1.29</td>
<td>6</td>
<td>0.003539</td>
<td>237.69 ± 3.78</td>
</tr>
<tr>
<td>COL</td>
<td>Colomee</td>
<td>25°36.269</td>
<td>66°24.597</td>
<td>2570</td>
<td>468.80 ± 3.33</td>
<td>1.52</td>
<td>5</td>
<td>0.003544</td>
<td>293.02 ± 4.32</td>
</tr>
<tr>
<td>CD</td>
<td>Cerra Durzanoe</td>
<td>25°39.292</td>
<td>66°20.971</td>
<td>2800</td>
<td>488.10 ± 3.12</td>
<td>0.78</td>
<td>7</td>
<td>0.003543</td>
<td>306.25 ± 4.48</td>
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<td>COB</td>
<td>Cobresf</td>
<td>23°38.623</td>
<td>66°17.604</td>
<td>3800</td>
<td>478.40 ± 3.46</td>
<td>4.04</td>
<td>7</td>
<td>0.003549</td>
<td>172.28 ± 2.65</td>
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<tr>
<td>GF</td>
<td>Fundiciones granitet</td>
<td>23°23.526</td>
<td>66°25.813</td>
<td>3972</td>
<td>160.40 ± 1.20</td>
<td>0.68</td>
<td>7</td>
<td>0.003546</td>
<td>161.78 ± 2.66</td>
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<tr>
<td>GA</td>
<td>Aguilar granitet</td>
<td>23°12.856</td>
<td>65°41.217</td>
<td>4050</td>
<td>150.40 ± 0.91</td>
<td>2.67</td>
<td>7</td>
<td>0.003547</td>
<td>94.68 ± 1.55</td>
</tr>
<tr>
<td>AC</td>
<td>Nevado de Acayc</td>
<td>24°24.547</td>
<td>66°10.688</td>
<td>4986</td>
<td>12.61 ± 0.25</td>
<td>1.25</td>
<td>7</td>
<td>0.003545</td>
<td>15.17 ± 0.93</td>
</tr>
</tbody>
</table>

1Adjusted for atmospheric argon and nucleonic interferences $^{40}$Ar/$^{39}$ArK = 0.0285; $^{40}$Ar/$^{39}$ArCa = 0.00018; $^{40}$Ar/$^{39}$ArCa = 0.00057.
2Mean square of weighted deviates.
3Irradiation factor.
4Corrected for $^{26}$Ar.$^{20}$Po.

Strecke, 2009]. During the middle late Miocene, eruptions associated with stratovolcanoes along the modern magmatic arc were controlled by inherited Paleozoic northwest striking shear zones [Coira et al., 1982; Almendinger et al., 1983; Alonso et al., 1985; Salfity, 1985].

3. Methods

3.1. Sampling

[10] We sampled broadly to constrain the large-scale spatial variations in basement thermal histories. We targeted well-exposed early Paleozoic plutonic rocks and overlying strata along the eastern Puna/Eastern Cordillera boundary region because they are potentially capable of preserving pre-Andean cooling signals (Table 1 and Figure 1). In general, samples were located close to the base of mountain ranges at 2.5–5.0 km elevations. We analyzed Ordovician basement rocks from the western Puna eruptive belt (Sierra de Macón = MAC), the eastern Puna eruptive belt (Cobres granite = COB, Complejo Oire = CO, Valle Lucacatao = VL, Colome = COL), and the Eastern Cordillera (Cerro Durzano = CD). We also analyzed two Cretaceous samples from the northern Puna/Eastern Cordillera boundary (Aguilar granite = GA, Fundiciones granite = GF), and one Tertiary sample along a NW-SE trending crustal lineament (OETL in Figure 1) in the Eastern Cordillera (Nevado de Acay = AC). For each sample, we collected ~3 kg of rock and isolated zircon and K-feldspar crystals using standard crushing, sieving, and density/magnetic separation techniques. We then used a binocular microscope to hand-select specific grains for our analysis.

3.2. Zircon U-Pb Geochronology

[11] The U-Pb isotope system is used to infer crystallization ages of magmatic zircon. Results of U-Pb ion microprobe analysis of 63 zircons from 9 granite samples (5–7 zircons per sample) are summarized in Figure 2 and Table 1. Complete methods and tabulated data are presented in the auxiliary material.1 Cathodoluminescence (CL) imaging generally revealed that core-rim overgrowth textures were uncommon. However, we did observe more complex textural relationships on a few individual grains (e.g., VL grain 6; Figure 2c). Our analysis spots were deliberately placed near the grain margins in order to probe the youngest growth layers and to avoid the older zircon interiors. Thus, while older (Grenville) ages documented for the Sierras Pampeanas [e.g., Casquet et al., 2005] were not detected in our analyses, no firm conclusion regarding inheritance of this age should be drawn since we did not measure grain interiors.

3.3. K-Feldspar $^{40}$Ar/$^{39}$Ar Thermochronology

[12] K-feldspar $^{40}$Ar/$^{39}$Ar thermochronology provides the basis for reconstructing the thermal history of the Andean crust [e.g., McDougall and Harrison, 1999]. Determination of thermal histories from $^{40}$Ar/$^{39}$Ar step heating of K-feldspars is based upon MultiDiffusion Domain (MDD) analysis [e.g., McDougall and Harrison, 1999]. In favorable circumstances, MDD modeling has proven capable of yielding continuous thermal histories between 350°C and 150°C [Lovera et al., 1989, 1997, 2002; Harrison et al., 2005]. Application of MDD modeling to $^{40}$Ar/$^{39}$Ar results from K-feldspar requires assumptions about the general form of the temperature-time path taken by any sample. We restrict our interpretation to the simplest possible solution, i.e., monotonic cooling.

[13] Assuming monotonic cooling, calculated thermal histories obtained from $^{40}$Ar/$^{39}$Ar K-feldspar step heating typically define a narrow range of possible temperature–time conditions [Lovera et al., 1997]. MDD modeling approaches have been developed that provide useful constraints on potential nonmonotonic cooling paths such as maximum bounds on peak temperatures associated with reheating [Quideulle et al., 1997; McDougall and Harrison, 1999; Harrison et al., 2000; Lovera et al., 2002]. However, MDD-derived models cannot unambiguously distinguish monotonic from nonmonotonic cooling paths without additional information. With the lack of independent thermal history constraints in our study area, removing the monotonic assumption would significantly increase the range of potential model thermal histories, preclude the straightforward use of closure temperature concepts, and introduce

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1Auxiliary material data sets are available at ftp://ftp.agu.org/apend/tc/2012tc003168. Other auxiliary material files are in the HTML. doi:10.1029/2012TC003168.
ambiguities in modeling continuous time–temperature paths [Reiners, 2009]. Because we have no clear indication of transient heating by nearby plutons or any other sources of crustal heat addition between early Silurian and Cretaceous times, we adopt the assumption of monotonic cooling as the most sensible first-order interpretation of the feldspar 
\[ ^{40} \text{Ar}/^{39} \text{Ar} \]
data.\[14] Some K-feldspars have been shown to yield anomalously old \[^{40} \text{Ar}/^{39} \text{Ar} \] ratios over the first several percent of \[^{39} \text{Ar} \] release, which has been attributed to thermal decrepitation of excess argon \(^{40} \text{Ar}_{\text{E}}\) bearing fluid inclusions [Harrison et al., 1994]. To improve the basis for correcting the low-temperature argon release from these feldspars, we performed isothermal quadruplicates with the low-temperature steps. Since the number of fluid inclusions that burst at a given temperature decreases with progressive heating, the first isothermal step is interpreted to yield an anomalously old age, while the second and successive steps are less affected, and thus yield younger ages. Generally, \(^{40} \text{Ar}_{\text{E}}\) is correlated with CI/K and an age spectrum correction for the effects of low-temperature \(^{40} \text{Ar}_{\text{E}}\) contamination is possible [Harrison et al., 1993, 1994]. However, corrections for CI-correlated \(^{40} \text{Ar}_{\text{E}}\) were ineffective for our samples due to a combination of high \(^{40} \text{Ar}_{\text{E}}/\text{Cl} \) and low CI/K values that resulted in erratic corrected ages because they were overly sensitive to measurement uncertainties in CI/K (i.e., only a small % of measured \[^{38} \text{Ar} \] was CI derived). Thus, to avoid interpretative difficulties in modeling the age spectrum, we did not attempt to fit the lowest-temperature portion of the age spectrum that is characteristically degraded by fluid inclusion-hosted \(^{40} \text{Ar}_{\text{E}}\).

Figure 2. The \(^{206} \text{U}/^{238} \text{U} \) versus \(^{207} \text{U}/^{235} \text{U} \) Concordia diagrams for U-Pb ion microprobe results for zircons from nine different samples. The individual analysis results are plotted as open (1 sigma errors) ellipses, bold ellipses are error ellipses, dashed ellipses represent analyses excluded from calculating the mean age. CL images show grain VL_g6 (Figure 3c) and GA_g6 (Figure 3g).
Another feature often manifested by samples are low-amplitude intermediate age maxima in the age spectra. Such features may result from recrystallization of diffusion domains after the onset of $^{40}$Ar retention [Lovera et al., 2002]. Suspected physical phenomena include low-temperature exsolution or partial alteration to albite and other phases and/or new growth of low-temperature adularia [e.g., Farsons et al., 1999; Lovera et al., 2002]. The impact of these phenomena is that newly formed, small domains inherit the $^{40}$Ar concentrations from their larger precursor domains [see Lovera et al., 2002]. Consequently, thermal history results from affected samples are generally less reliable than those for samples lacking such features. However, in cases where prominent intermediate age maxima occur for our samples (e.g., Sierra de Macón), the uncertainties associated with detailed MDD thermal history analysis is minimized because samples cooled rapidly.

Table 1 summarizes the key $^{40}$Ar/$^{39}$Ar results with correction factors for interfering neutron reactions determined from CaF$_2$ and K$_2$SO$_4$ crystals and calculated J factors. Detailed analytical results for individual analyses are provided in the auxiliary material. For each sample, we calculated 10 best fit MDD parameter sets. Five best fit monotonically decreasing cooling histories were calculated from each set of MDD parameters to yield 50 total solutions. Using these solutions, we determined the 90% confidence limits of the thermal history that correspond to the interpreted portion of the age spectrum.

4. Results

4.1. Zircon U-Pb Data and Analysis

Most of the 63 zircon grain analyses have concordant $^{206}$Pb/$^{238}$U ages (Figure 2). After excluding anomalous age results (e.g., grain 6 for GA and VL, Supp2), weighted mean $^{206}$Pb/$^{238}$U ages were calculated for the remaining analyses. Generally, values of the mean square of weighted deviates (MSWD) fell within the expected range for a homogenous population [Mahon, 1996]. With these samples, we interpret the results to characterize the age of magmatic crystallization within 1σ analytical uncertainties. The two exceptions are samples from the Cobres granite and the Aguilar granite with MSWD values of 4.0 and 2.8, respectively. While the age spread may signal heterogeneity in the zircon population (e.g., caused by inheritance or Pb loss), we adopted a conservative approach of increasing the stated uncertainties on the sample mean age by the square root of the MSWD.

4.1.1. Ordovician Granitoids

The Sierra de Macón is an elongated basement range within the Puna (Figure 1). It is the easternmost granitoid of the Ordovician western Puna belt [Poma et al., 2004]. Our $^{206}$Pb/$^{238}$U data indicate a crystallization age of 473.7 ± 4.3 Ma (Table 1 and Figure 2a). This age is consistent with radiometric ages determined by previous workers ($^{206}$Ar/$^{39}$Ar on hornblende: 482.7 ± 7.8 Ma [Koukharsky et al., 2002]) and agrees with regional age relationships for early Ordovician intrusions [e.g., Coira et al., 1999].

Four samples were obtained from the southern transition between the eastern Puna and the Eastern Cordillera (Figure 1). Our Ordovician samples mostly consist of biotite-hornblende bearing granodiorite and monzogranite and K-feldspar-rich granites that form the eastern flank that forms the western margin of the Puna plateau [Allmendinger, 1986]. Samples yielded weighted mean $^{206}$Pb/$^{238}$U ages between 469 and 491 Ma (481.9 ± 4.2 Ma (Complejo Oire), 491.2 ± 2.8 Ma (Valle Luracatao), 468.8 ± 3.3 Ma (Colomé), 488.1 ± 3.1 Ma (Cerro Durazno); Table 1 and Figures 2b–2e). We also sampled the Cobres pluton along the northeastern Puna margin (Figure 1). The zircons have a weighted mean $^{206}$Pb/$^{238}$U age of 478.4 ± 3.5 Ma (Table 1 and Figure 2f). All of these ages fall into the early to middle Ordovician and correlate with reported ages from the eastern Puna belt [e.g., Omarini et al., 1984; Lork and Balhburg, 1993; Coira et al., 1999; Viramonte et al., 2007; Pearson et al., 2012].

4.1.2. Cretaceous Granitoids

Our Cretaceous samples are located at the northern margin of our study area between the eastern Puna and the Eastern Cordillera and include the Aguilar granodiorite and the Fundiciones granite (Figure 1). Zircon U-Pb weighted mean ages are 150.4 ± 0.9 Ma for the Aguilar granodiorite and 160.4 ± 1.2 Ma for the Fundiciones granite (Table 1 and Figures 2g and 2h). The Aguilar age is significantly older than previously determined Rb/Sr crystallization ages and K-Ar ages of 110–133 Ma [Halpern and Latorre, 1973]. Although one zircon (g6) was clearly inherited (~250 Ma), CL imaging of the grains showed that the ion microprobe analysis sites successfully avoided inherited zircon domains, and hence support a late Jurassic crystallization age. This age is consistent with $^{206}$Pb/$^{238}$U ages of the Abra Laite pluton just west of the Aguilar that has an average age of ~153 Ma as well as the 140–153 Ma old Tausquillas Plutonic Complexes ~35 km to the south [Cristiani et al., 2005].

4.1.3. Tertiary Plutons

The Nevado de Acay pluton is the youngest intrusion we sampled. This biotite- and hornblende-bearing quartz-monzonite intruded low-grade metamorphic rocks of the Punoiviscana Formation and Upper Cretaceous limestones and pelites of the Yacoraite Formation in the Calchaqui Valley (Figure 1). The location of the pluton is thought to have been structurally controlled by the Olacapato–El Toro lineament (OETL) [Llambias et al., 1985]. Eight zircons yielded concordant $^{206}$Pb/$^{238}$U results with a weighted mean age of 12.6 ± 0.3 Ma (Table 1 and Figure 2i). Although previous ages indicated late Oligocene (26 ± 1 Ma [Mirre, 1974; Linares, 1979]) or early Miocene emplacement (18–19 Ma [Petrinovic et al., 1999]), our new younger age is more compatible with similar magmatic rocks in the region [Petrinovic et al., 1999; Mazzuoli et al., 2008; Kay et al., 2010].

4.2. K-Feldspar $^{40}$Ar/$^{39}$Ar Thermal Histories

In the following, we describe sample cooling histories by their locations within the broad tectonic framework (western Puna eruptive belt, southeastern Puna/ Eastern Cordillera boundary, southeastern Puna/Eastern Cordillera transition). Deriving depth estimates is difficult...
due to uncertainties in the paleogeothermal gradient. Modern temperature gradients in boreholes in the central Andes are $39 \pm 9^\circ C/km$ in the plateau area and $25 \pm 8^\circ C/km$ in the Eastern Cordillera [Henry and Pollack, 1988; Springer and Förster, 1998]. Estimated Cenozoic paleogeotherms in the central Andes range between $32$ and $19^\circ C/km$ with errors of $20\text{–}40\%$ [e.g., Carrapa et al., 2006; Deeken et al., 2006; Ege et al., 2007]. Thus, we estimate depths assuming an average paleogeothermal gradient of $25\text{–}35^\circ C/km$, applying the higher geotherm to most of the Paleozoic and Mesozoic estimates and the lower geotherm to Cenozoic times. These geotherms represent temperatures for an equilibrated crust and cannot be applied to intrusions. Thus, estimated crystallization depths are based on (1) the overall cooling history (i.e., how fast the sample cools down after crystallization) and (2) previously estimated crystallization pressures and temperatures in the central Andes (i.e., assuming generally shallow intrusion depths at pressure $<0.4$ GPa and/or $4\text{–}8$ km [Kay et al., 2010, and references therein]).

4.2.1. The Western Puna Eruptive Belt

[23] K-feldspar $^{40}$Ar/$^{39}$Ar step heating of the Sierra de Macon sample shows a minor age gradient between 410 and 430 Ma over the last 80% of $^{39}$Ar release (Figure 3a). The initial gas released exhibits an age gradient down to 250 Ma before it is obscured by initial low-temperature $^{40}$Ar$_E$ contamination. Although the detailed temperature-time history is slightly uncertain due to the prominent intermediate age maxima at $\sim10\%$ $^{39}$Ar release, our calculations indicate that it does not seriously impact our ability to extract the MDD thermal history, because the sample cooled rapidly.

[24] MDD modeling results indicate temperatures $<300^\circ C$ by $\sim420$ Ma (Figure 4a and Table 2), suggesting shallow crystallization depths ($<8$ km). Importantly, the Sierra de Macon cooled very rapidly ($\sim3.6^\circ C$/Myr) to $\sim150$ Ma, followed by a long period of thermal crustal stability with slow cooling ($<0.2^\circ C$/Myr). The Sierra de Macon stayed at $150\text{–}120^\circ C$ ($\sim4$ km depth) throughout most of the late Paleozoic and early Mesozoic. AFT data indicate cooling below $\sim115^\circ C$ at $\sim200$ Ma [Deeken et al., 2006].

4.2.2. The Southeastern Puna/Eastern Cordillera Boundary

[25] K-feldspar $^{40}$Ar/$^{39}$Ar data from samples along the eastern Puna eruptive belt indicate a monotonically increasing age spectrum (Figures 3b–3d and 3f). The southern Puna samples indicate cooling to $<300^\circ C$ between 400 and 360 Ma (Figures 4b–4d and Table 2). Temperatures remained above $200^\circ C$ ($>6\text{–}8$ km depth) until Jurassic/Cretaceous time. The Complejo Oire and Valle de Luracatao granites cooled slowly ($0.2\text{–}0.3^\circ C$/Myr) during the Paleozoic and the early Mesozoic. Both samples experienced rapid cooling in the late Jurassic/early Cretaceous with cooling rates increasing from $0.2$ to $0.6^\circ C$/Myr (Complejo Oire) and 0.3 to $1.3^\circ C$/Myr (Valle Luracatao), respectively (Figures 4b and 4c and Table 2). The Colomé sample is...
Figure 4
Table 2. Cooling History

<table>
<thead>
<tr>
<th>Sample</th>
<th>Location</th>
<th>Time (Ma)</th>
<th>Temperature (°C)</th>
<th>Cooling Rate (°C/Myr)</th>
</tr>
</thead>
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<tr>
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<td>0.30</td>
</tr>
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<tr>
<td></td>
<td></td>
<td>16.70</td>
<td>115</td>
<td>0.60</td>
</tr>
</tbody>
</table>

*Bold values indicate accelerated cooling phases.

characterized by slow cooling to temperatures of ~160°C (~4.5–6 km depth) by 134 Ma (Figure 4d).

[26] The Cerro Durazno range is proximal to the present-day southern Puna margin, but located within the Eastern Cordillera. The K-feldspar 40Ar/39Ar step heating data show a flat age spectrum at ~320 Ma (Figure 3e). Detailed interpretation of the Cerro Durazno results is complicated by a minor intermediate age maximum at ~10% 39Ar released, and the sample’s low apparent activation energy [Lovera et al., 1997]. However, MDD modeling results indicate that temperatures <300°C are not reached until ~300 Ma (i.e., much later than in the eastern Puna), while low temperatures of ~130°C are evident by 180 Ma (i.e., much earlier than for the eastern Puna samples; Figure 4e and Table 2). Two short phases of rapid cooling occur between 320–285 Ma (3.4°C/Myr) and 195–180 Ma (2.4°C/Myr) separated by slower cooling throughout the Permain and Triassic (Figure 4e). The shallow Cretaceous crustal depth (~4–5 km) implied by the Jurassic MDD results is consistent with an old AFT age of 69 ± 2 Ma [Deeken et al., 2006], indicating a temperature below ~115°C since the latest Cretaceous.

4.2.3. The Northern Eastern Puna/Eastern Cordillera Transition

[27] The Cobres granite K-feldspar 40Ar/39Ar data have a monotonically increasing age spectrum (Figure 3f). This sample cooled to ~300°C by ~230 Ma (Figure 4h). Although its thermal history is poorly constrained below ~210°C, slow cooling (0.3°C/Myr) is evident throughout the Mesozoic. Cooling rates start to increase to ~2°C/Myr in the early Tertiary (Figure 4h and Table 2).

[28] K-feldspar 40Ar/39Ar results indicate a late Jurassic crystallization age for the Aguilar pluton. Specifically, the high-temperature release steps yield 40Ar/39Ar ages that closely approach the ~150 Ma crystallization age (Figure 3g). A minor intermediate age maximum between 10 and 20% 39Ar released hampers our ability to calculate a thermal history from these results. Nevertheless, the results indicate a protracted residence above 300°C (>8.5 km depth) prior to 100 Ma (Figure 4g). The Aguilar intrusion is reported to be fault controlled [Cristiani et al., 2005], but low temperatures of ~<120°C (~4–5 km depth) are not attained before 25 Ma. The Aguilar granite records two periods of rapid cooling (~4–5°C/Myr) during the late Cretaceous-Paleocene (78–60 Ma) and the Oligocene (34–25 Ma) with slower cooling (~1°C/Myr) during the intervening Eocene interval (Figure 4g). Although Miocene AFT data from this sample (16.6 ± 1.2 Ma [Deeken et al., 2005]) are not inconsistent with our cooling history, thermal AFT modeling utilizing track length distributions could not simultaneously fit both the K-feldspar MDD and apatite thermal history constraints (A. Deeken, personal communication, 2012). One reason for the discrepancy could be the low number of datable grains and the scarcity of confined tracks in the AFT data (A. Deeken, personal communication, 2012).

[29] In contrast, the Fundiciones granite does not show protracted slow cooling above 300°C during the Cretaceous and instead has a flat age spectrum at ~150 Ma (Figure 3h). Rapid cooling (~7°C/Myr) occurred shortly after initial crystallization during the late Jurassic and early Cretaceous. This sample cooled to temperatures of ~150°C by 130 Ma (Figure 4g and Table 2). Subsequent cooling is characterized by an abrupt decrease in cooling rate to ~0.2–0.4°C/Myr without further rapid cooling. An AFT age of ~13.9 ± 0.9 Ma implies that the sample continued to reside at ~4–5 km depth until the middle Miocene [Deeken et al., 2005].

4.2.4. Eastern Cordillera Tertiary Plutons

[30] The Ar age spectrum for the Nevado de Acay has a well-defined plateau at ~12 Ma (Figure 3i). We attribute the age discrepancy between 206Pb/238U zircon and 40Ar/39Ar K-feldspar to minor 40Ar contamination of the K-feldspar and note that similar problems may have influenced earlier K-Ar studies [Mirre, 1974; Linares, 1979]. Regardless, rapid
cooling after pluton emplacement is consistent with both shallow emplacement and rapid exhumation (Figure 4i).

4.3. Summary

[31] The $^{40}$Ar/$^{39}$Ar K-feldspar analyses indicate diverse Paleozoic to early Cenozoic cooling histories in the Puna/Eastern Cordillera transition zone. The western Puna (Sierra de Macón sample) indicates stable and relatively cool crustal conditions characterized by slow cooling from below ~200°C (<6–8 km depth) since 400 Ma. Samples from the southeastern Puna were hotter overall and experienced slow cooling between 300°C and 200°C between the Devonian and Cretaceous with temperatures of ~200°C (~6–8 km depth) attained by 120 Ma. These samples (Complejo Oire, Valle de Luracatao, Colomé) reflect crustal thermal stability during most of Paleozoic and early to mid-Mesozoic times, followed by increased cooling rates during the early Cretaceous (~160–100 Ma). In contrast, the Cerro Durazno in the southern Eastern Cordillera cooled below 200°C by early Permian time (~270 Ma) and indicates rapid cooling in the early and middle Jurassic (~194–180 Ma).

[32] The northern Puna (Cobres and Aguilar granites samples) experienced slow protracted cooling until the late Cretaceous and faster cooling rates during the late Cretaceous/early Paleocene (78–55 Ma). Temperatures of <200°C (~6–8 km depth) were not attained until the Paleocene (60–55 Ma). These samples also show accelerated cooling during the Eocene. The northern Eastern Cordillera (Fundiciones granite) indicates rapid cooling in the early Cretaceous, reaching temperatures of <200°C by ~140 Ma, almost immediately following pluton emplacement. There is no independent evidence that the Fundiciones and Aguilar plutons initially crystallized at different depths. However, the initial rapid cooling of the Fundiciones from crystallization temperatures to ~150°C within 30 Ma and subsequent residence of the pluton between 150°C and 115°C strongly suggests either very shallow intrusion depths and/or very rapid magma ascent along faults. In contrast and despite its close proximity to rift bounding faults, the Aguilar granite resided at higher temperature throughout the Cretaceous and Paleogene, suggesting somewhat deeper intrusion levels.

5. Discussion

[33] Our new geochronologic data set provides insights into the exhumation history of pre-Cretaceous igneous basement rocks underlying the Puna and Eastern Cordillera in northwestern Argentina. The Eastern Cordillera and Puna regions have previously been characterized as having distinct pre-Paleozoic metamorphic and magmatic histories [e.g., Ramos, 2008]. Our data emphasize that these domains exhibit substantial contrasts in their Paleozoic and Mesozoic cooling histories. Furthermore, differences also exist between rock samples analyzed from within the northeastern and southeastern Puna/Eastern Cordillera transition. In sections 5.1–5.3, we discuss these differences in the context of the regional Paleozoic, Mesozoic, and Cenozoic tectonic history and compare them to existing thermal history information.

5.1. Paleozoic Evolution

[34] Evidence from high-grade rocks of northern Chile and northwestern Argentina implies high-temperature, low-pressure conditions of metamorphism, and a high thermal gradient at midcrustal levels during the early Paleozoic [Lucassen et al., 1996]. Previous K-Ar analyses indicate cooling of metamorphic rocks in the western Puna belt to temperatures of ~500°C at ~480–450 Ma [Lucassen et al., 1996] and ~300°C at 400 Ma [Lucassen et al., 2000; Damm et al., 1990; Maksaev, 1990]. Thus, initial rapid cooling of the Sierra de Macón below 300°C prior to the Devonian, and subsequent cooling to ~150°C by the earliest Jurassic (Figure 4a), points to protracted crustal residence at shallow depths (~8–4 km?) during this interval, assuming a geothermal gradient of 35°C/km. Protracted upper crustal residence is consistent with suggested fast subsidence rates in an early Ordovician back-arc basin in that region [Mon and Salfity, 1995; Coira et al., 1999] and suggests the establishment and persistence of a relatively stable thermal regime following earlier Ordovician tectonism and magmatism [e.g., Bahlburg and Herve, 1997]. Patterns of subsidence and facies distribution are compatible with the emplacement of supracrustal loads of 8 km thickness on the western margin of the Puna basin [Bahlburg and Furlong, 1996]. Inverse thermal modeling of AFT data from the Arizaro basin just west of the Sierra de Macón [Carrapa et al., 2009] indicates shallow crustal levels of ~140°C (~4 km) were achieved by ~200 Ma.

[35] Emplacement of Ordovician plutons in the region of the present-day eastern Puna and the Eastern Cordillera was coeval with regional deformation under high-temperature metamorphism [Becchio et al., 1999; Hongn and Riller, 2007; Wegmann et al., 2008]. Similar ages for all granitoids signifies simultaneous thermal events within the crust across the study area with abundant magmatism and intrusion at middle to lower crustal levels [Lucassen et al., 2002]. This scenario has been either associated with Famatinian continental arc magmatism [e.g., Coira et al., 2009b; Otamendi et al., 2010] or a broad “mobile” belt of thickened crust and magmatic activity [Lucassen et al., 2000]. Crustal temperatures of ~400–300°C by ~416–400 Ma have been reported in the Sierra de Quilmes, ~50 km to the south [Büttnner et al., 2005]. Our Luracatao range samples (Complejo Oire, Valle de Luracatao, Colomé) show protracted cooling throughout the Paleozoic with temperatures <300°C reached by the Devonian. A mechanically stable and shallow crust at this time is supported by a general lack of a post-Ordovician sedimentary record in northwestern Argentina prior to the latest Carboniferous [Bahlburg and Herve, 1997] and little to no metamorphism in the Cochina-Escaya complex sedimentary units [Bahlburg, 1990; Coira et al., 1999].

[36] In contrast, the Cerro Durazno granite in the southern Eastern Cordillera and the Cobres granite in the northern Puna indicate significant differences in their cooling histories, implying regional variations in the thermal and structural regimes. Relative to the other samples, the Cerro Durazno and Cobres granites resided at higher temperatures (>300°C) until the late Paleozoic. These higher temperatures could imply either somewhat deeper crustal levels (>8 km) or unrecognized transient heating. A somewhat deeper crystallization depth at midcrustal levels is consistent with observations indicating that the emplacement of the Cerro Durazno and Cobres plutons was coeval with regional deformation under high-temperature metamorphism [Hongn
and Riller, 2007]. Furthermore, emplacement of these granitoids has been genetically related to prominent shear zones [Hongn et al., 1999; Hongn and Riller, 2007; Wegmann et al., 2008]. The proximity between brittle or ductile deformation zones and magma ascent and emplacement suggests that tectonically assisted transport and emplacement of magma occurred [Hongn and Riller, 2007]. Thus, despite significantly less erosion levels north of the OETL, a deeper crustal residence of the Cerob granite is reasonable, assuming that nearby faults assisted in subsequent exhumation of the granitoid. Deformation and plutonism is known to have occurred in the early Carboniferous, followed by the development of intracratonic basins [Coira et al., 1982]. Rapid cooling of the Cerro Durazno at ~320 Ma could reflect vertical movements in relation to theses basins associated with fault reactivation [Mon and Salfity, 1995]. The cooling differences between the Cerro Durazno and Luracatao ranges point to variable pulses of deformation and/or spatial differences in the thermal structure underneath the eastern Puna and the Eastern Cordillera. We infer the early Jurassic cooling of Cerro Durazno to be related to the onset of local extension at that time.

5.2. Cretaceous Rifting

[37] Relatively stable Ordovician to Jurassic thermal conditions ended with prerift magmatism of the continental Salta rift in the late Jurassic (~160–150 Ma) [Galliski and Viramonte, 1988; Salfity and Marquillas, 1994]. Rifting created regional topographic gradients and a horst-and-graben crustal architecture (Figure 4) with Cretaceous sedimentary units unconformably overlying Paleozeug and Precambrian basement [Marquillas et al., 2005]. Our intrusion ages of ~160 to 150 Ma for the Fundiciones and Aguilar granites correlate with this late Jurassic magmatism. These plutons are located in the Tres Cruces branch of the Salta rift, in close proximity to older Paleozoic structures such as the SW-NE oriented Cobre lineament (Figure 4). [38] The Jurassic-Cretaceous cooling histories differ for samples depending on their location relative to the rift structures. The western Puna (i.e., Sierra de Macón) was not influenced by Paleozoic and Mesozoic rift-related faulting and subsidence. It indicates thermally and mechanically stable crust for the western Puna. In comparison, samples from the southeastern Puna indicate rapid cooling during the prerift phase (~153–125 Ma [Marquillas et al., 2005]) and the first synrift cycle (~125–98 Ma [Marquillas et al., 2005]), likely reflecting initial faulting and tectonic subsidence in the Brealito subbasin (Figure 4). Although MDD models for the Colomé pluton and Cerro Durazno range are only weakly constrained to Cretaceous and younger ages, AFT data from these samples indicate rapid cooling during the early Cretaceous [Deeken et al., 2006]. Interestingly, when combined with the AFT data, rapid cooling indicates a southward temporal trend, beginning at the Complejo Oire granite ~164–120 Ma, then Valle de Luracatao and Colomé ~150–100 Ma, and finally the Cerro Durazno Range 120–80 Ma. The precise age of individual rift structures is unknown, but rift-related fault displacement is the most likely cause for the rapid cooling of these intrusions. If true, these spatiotemporal variations in cooling imply that different faults were active at different times. Rapid cooling of the southern samples ceased with the end of the first synrift stage in the middle Cretaceous, probably associated with the termination of faulting [Bianucci et al., 1982] and a regional modification of rift margins in the Brealito subbasin (Figure 4). [39] In the north, the Cretaceous cooling history at Fundiciones differs substantially from the Cobres and Aguilar cooling histories. The Fundiciones granite has a pronounced alkaline geochemistry [Viramonte et al., 1999], consistent with minor partial melting of a depleted subcontinental mantle [Galliski and Viramonte, 1988]. We interpret the fast initial rapid cooling at Fundiciones as magmatic cooling following shallow emplacement depths and initial extension. In comparison, petrological analyses show that the Aguilar pluton has subalkaline affinities [Galliski and Viramonte, 1988], consistent with the Cretaceous Salta rift intracratonic setting. Both the Aguilar and Cobres granites resided at temperatures >250°C until the late Cretaceous, then cooled rapidly during the latest Cretaceous to middle Paleogene (~70–55 Ma). During this time, the Aguilar and Cobres granites were within the realm of the rift system (Figure 4) and might have been affected by thermal subsidence during the initial postrift stage (~70–60 Ma [Marquillas et al., 2005]). This interpretation is consistent with previous studies that have attributed the deposition of the late Cretaceous/early Paleogene sedimentary Balbuena group to postrift thermal subsidence [e.g., Galliski and Viramonte, 1988; Grier et al., 1991; Salfity and Marquillas, 1994; Marquillas et al., 2005]. However, the observed cooling could also be caused by shortening and hence reflect thrust-related crustal loading in northern Chile that induced flexural subsidence in the distal parts of the basin and overprinted the postrift thermal subsidence during the deposition of the Paleocene/Eocene Santa Barbara group [Deceles et al., 2011; Sik and Horton, 2011]. [40] The differences in cooling histories between the Cerro Durazno and the Luracatao samples along the southeastern Puna/Eastern Cordillera transition, as well as differences between the Fundiciones and the Aguilar and Cobres granites along the northeastern Puna/Eastern Cordillera transition, suggests important differences in the timing of cooling and fault initiation/reactivation along the eastern Puna and the Eastern Cordillera. This contrast could reflect long-standing Paleozeois disparities in the crust or may be related to differences in Paleozeois lithology, sedimentary cover and/or proximity to the magmatic arc/deformation front.

5.3. Cenozoic Cooling

[41] Paleogene crustal shortening caused the compressional inversion of rift-related normal faults [Grier et al., 1991; Cristallini et al., 1997; Kley and Monaldi, 2002; Kley et al., 2005]. Topographic uplift and exhumation of crustal segments at the eastern Puna/Eastern Cordillera boundary is constrained at 50 to 24 Ma by AFT data [Andriessen and Reutter, 1994; Coutand et al., 2001; Carrapa et al., 2005; Deeken et al., 2006]. Our K-feldspar results help to resolve the Cenozoic cooling history for only one sample. The Aguilar granite MDD modeling shows a phase of fast cooling from 35 to 25 Ma, consistent with the AFT data (Figure 4g). For all other samples, our results constrain cooling throughout the Cenozoic to have been less than ~150°C, the typical minimum temperature sensitivity for 40Ar diffusion in K-feldspar. This implies that total
Cenozoic unroofing was less than 6 km, assuming a 25°C/km geotherm.

Miocene volcanic centers are concentrated on NW-trending shear zones that mark long-lasting zones of lithospheric weakness [e.g., Riller and Oncken, 2003]. These zones with inferred strike-slip kinematics may have favored development of shallow magma chambers and eruptive centers [Petrinovic et al., 1999]. Our new age determination of ~12.6 Ma for the Nevado de Acay, located along the Olacapato–El Toro lineament, is

Figure 5. Shaded 30 m ASTER GDEM of the central Andes (~22°S–28°S). White diamonds represent our sample location; black symbols indicate previously published data. The numbers indicate the timing of shortening, deformation and/or enhanced cooling inferred from geothermochemistry, stratigraphy, and structural data. Superscript capital letters refer to references. Parentheses refer to our individual samples, with asterisk indicating interpretation from Deeken et al. [2006].
consistent with a ~15–10 Ma volcanic pulse in this region [Petrinovic et al., 1999; Kay et al., 2010]. The geochemical signature of the Nevado de Acay has been associated with a thin crust [Coira et al., 1993] and well-defined fractionational crystallization [Petrinovic et al., 1999]. This suggests that geothermal gradients may have been higher than normal, and that nearby samples record less denudation of 6–4 km. This is consistent with thin (~0.2–2 km) late Miocene to Plio-Pleistocene stratigraphic sequences in that area [Jordan and Alonso, 1987].

6. Implications

[43] Our results indicate variable cooling histories that suggest differences in (1) the thermal gradient between the western, northeastern and southeastern Puna since the early Paleozoic and (2) the timing of regional activation/reactivation of long-lived structural discontinuities that have influenced the Cenozoic tectonic evolution of the Puna and the neighboring foreland sectors [Hilley et al., 2005; Hongn et al., 2007; Strecker et al., 2007]. Samples of magmatic rocks from the northern Puna indicate residence at higher temperatures and/or greater crustal depth throughout the Paleozoic and Mesozoic in comparison to the southern Puna at present-day erosion levels. This temperature contrast probably persisted until the late Miocene when abundant ignimbrite volcanism in the northern Puna points to a higher degree of melting as to the south [Kay and Coira, 2009].

[44] In addition to detecting regional differences in the Paleozoic and Mesozoic cooling histories, our results reveal an episode of accelerated cooling focused within a narrow zone along the eastern Puna/Eastern Cordillera transition. Combining our data with available thermochronological, structural, and sedimentological data highlights synchronous rapid cooling along the present-day eastern and western Puna margins (Figures 5a and 5b). The onset of extensional tectonics in the Salta rift is recorded by cooling of the crust in the region that now comprises the eastern Puna margin during the late Jurassic and early Cretaceous. AFT analyses on rocks from the eastern Puna margin indicate old age populations of detrital grains, ranging between ~158 and 100 Ma [Coutand et al., 2006; Mortimer et al., 2007; Carrapa and DeCellettes, 2008], whose source region can be related to the outcropping Cretaceous strata (Figure 5a). Along the western Puna margin, K-Ar and AFT data record a rapid cooling phase between ~152 and 118 Ma that has been interpreted to reflect contraction and uplift along the western central Andes (Figure 5a) [Andriessen and Reutter, 1994; Juez-Larré et al., 2010]. Late Cretaceous to Paleogene rapid cooling correlates with thermal relaxation and cessation of rifting. Synchronous cooling recorded in AFT data along the eastern margin of the Puna indicates active tectonic deformation and exhumation during this time [e.g., Coughlin et al., 1998; Coutand et al., 2001; Coutand et al., 2006; Mortimer et al., 2007; Carrapa and DeCellettes, 2008], whereas late Cretaceous cooling along the western Andean margin is likely related to exhumation and eastward migration of the magmatic arc (Figure 5b) [e.g., Kraemer et al., 1999; Mpodozis et al., 2005; Juez-Larré et al., 2010].

[45] Our K-feldspar thermal histories are not well constrained beyond the Paleogene but do place upper bounds on the total amount of Cenozoic denudation that could have occurred in the this region of the plateau at <4–6 km. The AFT analyses of our samples suggest accelerated Eocene exhumation along the eastern Puna margin [Deeken et al., 2006]. In addition, Eocene deformation and exhumation has been proposed for many ranges along the eastern Puna/Eastern Cordillera. This region forms the transition to the broken foreland [e.g., Coutand et al., 2001; Coutand et al., 2006; Carrapa and DeCellettes, 2008; Bosio et al., 2009] and delineates the eastern border of a pronounced negative Bouguer gravity anomaly [Tassara et al., 2006]. Exhumation and surface uplift are also evident along a relatively narrow region along the western Puna margin [Jordan and Alonso, 1987; Reutter et al., 1991; Maksaev and Zentilli, 2000; Arriagada et al., 2006] and along the coast [Juez-Larré et al., 2010]. In contrast, tectonic activity during the middle Tertiary has been reported for only a few sites in the region that now forms the interior of the Puna plateau [Kraemer et al., 1999].

[46] Our data document periods of accelerated cooling corresponding to distinct tectonic periods along a spatially focused region in northwestern Argentina, approximately 250 km wide. Reactivation of inherited ductile and brittle deformation zones and metamorphic fabrics in regions with variable thermal structures and differences in crustal depth juxtaposed along these features is most likely responsible for our observed differences in the chronology and magnitude of exhumation. Structures that accommodated Eocene shortening along the eastern Puna margin correlate with pre-strained zones from the Cretaceous/Paleogene Salta rift system, which in turn are mainly associated with Ordovician crustal anomalies. Our thermochronologic data combined with previous studies emphasize the spatial correlation between Paleozoic orogenesis [e.g., Mon and Hongn, 1991; Lucassen and Franz, 2005; Hongn and Riller, 2007], late Jurassic-Cretaceous rift-related structures [e.g., Hongn and Seggiaro, 2001], and areas of Cenozoic mountain and plateau building. Taken together, these observations show that the Puna plateau and the adjacent eastern foreland region experienced a variable Paleozoic to early Cenozoic thermal and structural history and we suggest that this thermal and structural preconditioning, via inherited crustal anisotropies, may have influenced the modern boundaries of the Puna plateau.

7. Conclusions

[47] The Puna/Eastern Cordillera transition zone in northwest Argentina has experienced a variable Paleozoic to early Cenozoic history of cooling and exhumation. Post-crystallization cooling histories from Ordovician granites are noticeably different for samples along the northeastern compared to the southeastern Puna/Eastern Cordillera margin. Granitic basement from the western Puna eruptive belt indicates initial rapid cooling to 300°C by ~420 Ma. Subsequent slow cooling to ~150°C throughout the late Paleozoic and early Mesozoic (~380–200 Ma) points to protracted crustal residence from 8 to 47 km depths (assuming a geothermal gradient of 35°C/km) during this time, and suggests the establishment and persistence of a relatively stable thermal regime following earlier Ordovician tectonism and magmatism.
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