A thermochronometric view into an ancient landscape: Tectonic setting, development, and inversion of the Paleozoic eastern Paganzo basin, Argentina

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ABSTRACT

In this study, we utilize multiple thermochronometric methods, including apatite and zircon fission track, (U-Th)/He, and zircon U-Pb, to evaluate the cooling history and provenance of sedimentary strata of the late Carboniferous to Late Permian eastern Paganzo basin and adjacent basement rocks (Argentina). The strata in the study area represent a long-lived, composite basin system that is interpreted to have experienced multiple periods of deformation, and to have received sediment from a number of different source terranes. These strata are well exposed in the Sierra de Chepes of west-central Argentina. New thermochronometric data and field observations, together with published data from the surrounding mountains, allow us to reconstruct: (1) the cooling history of the underlying basement rocks and the highlands surrounding the basin, (2) the thermal history of the source areas that provided sediment to the basin, and (3) the timing of structural inversion of the basin. Our data suggest that parts of the Sierra Chepes were rapidly exhumed in Late Devonian–Carboniferous times; these exhuming areas supplied sediment to the adjacent basin. In contrast, the overlying red-bed strata originated from a slowly exhuming region located farther east or north of the basin within the Pampean orogenic belt or the Famatinian belt, respectively. Burial by latest Carboniferous–Early Cretaceous times and accompanying synorogenic strata, and the general modification of upper-crustal rocks over geologic time.

INTRODUCTION

Low-temperature thermochronometric methods like fission-track (FT) and (U-Th)/He dating of apatite and zircon have proven to be powerful tools to investigate rock exhumation in active mountain belts (e.g., Bernet et al., 2004; Blythe et al., 2007; Thiede et al., 2009; Enkelmann et al., 2010). In active tectonic settings, the spatial distribution of thermochronometric ages can be compared with observations of climate, topography, structures, and rock types, and this allows us to recognize correlations between crustal cooling and landscape evolution processes (e.g., Zeitler et al., 2001; Willett et al., 2003; Schuster et al., 2005). The reconstruction of tectonic and surface processes in ancient mountain belts, however, is often much more challenging due to the long-term erosion of topography and related syntectonic strata, and the general modification of upper-crustal rocks over geologic time.

To continue to expand the application of thermochronometric tools to understand the evolution of ancient mountain belts, in this study we present new results from multiple thermochronometric methods (apatite and zircon FT and [U-Th]/He analysis) and zircon U-Pb ages from early Paleozoic bedrock and Upper Carboniferous to Permian detrital samples of the eastern Paganzo basin exposed in the Sierras de Chepes of west-central Argentina. The study area was chosen because it includes an exceptionally well-exposed Carboniferous–Paleo–glacier valley filled with Upper Carboniferous to Permian strata of the eastern Paganzo basin that were deposited on early Paleozoic basement (Sterren and Martinez, 1996; Socha et al., 2006; Figs. 1 and 2). Our new thermochronometric and geochronometric results are discussed in context with published data from surrounding mountain belts of the Sierras de Pampeanas and allow the reconstruction of the Paleozoic through Cenozoic geological history of the eastern Paganzo basin and nearby basement exposures. Our study demonstrates that in some ancient settings, low-temperature thermochronometric approaches can yield meaningful first-order insights into the development of landscapes over long periods of geologic time.

GEOLOGICAL SETTING OF THE STUDY AREA

The Upper Carboniferous to Permian strata of the Paganzo basin crop out over a large part of west-central Argentina (Fig. 1; Fernandez–Seveso and Tankard, 1995). These strata are up to 1500 m thick and represent over 35 m.y. of
deposition in marine and continental environments (e.g., Fernandez-Seveso and Tankard, 1995; Limarino et al., 2006). The strata of the Paganzo basin contain a composite record of several different styles of deformation and tectonic settings, including Paleozoic strike-slip deformation, Mesozoic extension and inversion, and late Cenozoic contractional deformation (Fernandez-Seveso and Tankard, 1995; Ramos, 2010). Sedimentation in the eastern Paganzo basin started in the latest early Carboniferous to earliest late Carboniferous with deposition in marine and continental environments (e.g., Fernandez-Seveso and Tankard, 1995; Limarino et al., 2006). We refer to these glacial and glacial-interglacial stratigraphic units as the red beds of the upper section of the Paganzo Group in the text (Fig. 3).

The Sierra de Chepes consists of several smaller mountain ranges, including the Sierra de Los Llanos, the Sierra de Malanzan, the Sierra de Los Lujan, and the Sierra del Porongo, which are all located in the west-central part of Argentina (Figs. 1 and 2). The Sierra de Chepes is today part of the basement uplifts of the Sierra de Pampeanas (Fig. 4), which are a product of Miocene to Holocene flat-slab subduction of the Nazca plate and Juan-Fernandez Ridge (e.g., Barazangi and Isacks, 1976, 1979; Ramos et al., 2002). The basement rocks of the Sierra de Pampeanas are a product of series of Paleozoic orogenies that developed through the accretion of different terranes into the proto-Andean margin of Gondwana. Three tectonic phases have been identified: the Early Cambrian Pampean (580–510 Ma), the Late Cambrian–Ordovician Famatinian (500–440 Ma), and the Devonian Achalian orogenic cycles (420–350 Ma) (e.g., Toselli and Aceñolaza, 1978; Aceñolaza and Toselli, 1981; Omarini, 1983; Ramos et al., 1986; Baldó et al., 1996; Rapela et al., 1998, 2007; Martino, 1999; Siegesmund et al., 2004, 2010; Steenken et al., 2004; López de Luchi et al., 2007; Drobe et al., 2009, 2011). The Sierra de Chepes is mainly composed of granitoid rocks dated at 497–477 Ma, and small (couple of meters thick) discontinuous exposures of metasedimentary rocks (Fig. 3; Pankhurst et al., 1998; Stuart-Smith et al., 1999).

**METHODS**

**Samples**

During a reconnaissance trip across the Sierras de Pampeanas, we collected six samples from strata of the eastern Paganzo basin and four samples from the underlying basement rocks in the Sierra de Chepes (Table 1; Figs. 2 and 3). Three of the bedrock samples are part of the paleovalley floor beneath the Carboniferous strata of the Paganzo Group following the terminology of Azcuy and Morelli (1970). Overlying these strata as the red beds of the upper section of the Paganzo Group in the study area (Sterren and Martinez, 1996) and (modifed after Limarino et al., 2006). We refer to these glacial and glacial-interglacial stratigraphic units as the red beds of the upper section of the Paganzo Group (Fig. 1; e.g., Limarino et al., 2006; Gulbranson et al., 2010). Glacial and deltaic strata related to this initial depositional stage are preserved in the paleo-glacier valley between the towns of Olta and Malanzan in the Sierra de Chepes (Fig. 1; Fig. DR1; Sterren and Martinez, 1996; Socha et al., 2006). We refer to these glacial and glacial-fluvial strata as part of the lower section of the Paganzo Group following the terminology of Azcuy and Morelli (1970). Overlying these Carboniferous strata, fluvial red-bed sandstone and conglomerate characterize the upper section of the Paganzo Group in the study area (Sterren and Martinez, 1996). The red-bed strata occur at the eastern and western flanks of the

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**TABLE 1. SAMPLE INFORMATION AND LOCATION**

<table>
<thead>
<tr>
<th>Sample</th>
<th>Longitude (°W)</th>
<th>Latitude (°S)</th>
<th>Rock description</th>
<th>Time of deposition/stratigraphic unit</th>
</tr>
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<tr>
<td>Bedrock</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>28TR2</td>
<td>66°26.610</td>
<td>30°46.161</td>
<td>Granite</td>
<td>350–320 Ma; Malanzan Formation1</td>
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<tr>
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<td>66°22.526</td>
<td>30°39.126</td>
<td>Granite</td>
<td>324–318 Ma; Guandacol Formation (SI-1)2</td>
</tr>
<tr>
<td>28TR6</td>
<td>66°31.516</td>
<td>30°30.617</td>
<td>Granite</td>
<td>ca. 320 Ma; uppermost Malanzan Formation1</td>
</tr>
<tr>
<td>28TR7</td>
<td>66°21.730</td>
<td>30°38.826</td>
<td>Sandstone layers within fine sandstone and silt</td>
<td>315–305 Ma; Loma Larga Formation1</td>
</tr>
<tr>
<td>28TR8</td>
<td>66°32.657</td>
<td>30°48.328</td>
<td>Sandstone layers within silt and mudstone</td>
<td>350–320 Ma; Malanzan Formation1</td>
</tr>
<tr>
<td>28TR9</td>
<td>66°17.051</td>
<td>30°39.116</td>
<td>Mafic gneiss</td>
<td>324–318 Ma; Guandacol Formation (SI-1)2</td>
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<td>Detrital</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>29TR4</td>
<td>66°18.679</td>
<td>30°38.094</td>
<td>Sandstone with large dropstones</td>
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<td>Sandstone layer underneath big boulders</td>
<td>ca. 318 Ma; boundary SI-1/SI-22</td>
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<td>Granite</td>
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<td>30°48.328</td>
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<td>30°47.283</td>
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</tr>
</tbody>
</table>

*Note: Stratigraphic units are shown in Figure 1. References: 1—Net and Limarino (1999); 2—Gulbranson et al. (2010).*

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**Figure 1. Late Carboniferous paleogeography of the Paganzo basin and Rio Blanco back-arc basin (modified after Limarino et al., 2006). Darker areas denote paleo-topographic heights; white areas denote basins. Box shows the study area and map of Figure 2.**

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**Figure 2.** Paleoclimatic map of the Paganzo basin (modifed after Limarino et al., 2006). Darker areas denote paleo-climatic regions. White areas denote basins. Box shows the study area and map of Figure 2. **

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**Figure 3.** Paleoclimate map of the Paganzo basin (modified after Limarino et al., 2006). Darker areas denote paleo-climatic regions. White areas denote basins. Box shows the study area and map of Figure 2.
Eastern Paganzo basin evolution

Four sedimentary samples are from the lower section of the Paganzo Group that is composed of Upper Carboniferous glacial (28TR7, 29TR4, and 29TR6) and glacial-fluvial (29TR7) strata within the paleovalley (Figs. 2 and 3). Two samples are from the upper section of the Paganzo Group and are composed of red-bed sandstones (28TR1 and 29TR2). The contact between the strata of the lower section of the Paganzo Group and bedrock, as well as the U-shaped valley form that widens toward the west, is observed in the study region. In the Data Repository, we provide field pictures of the Upper and Lower Paganzo Group strata and observations of the contact between bedrock and strata (Fig. DR1 [see footnote 1]).
and Cuerda (1965) and is equivalent to unit SI-2 of Gulbranson et al. (2010; Fig. 1 herein). Sample 29TR7 is from the Loma Larga Formation (315–305 Ma), defined for the Sierra de Los Llanos by Andreis et al. (1986), which is part of the Tupe Formation (SI-2 unit) that was deposited 318–311 Ma (Gulbranson et al., 2010; Fig. 1 herein). Figs. 2 and 3; Table 1). The two red-bed samples (28TR1 and 29TR2; Fig. 3) are from the Solca Formation, defined for the Sierra de Los Llanos by Andreis et al. (1986), and equivalent to the Patquia Formation defined for the provinces of La Rioja and San Juan by Cuerda (1965) (Fig. 3). Their depositional age is less well constrained but is interpreted to be between the late Carboniferous and the Early Permian (<306 Ma; Gulbranson et al., 2010). The red-bed strata in the eastern study area are assumed to be Permian in age (<299 Ma) based on fossilized trees (Crisafulli and Herbst, 2008; Gulbranson, 2011, personal commun.). All samples were crushed and sieved to grain sizes between 80 and 300 µm. After washing the samples, the apatite and zircon grains were separated using standard magnetic and heavy liquid separation.

(U-Th)/He Analysis

Clear apatite and zircon grains without inclusions and other impurities were selected from the bedrock samples using a binocular microscope. After documenting the grain dimensions, single grains were packed in Nb-tubes for (U-Th)/He analysis. In general, we analyzed 3–6 aliquots per sample. The samples were analyzed in the Patterson helium-extraction line at the University of Tübingen, which is equipped with a diode laser to extract the gas. Apatite grains were heated for 5 min at 11 A, and zircon grains were heated for 10 min at 20 A. Each grain was heated and analyzed a second time to make sure that the grain was degassed entirely in the first step. The re-extracts generally showed <1% of the first signal. After helium analysis, the grain packages were sent to the University of Arizona at Tucson for U, Th, and Sm measurements using an inductively coupled plasma–mass spectrometer (ICP-MS).

The analytical errors of mass spectrometer measurements are generally very low (<2%); however, the alpha correction factor and the reproducibility of the sample age represent a much larger uncertainty on the (U-Th)/He age, and thus are a more honest display of the age precision. The scatter between the single-grain ages of the same sample is influenced by many factors, such as zoning of U and Th concentration, the alpha ejection, grain size, radiation damage, micro-inclusions, or any impurities. We report the single-grain ages with their analytical error and the mean (U-Th)/He age and the standard deviation as the sample error (Table 2). Due to the old (U-Th)/He ages, the range between single-grain ages appears large (tens of millions of years), but the sample age reproducibility (7%–18% for zircon and 5%–24% for apatite) is common for the (U-Th)/He method (e.g., Flowers et al., 2008; House et al., 2005; Stock et al., 2006; Hacker et al., 2011; Spotila and Berger, 2010). However, since the method is mostly applied to rocks that experienced Cretaceous–Cenozoic cooling, the absolute errors are usually small (<5 m.y.) and give the impression of a higher precision.

Apatite and Zircon Fission-Track Analysis

Fission-track analysis on apatite and zircon was carried out in the fission-track laboratory at the University of Tübingen. Apatite grains from both the bedrock and sedimentary samples were embedded in epoxy, grinded, and polished to expose internal surfaces of the apatites. The apatite mounts were etched in 5.5 M HNO₃ for 20 s at 21 °C to reveal the spontaneous tracks in apatite (Carlson et al., 1999). Afterward, the mounts were covered with a uranium-free muscovite external detector and irradiated with thermal neutrons at the research reactor facility at Garching (FRM-II, Germany). After irradiation, the external detectors were etched in 40% HF for 30 min to reveal the induced fission tracks. After etching, the external detector was placed back to its original position for track counting in apatite. Spontaneous and induced fission tracks were counted at a nominal magnification of x1000 using a Zeiss Axioscope microscope, by focusing first at the apatite surface and then up.
the same way as the apatite mounts. Zircon fissions were properly etched. After etching, the microscope to make sure that all zircon populations and etched for various times ranging from 6 to 12 h. Between the etching steps, we checked the appearance of the etched tracks under the microscope to make sure that all zircon populations were properly etched. After etching, the zircon mounts were prepared for irradiation in the same way as the apatite mounts. Zircon fission tracks were counted under a nominal magnification of x1000 (dry objective) using a Zeiss AxiosImager M2M microscope equipped with an Autoscan stage system.

**Zircon U-Pb Dating**

We measured the zircon U-Pb age of individual zircon grains using laser-ablation (LA) ICP-MS at the Museum für Mineralogie und Geologie (GeoPlasma Laboratory, Senckenberg Naturhistorische Sammlungen Dresden). We analyzed 120 grains per sample for the two detrital samples (29TR2 and 29TR4) and 30 grains for the bedrock sample (28TR4).

Analytical techniques and settings of the analysis of U-Pb isotopes of magmatic and inherited zircon by LA-ICP-MS are described as follows, and instruments settings are given in Table DR1 (see footnote 1). Zircons were analyzed for U, Th, and Pb isotopes by LA-ICP-MS techniques using a Thermo-Scientific Element 2 XR sector field ICP-MS coupled to a New Wave UP-193 Excimer laser system. A teardrop-shaped, low-volume laser cell constructed by Ben Jähne (Dresden) and Axel Gerdes (Frankfurt) was used to enable sequential sampling of heterogeneous grains (e.g., growth zones) during time-resolved data acquisition. Each analysis consisted of ~15 s background acquisition followed by 30 s data acquisition, using laser spot sizes of 25 and 35 µm, respectively. A common-Pb correction, based on the interference- and background-corrected 204Pb signal and a model Pb composition from Stacey and Kramers (1975), was carried out if necessary.
The necessity of the correction was judged based on whether the corrected $^{207}\text{Pb}^{206}\text{Pb}$ value was located outside of the internal errors of the measured ratios. Discordant analyses were generally interpreted with care. Raw data were corrected for background signal, common Pb, laser-induced elemental fractionation, instrumental mass discrimination, and time-dependent elemental fractionation of Pb/Th and Pb/U using an Excel® spreadsheet program developed by Axel Gerdes (Institute of Geosciences, Johann Wolfgang Goethe-University Frankfurt, Frankfurt am Main, Germany). Reported uncertainties were propagated by quadratic addition of the external reproducibility obtained from the standard zircon GJ-1 (±0.6% and 0.5%−1% for the $^{206}\text{Pb}^{206}\text{Pb}$ and $^{207}\text{Pb}^{207}\text{Pb}$, respectively) during individual analytical sessions and the within-run precision of each analysis. Concordia diagrams (2σ error ellipses) and concordia ages (95% confidence level) were produced using Isoplot/Ex 2.49 (Ludwig, 2001). The $^{206}\text{Pb}^{206}\text{Pb}$ age was taken for interpretation of all zircons older than 1.0 Ga, and the $^{207}\text{Pb}^{207}\text{Pb}$ ages were used for younger grains. For further details on analytical protocol and data processing, see Frei and Gerdes (2009). The Th/U ratios (Tables DR2 to DR4 [see footnote 1]) were obtained from the LA-ICP-MS measurements of investigated zircon grains. U and Pb content and Th/U ratio were calculated relative to the GJ-1 zircon standard and are accurate to ~10%.

RESULTS

This study presents data from four different thermochronometric systems that have been applied to two different kinds of samples, i.e., granitoid “bedrock” and sedimentary “detrital” samples. Each thermochronometric system yields ages of the sample that refer to the cooling below a certain closure temperature range and the thermal history within the temperature window where fission tracks anneal and helium diffusion occurs. In the simple case of continuous rock cooling through the temperature window where helium diffusion and FT annealing occur, the thermochronometric age represents the time of cooling below the closure temperature (Dodson, 1973). The zircon FT and (U-Th)/He system have the highest closure temperatures of 290–210 °C and 200–160 °C, respectively, depending on the cooling rate, grain size, and accumulated radiation damage (e.g., Yamada et al., 1995; Brandon et al., 1998; Reiners et al., 2002, 2004; Garver et al., 2005). The apatite FT and (U-Th)/He systems have lower closure temperatures of 110–100 °C and 80–50 °C, respectively (e.g., Green et al., 1986; Carlson et al., 1999; Schuster et al., 2006).

Because of the thermal sensitivity, subsequent heating due to burial or an increase in geothermal gradient can erase the cooling age of a sample. This property is particularly useful for sedimentary samples, because the thermochronometric results can be used to evaluate the postdepositional heating of sediment/strata in a basin setting. A reset sedimentary sample means that postdepositional burial was sufficient to heat the sample well above the closure temperature, which causes complete resetting of the original cooling signal due to the diffusive loss of helium or annealing of fission tracks. If heating was not high enough, but within the temperatures of partial helium retention or partial FT annealing, the sample is called partially reset and is characterized by apparent ages that are older and younger than the depositional age. An unset sample is a sedimentary sample that has not been exposed to high temperatures, and the apparent ages, therefore, predate depositional age and provide information about the cooling of the source rock from which the sediment was eroded.

Bedrock Samples

Figure 2 displays the results of the bedrock dating using apatite and zircon (U-Th)/He and apatite FT analysis. Analytical results are given in Tables 2 and 3. The zircon (U-Th)/He ages are oldest, and the mean ages range from 378 ± 39 Ma to 338 ± 43 Ma within the paleo–glacier valley; the sample outside the valley yielded an age of 39 Ma to 338 ± 43 Ma within the paleo–glacier valley. The bedrock sample located to the north and outside the paleo–glacier valley yielded an apatite (U-Th)/He age of 244 ± 13 Ma, which is similar (within error) to the apatite FT age (232 ± 27 Ma) and suggests rapid cooling in the Triassic (Table 2).

The apatite FT and (U-Th)/He ages of the paleo–glacier valley basement postdate the late Carboniferous and thus indicate that they were heated to temperatures of at least within the apatite FT partial annealing zone after their Carboniferous surface exposure (Figs. 5A and 5B). This heating caused partial or full resetting of the apatite systems, which may be one reason that two samples failed the $\chi^2$ test (FT ages) and show a larger scatter between single-grain aliquot measurements (apatite [U-Th]/He ages; Tables 2 and 3). Variations in grain size and radiation damage have been suggested as possible reasons for an observed spread between sample aliquots (e.g., Farley, 2002; Shuster et al., 2006). To investigate a possible correlation, we plotted the apatite and zircon (U-Th)/He ages of individual grains versus the effective uranium content (eU) and the diameter of the grain, but we did not find a correlation (Figs. DR2 and DR3 [see footnote 1]). We measured the diameter of the FT etch pits parallel to the crystallographic c-axis (Dpar) as a kinematic parameter for the apatite chemistry. Apatite chemistry may be responsible for different annealing behavior between grains (Carlson et al., 1999; Donelick and O’Sullivan, 2005). We did not observe variations among the Dpar values of individual grains in the bedrock samples, suggesting that the annealing behavior of the grains is similar.

Confined track length measurements were possible for two apatite samples (28TR2 and 29TR4). Because of the thermal sensitivity, subsequent heating due to burial or an increase in geothermal gradient can erase the cooling age of a sample. This property is particularly useful for sedimentary samples, because the thermochronometric results can be used to evaluate the postdepositional heating of sediment/strata in a basin setting. A reset sedimentary sample means that postdepositional burial was sufficient to heat the sample well above the closure temperature, which causes complete resetting of the original cooling signal due to the diffusive loss of helium or annealing of fission tracks. If heating was not high enough, but within the temperatures of partial helium retention or partial FT annealing, the sample is called partially reset and is characterized by apparent ages that are older and younger than the depositional age. An unset sample is a sedimentary sample that has not been exposed to high temperatures, and the apparent ages, therefore, predate depositional age and provide information about the cooling of the source rock from which the sediment was eroded.

<table>
<thead>
<tr>
<th>Sample</th>
<th>No.</th>
<th>$N_i$</th>
<th>$N_j$</th>
<th>$\rho_i$ (10$^5$ cm$^{-2}$)</th>
<th>$\rho_j$ (10$^5$ cm$^{-2}$)</th>
<th>$\chi_i$ (%)</th>
<th>Age (Ma ± 1σ)</th>
<th>n length (µm)</th>
<th>Mean length (µm)</th>
<th>Dpar (µm)</th>
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<td>28TR2</td>
<td>49</td>
<td>2342</td>
<td>641</td>
<td>1.34</td>
<td>0.383</td>
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<td>130</td>
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<tr>
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<td>41</td>
<td>2271</td>
<td>659</td>
<td>2.68</td>
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<td>119</td>
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<td>100</td>
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<td>0.520</td>
<td>0</td>
<td>$215 ± 20$</td>
<td>77</td>
</tr>
</tbody>
</table>

Note: n—number of measured confined tracks; $P_s$—spontaneous track density; $P_i$—induced track density; $N_i$—number of spontaneous fission tracks; $N_j$—number of induced fission tracks; $P_i$—track density for the dosimeter glass (IRMM540), zeta calibration factor of 241 ± 8 a cm$^{-2}$. The pooled age (bold) is calculated and can be interpreted as the sample cooling age; all other samples are mean ages (normal) calculated for samples that show more than one age population (see Table 4 for component ages).
Eastern Paganzo basin evolution | RESEARCH

TR4) and yielded a unimodal length distribution with mean track lengths of 12.5 ± 1.3 μm and 12.6 ± 1.5 μm (Fig. 5). We present the apparent ages and spread in single-grain ages for the three thermochronometric systems in a temperature versus time diagram together with the field observations in Figure 5. Additionally, we analyzed the zircon U-Pb age of sample 28TR4, which yielded a concordia age of 489.3 ± 1.7 Ma (1σ; Fig. DR5 [see footnote 1]). Sample 28TR6 is located north of the paleo–glacier valley, and the cooling ages differ from those samples collected in the valley. This sample was located at a much deeper crustal level during the early Carboniferous in comparison to those samples in the paleovalley. The mean ages and single-grain ages are presented, revealing the thermal history of the sample, in Figure 5.

**Sedimentary Samples**

For the detrital samples, we analyzed at least 100 grains per sample using the apatite and zircon FT method, and zircon U-Pb dating. The analytical details of single-grain measurements are given in the Data Repository (Tables DR2 to DR6 [see footnote 1]). The mineral separation process for the red-bed samples of the upper section of the Paganzo Group, however, did not yield any apatites for sample 28TR1 and only 20 apatite grains that could be analyzed for sample 29TR2. For the zircon FT analysis, we counted more than 100 grains for all samples, but one (29TR6), which yielded only 70 grains. Distributions of the grain ages from each sample were analyzed using a grain-age-deconvolution and binomial peak-fitting procedure (Galbraith and Green, 1990; Brandon, 1992, 1996; Galbraith and Laslett, 1993). The peak-fitting results are summarized in Table 4 and displayed in Figure 6.
Detailed tables of single-grain ages are provided in the Data Repository (see footnote 1).

With the exception of one sample (28TR1), the detrital zircon FT age distribution is generally older than the late Carboniferous deposition age, with peak ages ranging from 333 ± 40 Ma to 504 ± 70 Ma (Table 4). The detrital apatite FT ages are mostly younger than the deposition ages and are therefore reset (or partially reset), with mean apatite FT ages ranging from 209 ± 7 Ma to 245 ± 8 Ma (Table 3; Table DR5 [see footnote 1]). The apatite age distributions show generally two age populations, with the youngest-age-population peak ranging between 171 ± 10 Ma and 133 ± 13 Ma and older populations with age peaks between 269 ± 25 Ma and 239 ± 21 Ma (Table 4). Sample 28TR7 yielded three age populations that peak at 142 ± 20 Ma, 215 ± 90 Ma, and 322 ± 200 Ma, thus indicating that some grains are not reset. This sample is therefore interpreted as partially reset (Table 4).

The confined track length was measured on each sample, which yielded unimodal length distributions with a mean track length ranging from 11.8 ± 1.6 μm to 13.6 ± 1.3 μm (Table 3; Fig. DR4 [see footnote 1]). The track length distributions are well constrained, with ~300 confined tracks measured per sample. We could not find a difference between the track length distribution of grains that yield younger FT ages and the length distribution of grains with older FT ages.

**DISCUSSION**

**Cambrian to Carboniferous Cooling History**

The granitoid basement rocks of the Sierra de Chepes are part of the Famatinian belt and underlie the eastern Paganzo basin. The basement rocks represent a continental magmatic arc and its metasedimentary host rocks that formed during the Ordovician along the proto-Andean margin of Gondwana (Aceñolaza and Toselli, 1981). Recent petrologic and geochronological studies of the basement rocks in the Sierra de Velasco, located north of our study area (Fig. 4), suggested a two-stage evolution for the exhumation of the Famatinian belt (de los Hoyos et al., 2011). Stage I suggests moderate exhumation rates (0.3–0.8 mm/yr) during the Early to Middle Ordovician driven by erosion of overthickened crust, followed by stage II, which is defined by long-lasting slow exhumation (0.01–0.09 mm/yr) from the Late Ordovician to early Carboniferous. This northern part of the Famatinian belt was additionally affected by post-orogenic shallow magmatic intrusions in the early Carboniferous (monazite U-Pb: 360 Ma; de los Hoyos et al., 2011), but there is no evidence for...
maggmatic activity during the Carboniferous in the Sierra de Chepes and farther south. The Sierra de Chepes basement predominantly consists of I-type and peraluminous granitoid rocks with Late Cambrian to Ordovician crystallization ages (Fig. 4; zircon U-Pb: 477–497 Ma; Pankhurst et al., 1998; Stuart-Smith et al., 1999; this study). Only small patches of greenschist- to amphibolite-grade metasedimentary rocks occur in the Sierra de Chepes with a suggested maximum depositional age of 510 Ma (Drobe et al., 2011), and an Early to Middle Cambrian age of metamorphism (e.g., Pankhurst et al., 1998; Dahlquist et al., 2005).

The bedrock and detrital samples of our study have been analyzed using four different low-temperature thermochronometric systems to reconstruct their thermal history in the upper crust. According to their low closure temperatures, we not only expect the apatite systems to yield the youngest ages, but also that the Paleozoic thermal record may have been easily erased and reset due to subsequent phases of heating and cooling. This is evident by the Mesozoic apatite cooling ages (Tables 2 and 3), which are discussed later herein (sections on “Late Paleozoic Heating and Triassic Cooling History”). The three bedrock samples reveal that the last cooling through ~180 °C occurred in the Late Devonian through early Carboniferous (ca. 378–338 Ma; Table 2). We interpret this cooling to be related to rock exhumation, rather than lowering of the geothermal gradient, based on the fact that basement rocks were exposed at the surface by the end of the early Carboniferous. With surface exposure by ca. 324–318 Ma, our data suggest cooling from ~180 °C to 60–14 °C in 60–14 m.y. Assuming a geothermal gradient of 25 ± 5 °C/km and a surface temperature of 10 ± 5 °C, this suggests a moderate exhumation rate of 0.1–0.6 mm/yr during the early Carboniferous. Support for such moderate exhumation rates also comes from the detrital zircon FT and U-Pb ages of the Carboniferous strata discussed next. This result suggests that different parts of the Famatamine belt experienced different exhumation histories. While the Sierra de Velasco located to the north reveals exhumation with very slow rates during the early Carboniferous (de los Hoyos, 2011), the Sierra de Chepes were exhumed with rates an order of magnitude faster, possibly due to more surface uplift combined with efficient erosion by fluvial or glacial systems. Rapid rock exhumation is generally expected in regions of high uplift rates, creating local relief and elevation, which would also support the development of alpine glaciers that formed the preserved paleo–glacier valleys.

Late Carboniferous to Permian Cooling History and Provenance of Basin Strata

The combination of FT and U-Pb dating on detrital zircons of the late Paleozoic strata allows both identification of the provenance of the sediment supplied to the eastern Paganzo basin and also information about the cooling/exhumation history of the source terranes that provided sediment to this basin. The spread in the single-grain FT ages within the samples is generally wide (Fig. 6; Table DR6 [see footnote 1]), which is typical for detrital zircon FT dating due to the varying concentration of U and Th between individual zircon grains. The damage produced by the spontaneous fission of uranium and the alpha decay of uranium and thorium is accumulated at low temperatures and causes the transition from crystalline to amorphous (metamict) zircon. The resulting variation in the crystalline-to-metamict zircon properties causes a wide range of FT preservation between individual grains within a sample (Garver et al., 2005). That means the sample is a mixture containing low-retentive zircons and high-retentive zircons, which accumulate and anneal fission tracks at low or high temperatures, respectively (e.g., Seward and Rhodes, 1986; Carter, 1990; Garver et al., 2002, 2005). This temperature range can be from <200 °C to >350 °C (e.g., Tagami et al., 1998; Rahn et al., 2004; Garver et al., 2005). Despite the large spread in the zircon FT age distributions, we assume that the ages are not reset, consistent with the zircon (U-Th)/He age of the underlying basement, which are also not reset and reveal an early Carboniferous cooling.

The results of the zircon FT age distribution and binomial peak fitting reveal that all four samples of the lower section of the Paganzo Group yield only one age population that peaked in Late Devonian–early Carboniferous times, whereas the younger red-bed sample (29TR2) of the upper section yielded two populations with peaks in Silurian and Cambrian times (Table 4; Fig. 6). The other red-bed sample (28TR1) yielded zircon ages that are partially reset, with age population peaks that are both older and younger than the depositional age (Table 4; Fig. 6). That means the zircon ages of that sample do not provide cooling information from their source area, and this sample is therefore excluded for our discussion here.

The zircon FT age populations found in the sedimentary samples represent the average cooling signal in the catchment within the source region. There is a clear difference between the glacial and glacial-fluvial strata of the Lower Paganzo Group and the overlying red-bed sample of the Upper Paganzo Group. Each of the four samples from the Lower Paganzo Group...
yields one age population with peak ages that range from 369 to 333 Ma. Taking these age populations and the deposition age, these samples suggest a lag time between 9 and 58 m.y. The lag time is the time the rock needed to cool below the zircon FT closure temperature, to be exhumed to the surface, and then to be eroded, transported, and finally deposited in the basin (e.g., Garver et al., 1999; Bernet and Garver, 2005). Similar to the basement samples, we can use the lag time to estimate the range of exhumation rates for the source area of these late Carboniferous strata. Using the zircon closure temperature of 250 ± 40 °C, an ambient temperature of 10 ± 10 °C, and a geothermal gradient of 25 ± 5 °C/km, the source area for the glacial strata was also exhumed with moderate to rapid rates ranging from 0.1 to 1.6 mm/yr. These exhumation rates overlap with those we estimated for the underlying basement (0.1–0.6 mm/yr) and suggest a proximal source area for the glacial strata. The range of exhumation rates derived from the detrital samples is larger due to the wide range of the zircon FT closure temperatures and the larger uncertainty associated with the wide range of the zircon FT closure temperature and the geothermal gradient as above, to be exhumed to the surface and then to be eroded, transported, and finally deposited in the basin. In contrast, much lower exhumation rates are derived for the red-bed sample (29TR2). The two age populations peak at 432 and 504 Ma, suggesting a lag time of at least 133 m.y. and 205 m.y. for each age peak (assuming a maximum deposition age of 299 Ma; Table 1). Using the same parameters for the closure temperature and the geothermal gradient as above, the sediment source area for this upper section of the Paganzo Group was an area with exhumation rates on an order of magnitude lower (0.03–0.1 mm/yr) than the strata of the lower section of the Paganzo Group and underlying basement rocks. Alternatively, the sediment source area may have experienced higher exhumation rates, but the amount of rock exhumation was not enough (≥6–14 km) to exhume rocks from below the closure temperature depth. In any case, the zircon FT data suggest that the source area for the red-bed strata was different and located farther away than that for the underlying glacial and glacial-fluvial strata. This is also supported by the fact that the red beds are interpreted as having been deposited in braided river environments (Limarino et al., 2006) and yielded two distinct zircon FT age populations, suggesting that the fluvial catchment was probably large and consisted of source areas with different exhumation histories.

To identify possible source regions, we can compare the zircon FT age populations with published biotite K-Ar cooling ages from the surrounding areas (Fig. 4). Biotite K-Ar ages refer to cooling below ~350 °C and are commonly used to study metamorphic processes (McDougall and Harrison, 1999, and references therein). Devonian and Carboniferous biotite cooling ages (418–315 Ma; Steenken et al., 2010) have been reported from gneiss, mylonite, and pegmatite of the Sierra de San Luis, which is located south of our study area and is part of the Famatinian belt (29TR2; Steenken et al., 2008). Carboniferous apatite U-Pb and biotite Rb/Sr ages have also been reported from metagranodiorite and mylonite of the Tinogasta-Pituil-Antinaco shear zone (Höckenreiner et al., 2003). This NNW-trending shear zone separates the Pampean and the Famatinian belt and is exposed today north of the Sierra de Chelces. This suggests that sediment in the glacial and glacial-fluvial strata may have originated from the nearby Famatinian belt (Sierra de Chelces) or the Tinogasta-Pituil-Antinaco shear zone. In contrast, Or dovician and Silurian biotite cooling ages (480–400 Ma) have been reported from gneiss and mylonite of the Sierras de Cordoba located east of the study area, which are part of the Pampean orogenic belt (Fig. 4; Steenken et al., 1998; Schwartz and Gromet, 2004; Schwartz et al., 2008; Siegesmund et al., 2010). In summary, a rapidly exhuming source located near the eastern Pampean basin supplied sediment to the basin during the latest Carboniferous and Permian.

### Late Paleozoic Heating and Triassic Cooling History

As expected, the apatite FT and (U-Th)/He cooling ages are younger than the zircon ages due to their lower closure temperatures. The apatite ages of the basement samples and the overlying strata, however, are also younger than the depositional age of the late Carboniferous to Permian strata (apatite FT: 259–225 Ma and apatite [U-Th]/He: 175–142 Ma; Table 2; Tables 1 and 2). This suggests that the basin and the overlying Paleozoic strata were heated after their surface exposure and deposition, respectively, to temperatures <160 °C, the lower temperature limit for the zircon (U-Th)/He closure temperature. This heating caused a partial and/or total loss of the previously accumulated daughter products in apatite (i.e., diffusion of helium and annealing of fission tracks), but not in zircon.

The unimodal length distributions of the five detrital samples (11.8 μm to 13.6 μm; Table 3; Fig. DR3 [see footnote 1]) generally support the interpretation of full resetting of the apatite FT system; however, the large spread observed in the age distribution suggests that partial resetting of the apatite is also a likely interpretation. A partially reset sample is indicated by a bimodal and/or very broad length distribution with shorter tracks (e.g., Gleadow et al., 1986; Green et al., 1989). The binomial peak fitting...
found generally two age populations with ages postdating deposition. Only one sample 28TR7 (westernmost sample, Fig. 2) yielded a third population that predates deposition and thus is clearly a partially reset sample. The unimodal length distribution of the underlying bedrock suggests maximum heating during Permain time to temperatures in the lower part of the apatite partial annealing zone, probably between 80 °C and 140 °C, followed by cooling (Figs. 5A and 5B).

Previous studies of the Upper Paleozoic strata in the Pagozno basin suggested sedimentation with a general thinning of the strata toward the east. Based on the interpretation of seismic profiles in the La Rioja Basin located to the north (Fig. 4), Fisher et al. (2002) suggested a maximum thickness of 800 m for the Upper Paleozoic strata. Fernandez-Seveso and Tankard (1995) estimated up to 2000 m thickness for the Upper Carboniferous Guadacol Formation and up to 730 m of Permian strata. In total, the estimated thickness of Upper Paleozoic strata appears insufficient to bury the basement to depths with temperatures more than 80 °C for a normal geothermal field. It is possible that estimates of the late Paleozoic strata thickness are too low due to subsequent erosion. We, however, suggest a combination of basement burial by Upper Paleozoic strata and an increase in the geothermal gradient as the probable reasons for the heating of the basement rocks. The southwestern margin of Gondwana was affected by widespread magmatic activity throughout the late Carboniferous to Permian due to major geodynamic reorganization from convergent plate margin to extensional regimes (e.g., Uliana and Japas, 2009; Ramos, 2010). This magmatic event is represent by the Choyoi igneous province, which covers an area of ~500,000 km² of western Argentina and eastern Chile, with a thickness of up to 2000 m (e.g., López-Gamundi, 2006; Strazzere et al., 2006; Kleiman and Japas, 2009; Rocha-Campos et al., 2011). This large magmatic event must have, undoubtedly, influenced the geothermal structure of the Sierra de Chepes region.

Based on this reasoning, we suggest that the detrital apatite FT ages are reset (except sample 28TR7). The older of the two apatite FT age populations found in the reset samples ranges from 269 ± 25 Ma to 239 ± 21 Ma and is similar to the underlying bedrock apatite FT ages, which range from 259 ± 34 Ma to 225 ± 13 Ma. All these ages indicate that cooling occurred in the latest Permian through Triassic after the late Paleozoic burial/heating event (Tables 2 and 3). The Triassic cooling is supported by the bedrock sample north of the paleo–glacier valley (28TR6; Fig. 2), which yielded similar apatite FT and (U-Th)/He ages of 232 ± 27 Ma and 244 ± 13 Ma, respectively, suggesting a phase of rapid cooling in the Triassic (Tables 2 and 3). The zircon (U-Th)/He age of the same sample is 289 ± 51 Ma (Table 2). The entire temperature-time path of this sample suggests rapid cooling in the Permian through Triassic from temperatures >180 °C to below ~45 °C (Fig. 5C).

The Triassic Ischigualasto Basin (Fig. 4) could be a possible sink for the eroded “missing” strata that may have buried the currently exposed Upper Carboniferous strata. This rift basin formed during regional extension at the beginning of Gondwana breakup and contains up to 6 km of clastic strata (Milana and Alcobber, 1994; Stipanicic, 2002). The Ischigualasto Basin formed just west of the Sierras de Chepes and is characterized by a thick section of clastic strata and basalt flows (Gonzalez and Toselli, 1975; Valencio et al., 1975; Milana and Alcobber, 1994; Stipanicic, 2002; Zerfas et al., 2004). The Sierras de Chepes itself is considered to have been a highland at this time, which is supported by our temperature-time paths of the basement samples outside and inside the paleo–glacier valley (Fig. 5). Evidence for late Carboniferous–Permian heating also comes from a granitoid sample of the Sierra de Cordoba, located east of our study (Fig. 4). The K-Ar analysis of muscovite and potassium feldspar suggests cooling well below 120 °C at 320 Ma, followed by a slight heating event that did not exceed 120 °C at 265 Ma (Jordan et al., 1989; Richardson et al., 2013).
The zircon cooling ages provide estimates of Paleozoic exhumation rates for both the basement rocks of the Sierra de Chepes and the sediment source areas of the Carboniferous glacial and glacial-fluvial strata in the eastern Paganzo basin. For the Sierra de Chepes and the nearby surrounding terranes that provided sediment to the late Carboniferous basin, we derived long-term (>10^4 yr) exhumation rates of 0.1–1.6 mm/yr for Late Devonian–Carboniferous times. These exhumation rates, derived from zircon thermochronometric systems, are comparable to those reported from tectonically active mountain ranges like the European Alps (0.2–0.3 mm/yr; Bernet et al., 2009), the Himalayas (0.3–1.4 mm/yr; e.g., Bernet et al., 2006; Blythe et al., 2007; Thiede et al., 2009; Enkelmann et al., 2011), and the Andes (0.2–1.3 mm/yr; e.g., Garver et al., 2005). In contrast, exhumation rates are an order of magnitude lower in ancient, currently not tectonically active mountain belts, such as the Appalachians (0.07–0.08 mm/yr; Roden-Tice and Tice, 2005), Scandinavia (<0.02 mm/yr; e.g., Hendriks et al., 2007, and references therein), or Northern (0.02–0.05 mm/yr; Lisker et al., 2006) and Southern Victoria Land in Antarctica (0.08–0.1 mm/yr; Fitzgerald et al., 2006). All of these active and ancient orogens constitute mountain belts with >1500 m peak elevations, and all of them are glaciated or have been glaciated throughout the Quaternary. However, the average long-term exhumation rates differ by an order of magnitude between the tectonically active and ancient “inactive” mountain belts. Carving deep glacial valleys into a tectonically inactive mountain belt often produces impressive topographic relief and may also result in young cooling ages of the lower-temperature systems such as apatite (U-Th)/He, but it does not result in the rapid exhumation of rocks originating from greater depths below the closure temperatures of zircon FT or the (U-Th)/He system. However, this is the case when efficient fluvial or glacial erosion removes the actively uplifting surface in tectonically active mountain belts. In these types of settings, active mountainous regions show higher relief with peak elevations of >4000 m. We thus could reasonably speculate that this region of the Sierra de Chepes and parts of the orogenic Famatinian belt must have been significant mountain ranges that were tectonically active in the Carboniferous. Based on the results of our study, we suggest that at least parts of the orogenic belts in west-central Argentina were characterized by surface uplift and rapid rock exhumation with amounts that were sufficient to expose rocks from depths below the zircon FT closure temperatures. In contrast, the area located farther east of the Paganzo basin (Pampean orogenic belt), which delivered sediment to the red-bed strata of the upper section of the Paganzo Group, was tectonically inactive in the late Carboniferous through Permian and probably was characterized by much lower topography relative to the Famatinian belt. Our findings show that we can reconstruct the first-order late Carboniferous topography of the eastern Paganzo basin and surrounding highlands; our study provides an example of the type of information that can be acquired with thermochronometric data from older landscapes.

The interpreted deposition of at least 2.5 km of strata on top of the currently exposed Upper Carboniferous paleovalley strata also indicates that during the Permian, the area of the Sierra de Chepes was subsiding and part of a sedimentary basin. Inversion of the eastern Paganzo basin, marked by rapid cooling and exhumation, occurred in Triassic–Early Jurassic time with the beginning of Mesozoic rift development associated with opening of the Atlantic Ocean (Uliana and Biddle, 1988; Uliana et al., 1989; Zerfass et al., 2004). Since that time, the


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