Late Miocene upper-crustal deformation within the interior of the southern Puna Plateau, central Andes

Renjie Zhou1,2,* and Lindsay M. Schoenbohm1,2
1DEPARTMENT OF EARTH SCIENCES, UNIVERSITY OF TORONTO, 22 RUSSELL STREET, TORONTO, M5S 3B1, ONTARIO, CANADA
2DEPARTMENT OF CHEMICAL AND PHYSICAL SCIENCES, UNIVERSITY OF TORONTO MISSISSAUGA, 3359 MISSISSAUGA ROAD NORTH, MISSISSAUGA, L5L 1C6, ONTARIO, CANADA

ABSTRACT

The origin and evolution of the central Andes, a noncollisional orogenic system, have been hypothesized to evolve as a result of several dynamic processes, including formation of an eastward-propagating orogenic wedge, segmentation into rhomb-shaped basins as a result of N-S gradients in crustal shortening, reactivation of inherited deep structures, and lithospheric foundering. How these proposed processes dominate the orogen spatially and temporally is uncertain; however, constraining the timing of upper-crustal deformation is critical for investigating these models. We document the formation and deformation of the Pasto Ventura basin (NW Argentina) in the southern Puna Plateau. Through field mapping, deformation analysis, secondary ion mass spectrometry U-Pb dating of zircon from interbedded volcanic ashes, and 40Ar/39Ar geochronology of volcanics, we show that major basin formation started ca. 11.7–10.5 Ma and continued until at least ca. 7.8 Ma. The basin underwent syndepositional faulting and folding from ca. 10 to 8 Ma. Contractional deformation in the Pasto Ventura basin ended between ca. 7.3 and 4 Ma, based on the onset of regional horizontal extension. Data from the Pasto Ventura region allow us to bridge existing data and complete a regional compilation of upper-crustal deformation for the Puna Plateau. Our analysis shows that late Miocene formation and deformation of the Pasto Ventura basin represent an important out-of-sequence contractional event in the southern Puna Plateau. While a number of geodynamic processes likely shape the evolution of the southern Puna, multidisciplinary data sets, including deformation in the Pasto Ventura basin studied here, highlight the role of the formation and detachment of a late Miocene lithospheric dip in causing the upper-crustal deformation on the southern Puna Plateau since the mid–late Miocene.

INTRODUCTION

Studies of the central Andes have enriched our understanding of orogenic plateaus, such as the Altiplano-Puna Plateau, along noncollisional plate boundaries. In the absence of plate collision, other processes must account for generating the high strain, thick crust, and topography currently observed in the interior part of the overriding South American plate. Understanding the evolution of parts of this broad orogenic system requires understanding the overall physical and chemical parameters controlling tectonic coupling between the subducting plates (e.g., Can- tiano et al., 2011; Lamb and Davis, 2003; Sobolev and Babeyko, 2005). Horizontal forcing from subduction of the Nazca plate and formation of the associated orogenic wedge would result in a deformation front in the central Andes that has migrated in an in-sequence manner from W to E through the interior of the plateau to the modern foreland (e.g., Carrapa et al., 2011; DeCelles et al., 2011; Horton, 2005; McQuarrie, 2002a). Yet, horizontal forcing may preferentially activate preexisting crustal weak zones inherited from the Cretaceous rifting event, causing distributed, out-of-sequence crustal shortening (e.g., Hongn et al., 2007; Sobel et al., 2003; Sobel and Strecke, 2003). Neogene compressional basins, bounded by orogen-parallel and NE-striking fault zones, form rhomb-shaped crustal segments in map view (Fig. 1B; Riller et al., 2012; Riller and Oncken, 2003). Together with an overall N-S gradient in crustal shortening, these structures have accommodated deformation propagating north and south from the center of the Bolivian orocline since the Eocene (Riller et al., 2012; Riller and Oncken, 2003). In addition, significant reduction in the mass of the subcontinental lithospheric mantle beneath this region, either through partial melting or lithospheric foundering, appears to control rates of growth and the width of the central Andes during modulated periods of active thrusting in the orogenic foreland (Dahlen, 1990; DeCelles et al., 2009; Garzione et al., 2006; Kay et al., 1994; Kay and Coira, 2009; Molnar and Garzione, 2007; Schoenbohm and Carrapa, 2015). Locally, the formation and detachment of foundering lithospheric mantle would stretch the upper crust. As predicted by numerical and analog models, crustal shortening takes place in the region above a lithospheric foundering event during the formation of the foundered block prior to its detachment; crustal extension then takes over immediately after the lower lithospheric block detaches and the asthenospheric materials upwell (Göğüş and Pysklywec, 2008; Pysklywec and Cruden, 2004).

While a better understanding of geodynamic mechanisms operating in the central Andes requires a multidisciplinary approach, one key to evaluating these models is a comprehensive deformation history of the plateau and its margins. In the southern Puna Plateau, existing data are extremely sparse. Questions remain over the timing and causes of deformation and the applicability of orogen-wide geodynamic models for the southern Puna Plateau (e.g., Allmendinger et al., 1997; Carrapa and DeCelles, 2008; Carrera et al., 2006; DeCelles et al., 2011; Hain et al., 2011; Hongn et al., 2007; McQuarrie et al., 2005; Kay et al., 1994; Sobel et al., 2003; Strecke et al., 2007). In this investigation, we present mapping and deformation analysis for the Pasto Ventura basin, the only exposed basin on the southernmost Puna Plateau (NW Argentina). High-precision secondary ion mass spectrometry (SIMS) zircon U-Pb geochronology of interbedded ash beds and 40Ar/39Ar geochronology of newly identified, deformed basaltic trachyandesite flows provide age constraints on sedimentation
and contractional deformation. We then discuss deformation of the southern Puna Plateau and adjacent regions, supplemented with other geochronological, geochemical, and geophysical studies, and we explore possible geodynamic processes in the southern Puna Plateau.

BACKGROUND

The Andes is the major topographic feature in South America, extending from tropical Venezuela to glaciated southern Chile. In the central Andes (~15°S to ~27°S), the Central Andean Plateau (the Altiplano-Puna Plateau) is characterized by a high-elevation (>3 km), low-relief interior plateau, surrounded by the higher, more rugged Western and Eastern Cordilleras, the Sierra Pampeanas, and the Santa Barbara Ranges (Fig. 1; Allmendinger et al., 1997; Isacks, 1988; Strecker et al., 2007). The plateau region currently experiences semiarid to arid climate conditions, is internally drained, and is divided at ~22.5°S latitude into the Altiplano Plateau to the north and the Puna Plateau to the south (Fig. 1). The Nazca plate has subducted nearly orthogonally beneath the South America plate boundary since the Cretaceous, with a present speed of ~8 cm/yr (Marrett and Strecker, 2000). Geophysical investigations indicate that the crustal thickness in the Puna is spatially variable, but it ranges from ~50 to 68 km (McGlashan et al., 2008; Whitman et al., 1992, 1996; Yuan et al., 2002).

Cenozoic arc volcanic rocks are widespread along the Western Cordillera and the western Altiplano-Puna Plateau (Fig. 1). The post–late Miocene Altiplano-Puna volcanic complex between latitudes of 21°S and 24°S, covering ~50,000 km², represents an intense episode of felsic volcanism of the central Andes (de Silva, 1989). Ignimbritic complexes are also present across the Puna Plateau, often as collapsed calderas, and they are offset from the modern arc (e.g., Kay et al., 2010; Guzmán and Petrinovic, 2010; Guzmán et al., 2011). Across the Puna Plateau, there are well-preserved monogenetic volcanic cinder cones and associated lava flows; these are dominantly mafic, late Miocene to Holocene, confined to the plateau region, and compositionally distinct from arc volcanics (Figs. 2; Kay et al., 1994; Risse et al., 2008). Pre–late Miocene mafic to intermediate volcanic lavas are also present on the southern Puna Plateau, but their extents and ages are poorly constrained, especially for the southern Puna (Fig. 2A; Roy et al., 2006; Schnurr et al., 2006).

The deformation history of the Puna Plateau, the adjacent Eastern Cordillera, and the Sierra Pampeanas has been studied for several decades, where shortening is primarily accommodated by steeply dipping, bivertical thrust faults that cut deeply into the crust (>25 km; Allmendinger et al., 1997; Cristallini et al., 1997; Cristallini and Ramos, 2000; Jordan and Allmendinger, 1986; Kley and Monaldi, 1998; Kley et al., 1999; Monaldi et al., 2008; Pearson et al., 2013). This thick-skinned tectonic style may be genetically linked to the Salta Rift, which evolved through Cretaceous-Paleogene time (ca. 160–60 Ma) in NW Argentina and cuts across the current Santa Barbara system, parts of Eastern Cordillera, and the Puna Plateau (Galliski and Viramonte, 1988; Marquillas et al., 2005; Salfity and Marquillas, 1994).

Despite several studies from the Eastern Cordillera, data on the timing of Cenozoic upper-crustal deformation such as basin deformation, sedimentation, and bedrock exhumation in the Puna Plateau are sparse, particularly in the southern plateau, mainly due to the extensive volcanic rocks and colluvial cover (Fig. 3). This paucity of data has allowed a number of different, sometimes opposing models for the causes of deformation in the Puna Plateau. Some models emphasize the W-E propagation
of both deformation and basin sedimentation, arguing for a propagating, thick-skinned wedge (e.g., Carrapa et al., 2011; DeCelles et al., 2011). Others argue that upper-crustal deformation is driven by a northward increase in shortening in the plateau, resulting in a southward younging of deformation and basin sedimentation (Fig. 1B; e.g., Riller and Oncken, 2003). Other models draw attention to out-of-sequence deformation observed across the plateau and its adjacent Eastern Cordillera, attributing this to irregular reactivation of faults in a broken foreland (e.g., Hain et al., 2011), change of the mass of a critical taper (e.g., DeCelles et al., 2009), or to lithospheric foundering (e.g., Kay et al., 1994). We summarize existing shortening data in the following paragraphs, separating deformation inferred from rapid cooling deduced from bedrock thermochronology (exhumation in Fig. 3) from deformation inferred from dating of sedimentary strata that predate, overlap, or postdate deformation (pre-, syn-, and post-deformation in Fig. 3). We note that several pre-Cenozoic bedrock cooling episodes have been documented in the Puna Plateau and its adjacent Eastern Cordillera (e.g., Carrapa et al., 2014; Deeken et al., 2006; Löbens et al., 2013; Sobel and Strecker, 2003) and could be related to exhumation associated with extension along the Cretaceous Salta Rift (Carrapa et al., 2014; Sobel and Strecker, 2003). However, the pre-Cenozoic history of the central Andes is beyond the scope of this paper and is not included in the following summary.

In the northwestern Puna Plateau, modeled apatite fission-track data reveal that deformation to the west of the Salar de Arizaro (AR in Fig. 3) took place ca. 42–33 Ma, followed by exhumation around 20 Ma (Schoenbohm and Carrapa, 2015). To the east of the Salar de Arizaro, isotopic ages from deformed strata constrain deformation to after ca. 14 Ma (Alonso et al., 1991; Boyd, 2010; Marrett et al., 1994), consistent with proposed 15–8 Ma exhumation based on thermochronology data (Fig. 3C, section A–A'; Carrapa et al., 2009). To the east of the northern Puna Plateau in the
Late Miocene southern Puna Plateau deformation

Figure 3. Compilation of documented deformation across the southern Altiplano Plateau and the Puna Plateau. (A) Topographic map (Shuttle Radar Topography Mission [SRTM] 90 m digital elevation model) of the southern Altiplano and the Puna, central Andes. The 3 km contour is indicated in black, and the internal drainage area is outlined with blue lines (derived from U.S. Geological Survey HydroSHEDS data). It is noted that some studies directly documented the timing of deformation based on structural relationships (structural data), while others made use of low-temperature thermochronology to date exhumation (exhumation data). The timing of exhumation may relate to structural deformation, but it could also reflect other possible causes such as climate-driven erosion (e.g., Barnes et al., 2012; Sobel and Strecker, 2003). Deformation from locations north of ~24°S is shown in circles, with references in square brackets (sources: 1—Hammerschmidt et al., 1992; 2—Reutter et al., 1996; 3—Mpodozis et al., 2005; 4—Arriagada et al., 2006; 5—Siks and Horton, 2011; 6—Cladouhos et al., 1994; 7—Echavarría et al., 2003; 8—Horton, 1998; 9—Ege et al., 2007). (B) Detailed map of documented deformation in the Puna Plateau and its adjacent Eastern Cordillera (sources: a—Alonso et al., 1991; b—Marrett et al., 1994; c—Boyd, 2010; d—del Papa et al., 2013, Hongn et al., 2007; e—Cristallini et al., 1997; f—Carrapa et al., 2011; g—Mortimer et al., 2007; h—Carrapa et al., 2009; i—Carrapa et al., 2005; j—Sobel and Strecker, 2003; k—Coutand et al., 2001; l—Carrapa et al., 2006; m—Schoenbohm and Carrapa, 2015; n—Deeken et al., 2008; o—Carrapa et al., 2005; p—Carrapa et al., 2013; q—Vezzoli et al., 2012; r—Pearson et al., 2013; s—Carrapa et al., 2014). (C) Plots of the timing of deformation along topographic gradients. The elevation swath profiles are extracted from 50-km-wide swath boxes (shown on Fig. 3B), based on 90-m-resolution SRTM data. On the plots, the constrained timing of pre-/syn-/postdeformation is marked by short, horizontal bars. On the topographic swath profiles, the plateau is the higher-elevation region, bounded by the more steeply sloping eastern/southeastern flanks of the plateau.
Eastern Cordillera, apatite fission-track thermochronology indicates range-scale exhumation and adjacent intramontane basin development migrating eastward into the Eastern Cordillera (Cumbres de Luracatao Range) by 21 Ma (Fig. 3C, section A–A’; Deeken et al., 2006). At the southern end of the Eastern Cordillera, syndevelopmental deformation occurred at ca. 40 Ma (Fig. 3C, section A–A’; del Papa et al., 2013; Hongn et al., 2007).

In the southern Puna Plateau, especially south of 26°S, there are few data constraining deformation, partly because of limited access and partly because of extensive surficial deposits from the Cerro Galán ignimbrite complex (GL, Fig. 3B), which formed from 6.6 to 2 Ma (Fig. 3; Kay et al., 2010), obscuring observations of the underlying geology in much of the region. The 29–24 Ma exhumation of the Calalaste Range west of the Salar de Antofalla (ANT, Fig. 3) is the only other region containing documented Cenozoic shortening on the southern Puna Plateau (Carrapa et al., 2005). Deformation of the eastern and southern flanks of the southern Puna Plateau and in the adjacent Sierras Pampeanas was diachronous. Apatite fission-track data from an E-W transect (Fig. 3C, section B–B’) across the Chango Real pluton (CR, Fig. 3) indicate that the southeastern flank of the Puna Plateau was exhumed ca. 38–29 Ma (Figs. 3B and 3C; Coutand et al., 2001). Farther east, U-Pb geochronology of intercalated ashes and detrital fission-track data from strata within the El Cajón basin (Caj, Fig. 3) indicate uplift of the eastern plateau margin (west margin of basin) from 13.6 to 10.7 Ma, and of the Sierra de Quilmes to the east of the basin from ca. 10 to 6 Ma (Mortimer et al., 2007; Schoenbohm et al., 2007). Exhumation of the next range to the east, the Sierra del Aconquija (Acon, Fig. 3), occurred around 5.5 Ma (Fig. 3C, section B–B’; Sobel and Strecker, 2003). On the southern flank of the Puna Plateau (Fig. 3C, section C–C’), apatite fission-track and (U-Th)/He thermochronology indicate rapid exhumation at ca. 21–14 Ma in the Cerro Negro (Carrapa et al., 2014). This exhumation is also recorded by detrital apatite fission-track data in sediments preserved in the Fiambalá basin (Fia, Fig. 3) with a strong 14 Ma signal, suggesting the southern margin of the Puna Plateau was exhuming during the middle Miocene (Carrapa et al., 2006), and this is in line with the estimate of pre–9 Ma local relief establishment (Monteiro-López et al., 2014).

METHODS AND RESULTS

Geological Mapping in the Pasto Ventura Basin, NW Argentina

The landscape of the southern Puna Plateau is dominantly covered by recent lava flows, ignimbrites, and colluvium. The Pasto Ventura region (~26°48’S, ~67°16’W) is located on the southern Puna Plateau (Fig. 2) and outcrops in a narrow, elongated, roughly N-S-striking basin that contains sedimentary rocks of primarily Neogene age (Schoenbohm and Carrapa, 2015; Zhou et al., 2013; Fig. 4). The Pasto Ventura basin is the only exposed, relatively continuous sedimentary record in the southernmost Puna Plateau (Fig. 2A). We mapped this basin using an aerial photograph base (Instituto Geografico Militar, Argentina) through both field and remote mapping, supplemented with satellite images and digital elevation models (DEMs), including Landsat, Google Earth, ASTER (Advanced Space-borne Thermal Emission and Reflection Radiometer), and SRTM (Shuttle Radar Topography Mission). We present two maps of the Pasto Ventura basin, covering the majority of its extent: the north Pasto Ventura map (the NPV map; Fig. 5) and the south Pasto Ventura map (the SPV map; Fig. 6). Our maps update previous work in parts of the Pasto Ventura basin (Allmendinger et al., 1989; Schoenbohm and Carrapa, 2015; Zhou et al., 2013) and, along with new structural, geochronological, and geochemical data, allow us to explore the deformation history and dynamics of this region.

Most of the basin-fill units are exposed along a river-cut valley running approximately N-S through the Pasto Ventura basin, with the rest of the region largely covered by Quaternary sediments and Quaternary mafic volcanic rocks. Quaternary units overlying the deformed Neogene strata include modern eolian sand dunes (Qe), Quaternary alluvial and fluvial sediments (Qaf), and Quaternary colluvial and terrace sediments (Qct). Unit Qe is usually found along the lee side of river-cut channels or hills, where locally high topography favors the accumulation of large volumes of sand carried by prevailing northwesterly winds. Unit Qaf is sand-/gravel-grade sediments found within river channels (fluvial origins) or fans (alluvial origins). Unit Qct comprises low-slope surfaces covered by various unconsolidated, thick (up to several meters) sediments. The clasts composing unit Qaf and unit Qct are mainly derived from metamorphic bedrock (the Puncoviscana Formation) and mafic and intermediate lava
Figure 5. Geological map of the northern Pasto Ventura basin (the NPV map). Ages with (*) are from Schoenbohm and Carrapa (2015). The map base is aerial photography. (Upper right) $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology result for sample PV11B-01, which was sampled from the deformed basaltic trachyandesite unit (unit Nfl).
Figure 6. (A) Geological map of the southern Pasto Ventura basin (the SPV map). The map base is aerial photography. (B) Cross sections for the geological map of the southern Pasto Ventura basin (the SPV map). Locations are shown in Figure 2A.
flows. The clasts are generally ~5–10 cm in size and are characterized on the surface by wind-blown erosion surfaces.

The Quaternary mafic volcanic units include Qbc (cinder cones), Qbf (lava flows), and Qwb (wind-blown clasts). Unit Qwb is characterized by dark colors on images and composed of irregular-shaped clasts derived directly from nearby lava flows (Qbf) or cinder cones (Qbc). The recent volcanism is characterized by monogenetic volcanoes with associated lava flows and is spatially and temporally linked with post-Pliocene extension (e.g., Allmendinger et al., 1997).

We divide the Neogene sedimentary rocks into four units (units Ns-1, Ns-2, Ns-3, and Ns-4; Figs. 5 and 6). The thickness of these units varies across the region, ranging from 100 to 200 m to up to 1 km, and it may reflect varying depositional conditions throughout this relatively small basin. Unit Ns-4 is inferred to be the oldest among the four units because it does not contain interbedded volcanic ashes, whereas this region has been experiencing felsic volcanism since ca. 20–16 Ma (Kay et al., 2010). Unit Ns-4 outcrops along and beyond the northwest edge of the NPV map (Fig. 5), where it lies in depositional contact with the bedrock basement (unit M). It consists of medium to coarse cross-beded red sandstone and siltstone with occasional cobble layers (~5–10 cm in thickness) and pedogenic carbonate nodules (Fig. 4). In the central part of the basin, unit Ns-3 is a mudstone with thin (up to 20 cm) interbedded fine sandstone to siltstone layers (Fig. 4B). This unit, as well as units Ns-2 and Ns-1, is interbedded with volcanic ashes or pumice layers, with thicknesses varying from several centimeters to ~1 m (Fig. 4B). Locally, cross-beding is observed within fine sandstone to siltstone layers. The mudstone alternates between green and red in color on a scale of several tens of centimeters to ~1 m. Within the red mudstone, gyspsum deposits are found as thin (~1–2 cm) layers. Conformable with unit Ns-3, unit Ns-2 is dominantly composed of fine- to medium-grained sandstones, with interbedded conglomerate layers (Fig. 4C). These conglomerate layers are typically ~0.5 m thick with clasts of dominantly phyllite, basalt, and granite, ranging in size from 1 to 5 cm. The clasts are well rounded and relatively uniform in size within each bed. Conglomerate beds make up a larger portion of the unit in the southern part of the basin. Within the sandstones, cross-beding is widely observed. Soft-sediment deformation such as convolute bedding is also observed and is only found locally within isolated beds. Conformable with unit Ns-2, unit Ns-1 is a medium-grained eolian sandstone featuring low-angle (usually ~30°) and large-scale (up to several meters) eolian cross-beds (Fig. 4A). In the southeast part of the NPV map, outcrops of unit Ns-1 found along a river-cut valley appear to be hydrothermally altered and are brown to black in color, but the strong eolian cross-beds are preserved (Fig. 4B).

Unit Nfl constitutes volcanic flows that crop out in the center of the NPV map (~67°14′S, 26°47′W; Fig. 5). Nfl flows contain phenocrysts that are mainly plagioclase (up to ~2 cm) with sparse olivine; the groundmass contains plagioclase and olivine that have been mostly altered to iddingsite, and calcite-filled vesicles are also present (Fig. DR3). Unit Nfl is overlain by unit Ns-3 red mudstone and therefore predates deposition of units Ns-3, Ns-2, and Ns-1. However, the relationship between Nfl and Ns-4 is unconstrained by mapping. We performed geochemical and 40Ar/39Ar geochronological analysis on samples from unit Nfl in order to get a better understanding of its origins and to put better constraints on basin sedimentation and deformation. Details are included in the following sections.

**U-Pb Geochronology of Volcanic Ashes**

We sampled volcanic ash layers from deformed sedimentary units in order to place age constraints on deposition and deformation (Figs. 5 and 6; Table 1). Zircons were separated following standard methods at the Jack Satterly Geochronology Laboratory at University of Toronto (e.g., Barker et al., 2011). U-Pb measurements were performed at the SIMS Center at University of California Los Angeles (UCLA) following methods described by Schmitt et al. (2003).

We dated 16 ashes, six of which are from unit Ns-3 and 10 of which are from unit Ns-2 (Table 1). For each ash, ~10–20 grains were dated. The age of an ash was calculated from individual grain ages in a clustered age population, excluding very young (Quaternary age) and old (up to 100s or 1000s Ma) ages that likely reflect contamination by Quaternary volcanic activities in the region and detrital input from the plateau bedrock, respectively. We combine our new data with published data (Schoenbohm and Carrapa, 2015) to present a complete data set for all of the dated volcanic ashes for the Pasto Ventura basin (Table 1). Due to uncertainties in the dating technique, some ages overlap or are identical with each other. We group them into Ns-1, Ns-2 upper, Ns-2 lower, Ns-3 upper, and Ns-3 lower based on field relationships (Table 1). Ages of volcanic ashes in Ns-3 range from 10.4 ± 0.10 Ma to 9.8 ± 0.10 Ma. Those from Ns-2 range from 9.59 ± 0.10 Ma to 8.32 ± 0.12 Ma. In addition, Schoenbohm and Carrapa (2015) reported 7.77 ± 0.21 Ma and 7.88 ± 0.58 Ma ashes from unit Ns-1, and a 10.50 ± 0.10 Ma for unit Ns-2 (Table 1).

**40Ar/39Ar Geochronology of Lava Flows and Monogenetic Cinder Cones**

The 40Ar/39Ar geochronology experiments were conducted at the U.S. Geological Survey in Denver, Colorado. For this study, neutron irradiated basalt rock fragments, free of obvious alteration and phenocrysts, of ~1 mm³ were analyzed. A summary of the methods and the tabulated results are given in the GSA Data Repository (see footnote 1).

Unit Nfl was deformed and exhumed in a dome-like feature, as the result of interference between fold 1 and fold 2 structures (see later herein; Fig. 5). Unit Nfl also constitutes the floor of the eastern Pasto Ventura basin (Fig. 5), thus providing important constraints on the timing of deposition and deformation for the Pasto Ventura basin. We sampled unit Nfl from an outcrop immediately below the base of unit Ns-3 (sample PV11B-01) for 40Ar/39Ar dating. A 40Ar/39Ar plateau age (three or more contiguous heating steps comprising ≥50% of the 40Ar released with ages overlapping at the 2σ level of uncertainty) of 11.69 ± 0.01 Ma was determined for this sample (11 heating steps and 55.2% total 40Ar released; Fig. 5).

The study area also contains several monogenetic mafic volcanic cinder cones and associated lava flows (Fig. 2). The cinder cones are relatively small in size (typically 500 to 800 m in diameter), and the flows are usually confined to within 1–3 km from the sourcing cinder cones. They are important indicators for geodynamic processes such as lithospheric foundering and are clearly associated with upper-crustal extension (e.g., Allmendinger et al., 1989; Marrett and Emerson, 1992; Kay et al., 1994; Ruch and Walter, 2010). In this study, we compiled existing geochronological data for the mafic monogenetic volcanism for the Pasto Ventura region, and we add five new 40Ar/39Ar ages (Table 2). Five mafic volcanic...
samples (PV07-PV1-02, P07PV3UR3-1, P07SUR1-01, P09B-09, P09B-12) were collected for geochronology using 40Ar/39Ar step-heating analyses, and these yielded ages of 0.68 ± 0.06 Ma (sample PV07-PV1-02), 0.42 ± 0.05 Ma (sample P07PV3UR3-1), 0.43 ± 0.07 Ma (sample P07SUR1-01), 0.57 ± 0.04 Ma (sample P09B-09), and 0.45 ± 0.02 Ma (sample P09B-12) (Table 2).

Whole-Rock Geochemistry

Two samples of volcanic rocks (PV11B-01 and PV11B-04) obtained from the northeastern and southwestern edges of the mapped Nfl unit, respectively (Fig. 5), were processed for whole-rock geochemistry (Table 3). We also collected one sample from a similar flow to the east of the NPV map area (sample PV10B-03; Fig. 2), which was assigned to a dacite/andesite unit or Tertiary mafic volcanic rocks by previous workers (e.g., Roy et al., 2006; Schnurr et al., 2006). Major and trace elements including rare earth elements (REEs) were determined by whole-rock X-ray fluorescence (XRF) and inductively coupled plasma–mass spectrometry (ICP-MS) at the Geoanalytical Laboratory in the School of Earth and Environmental Sciences at Washington State University (e.g., Johnson et al., 1999; Table 3).

The geochemical similarities between samples PV11B-01 and PV11B-04 confirm our field and petrographic observations for the extent of unit Nfl. Results also show that the newly identified unit Nfl in the center of the NPV map area shares similarities with a sample (PV10B-03) from the previously mapped, voluminous intermediate-mafic unit to the east (e.g., Roy et al., 2006; Schnurr, 2006). Given the very similar geochemical signatures between PV10B-03 and PV11B-01/04, we describe them as one group (called the Nfl samples) in the discussion. The Nfl samples are shoshonitic basaltic trachyandesites. They have low Mg# (20.6–43.4), low MgO (1.2–3.8 wt%), high K2O (3.3–4.1 wt%), and high Al2O3 (17–20 wt%; Table 3). They are depleted in Ni and Cr and high field strength elements (HFSEs) such as Ta, Nb, and Ce. Some of the large ion lithophile elements (LILEs), especially Ba and Nb, are also depleted. The 11.69 ± 0.01 Ma 40Ar/39Ar plateau age of PV11B-01 might be indicative of the more extended Tertiary andesite/dacite unit on the SE edge of the Puna Plateau (Roy et al., 2006; Schnurr, 2006), the age of which has not previously been documented.
Deformation Analysis: Faults, Folds, and Syndepositional Deformation

A major NW-dipping thrust fault (fault 1), with an ~N32°E strike, cuts much of the NPV map, dicing out to the southwest by ~26°50’S, before reaching the SPV map. This thrust fault is a member of the El Peñón–Pasto Ventura fault group (Allmendinger et al., 1989). Although this major thrust fault has undergone recent extensional reactivation (Allmendinger et al., 1989), as evidenced by normal sense displacement of two ca. 0.8 Ma and ca. 0.38 Ma mafic cinder cones to the north (out of the NPV map; Zhou et al., 2013), we focus on its earlier, contractional phase because it appears to have controlled the deformation of the Pasto Ventura basin. It carries bedrock (unit M) in its hanging wall over sedimentary units in its footwall along the western margin of the basin. Because of modern elolian deposition and pervasive physical weathering and colluvial transport of material, we were unable to locate any well-preserved exposures of the footwall plane. However, the fault trace on the surface is clearly evident by a distinct topographic rise, and therefore it enables us to calculate an ~33° dip for fault 1 based on mapped fault trace and topography. To the east of fault 1, a secondary thrust (fault 2) bounds the fold 1 syncline on the west. Fault 2 strikes ~N40°E and dips ~45° to the west based on field observations. It brings unit Ns-3 to the surface and tilts the strata ~28° to 42° dip for fault 1 based on mapped fault trace and topography. To the west. Fault 2 strikes ~N40°E and dips ~45° to the west based on field observations. It brings unit Ns-3 to the surface and tilts the strata ~28° to 42° dip for fault 1 based on mapped fault trace and topography. To the west.

Strata of unit Ns-3 have been folded into a syncline (fold 1) and are locally overturned within ~300 m of fault 1 (Fig. 5). Fold 1, with an ~80° ~W-dipping axial plane, runs across the entire NPV map, with an ~N30°E strike in the north and an ~N10°E strike in the south. This syncline has an axial culmination, and thus younger units make up the core of the syncline in the north and south compared to the central part of the NPV map. From the north to the south, fold 1 becomes more open, with the interlimb angle changing from ~30°–40° to ~125°. Based on cross sections reconstructed from surface data, shortening accommodated by fold 1 is estimated to be ~0.81 km (~32%; Fig. 5, section A–A′) between fault 1 and fault 2 and ~0.94 km (~34%; Fig. 5, B–B′) between fault 1 and fault 3 farther south. We suggest that there is a WNW-ESE–directed, open anticline (fold 2), which subsequently folded fold 1 and is responsible for forming the axial culmination of fold 1. Volcanic unit Nfl is thus exposed at the base of the eastern section in an interference dome-like structure. This second folding event postdates most of the ESE-WNW–directed deformation in the region.

Fault 3 is paired with a syncline in its hanging wall to the east (fold 3), which is open in shape and has only minor significance for accommodating shortening (~10%). Fault 3 dies out and is replaced by a south-plunging anticline (fold 4) to its south. The axial plane of fold 4 is vertical with an interlimb angle of ~150°. The fold 4 anticline is paired with a syncline (fold 5) to its east, which is also south-plunging and gently folded. Across the cross section C–C′ (Fig. 5), the shortening accommodated by fold 1, fold 4, and fold 5 is ~0.46 km (10%). Unit Nfl, the volcanic flow, contains ~50-cm-spaced layers bounded by parallel parting surfaces that consistently dip ~22° to the east along the eastern edge of the unit, similar to the underlying unit Ns-3 (Fig. 4G). On satellite and air-photo images, the layers can be traced along strike, following the base of the mapped unit Ns-3, forming “V-shaped” curvatures at river valleys. The orientation of the layering in unit Nfl is close to the orientation of bedding of the base of overlying unit Ns-3, consistent with primary lavas flow within unit Nfl that were once flat-lying. Deformation by approximately NNE-striking folding and thrust faulting, and possibly subsequent anticline folding (fold 2), tilted unit Nfl. Therefore, unit Nfl (11.7 Ma, see following sections) predates deformation of units Ns-3 to Ns-1 in the Pasto Ventura basin.

In southern Pasto Ventura (the SPV map; Fig. 6), fault 4 and fault 5 thrust faults mark the southern continuation of fault 1 from the NPV map.
Fault 4 brings the bedrock (unit M) to the surface in its hanging wall, whereas fault 5 is confined to unit Ns-1. We suspect both faults dip relatively steeply because of their linear map trace, and they die out rapidly to the south. The anticline, fold 8, dominates the deformation in the SPV map. The axial plane of fold 8 strikes ~N20°–30°S, although the trace is somewhat sinuous. This fold brings the lower unit, Ns-3, to the surface. In the middle of the SPV map, strata from two limbs of fold 8 dip toward the NE and NW, indicating this fold is plunging to the north. Fold 8 is open in the south (interlimb angle ~110°) and becomes tighter to the north (interlimb angle ~40°). Fold 8 is paired with syncline fold 6 in the northwest and syncline fold 7 in the middle and south of the map. Axial planes of both fold 6 and fold 7 strike ~N30°E, parallel with syncline fold 8, and are tight (interlimb angle ~70°–80°). The axial planes of fold 6, fold 7, and fold 8 are subvertical (~80°, ~85°, ~80°, respectively), but dip slightly to the west. The amount of shortening estimated from the cross section to the north (D–D') is ~1.45 km (~47%) and from the south section (E–E') is ~3 km (~45%; Fig. 6).

The mapped NNE-striking, dominantly WNW-dipping thrust faults and associated folds suggest that the Pasto Ventura basin was deformed primarily under approximately WNW-ENE contraction. We are able to constrain the amount of shortening based on cross sections. The amount of approximately E-W shortening ranges from ~35%–45% to 10%–15% (Figs. 5 and 6). Some minor NNE-SSW contraction is possible as a later phase of deformation, evidenced by the formation of a WNW-ENE-striking fold 2 anticline in the NPV map.

Several lines of evidence point to syndepositional deformation in the Pasto Ventura basin. In the SPV map, an angular unconformity formed within the Ns-3 mudstone-siltstone (Figs. 7A and 7B). Units below and above the angular unconformity are indistinguishable in lithology but are oriented differently (the beds above the unconformity are oriented 230°/66°; the beds below the unconformity are oriented 216°/28°; Figs. 7A and 7B). SIMS-dated zircons from ashes above this unconformity yield an age of 9.22 ± 0.08 Ma (sample PV11A-06) and 9.17 ± 0.10 Ma (sample PV11A-61), respectively. Taken together, these internal unconformities and age data for interbedded volcanic ashes imply that deposition and deformation must have taken place synchronously in the Pasto Ventura basin around ca. 9–10 Ma.

**DISCUSSION**

**Timing of Formation and Deformation of the Pasto Ventura Basin (NW Argentina)**

The sedimentation history for the Pasto Ventura basin began with deposition of the oldest sedimentary unit in the basin, unit Ns-4. We do not have direct age constraints for unit Ns-4, and no stratigraphic relationship between units Ns-4 and Nfl has been observed in the field. However, we argue that unit Ns-4 likely predates unit Nfl (11.69 ± 0.01 Ma, sample PV11B-01 in this study) given the fact that the Puna Plateau started to be characterized by felsic volcanism since 20–16 Ma (mostly since 14–12 Ma), and unit Ns-4 does not contain any of the volcanic ashes so prevalent in younger strata (e.g., Kay et al., 2010). The presence of paleosol horizons within unit Ns-4 (Fig. 4F) indicates that the Pasto Ventura basin was undergoing slow accumulation and prolonged subaerial exposure, with possible periods of erosion or at least sedimentary hiatuses. Following deposition of unit Ns-4, a basaltic trachyandesite flow (unit Nfl) covered the basin at ca. 11.7 Ma. The onset of major accumulation in the basin, units Ns-3, Ns-2, and Ns-1, which consist of fluvial/alluvial, lacustrine, and eolian sediments, started after ca. 11.7 Ma and before ca. 10.5 Ma, the ages of the oldest volcanic ashes obtained from the Pasto Ventura basin (Table 1; sample P-Ash-01 in this study; sample PVN75 in Schoenbohm and Carrapa, 2015). Basin strata continued to accumulate until at least ca. 7.8 Ma, the age of the youngest ash dated in the basin (Schoenbohm and Carrapa, 2015). As no outcrop of any younger basin fill is observed in this region and the paleoenvironment for Ns-1 is similar to modern conditions in the Pasto Ventura region, we infer that unit Ns-1 is the last unit deposited in the Pasto Ventura basin and continued to be deposited after 7.77 ± 0.21 Ma (Schoenbohm and Carrapa, 2015).

Volcanic flow unit Nfl floors most of the exposed Pasto Ventura basin, dips consistently with immediately overlying unit Ns-3, and was tilted during basin deformation (Fig. 5). Therefore, its age of 11.7 Ma must predate the onset of the late Miocene syndepositional deformation in the Pasto Ventura basin. Angular, internal unconformities indicate ongoing syndepositional deformation of the Pasto Ventura basin from 10 to 9 Ma. The major thrust fault in the NVP map developed around the same time and truncates a 9.51 ± 0.10 Ma ash (sample PV11A-07) in its footwall. Unit Nfl, underlying unit Ns-3, was possibly folded by a NNE-SSW–striking anticline, enhancing its exposure at the surface. The influence of this anticline is minor and could be coeval with fault 2 and fault 3, but its exact timing is unclear. In the southern NVP map (Fig. 4), an ash close to the core of a syncline is dated at 8.32 ± 0.12 Ma (sample PV10A-02), and Schoenbohm and Carrapa (2015) documented tilted strata younger than 7.77 ± 0.21 Ma (in Ns-1 eolian sandstone). We therefore infer that contractional deformation, in the form of thrust faulting, folding, tilting, and formation of internal unconformities, initiated between 11.69 ± 0.01 Ma and 9.99 ± 0.12 Ma and continued until after 7.77 ± 0.21 Ma in the Pasto Ventura basin. Maximum shortening occurred in the southern part of the basin, in which 47% shortening took place in a WNW-ENE direction (Fig. 6).

Across the Puna Plateau, contractional deformation ended and was succeeded by extensional deformation during the late Miocene and

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**Figure 7. Uninterpreted (A, C) and interpreted (B, D) field photos for syndepositional structures and growth strata. Photo locations are marked in Figure 5 (for C) and Figure 6 (for A).**
Late Miocene southern Puna Plateau deformation

Research

Early Pliocene (e.g., Allmendinger et al., 1997). We use the age of recent regional extension of the southern Puna Plateau to constrain the cessation of contractional deformation of the Pasto Ventura basin. One important indicator constraining the timing of regional extension is the age of post–mid-Miocene monogenetic mafic volcanism, because dense mafic magmas could not have traveled easily through continental crust under a compressional stress regime (Allmendinger et al., 1997; Marrett and Emerman, 1992) and because of the association of mafic cinder cones with normal and strike-slip faults (e.g., Allmendinger et al., 1989). Several authors have suggested that the shift from contraction to extension likely took place diachronously across the plateau (Marrett and Emerman, 1992; Montero-López et al., 2010; Riller and Onccken, 2003; Riller et al., 2001; Risse et al., 2008; Schoenbohm and Strecke, 2009; Eckelmann et al., 2013; Zhou et al., 2013). Fault 1 cuts two mafic cinder cones with 0.76 ± 0.16 Ma and 0.47 ± 0.03 Ma ages to the north of the NPV map and offsets them by different amounts in a normal, right-lateral sense (Zhou et al., 2013), which indicates ongoing extension in the Pasto Ventura region between ca. 0.8 and 0.5 Ma. Other crosscutting relationships between dated units and extensional faults indicate the extension from south of 26°S on the southern Puna Plateau was established by 5 Ma (north of the Pasto Ventura basin; Montero-López et al., 2010) and 4 Ma (south of the Pasto Ventura basin; Schoenbohm and Strecke, 2009). Taken together, we conclude that the onset of extension, and thus termination of the contractional deformation in the Pasto Ventura region, likely took place between 7.8 Ma and 5–4 Ma based on regional observations (Montero-López et al., 2010; Schoenbohm and Carrapa, 2015; Schoenbohm and Strecke, 2009), and certainly by 0.8 and 0.5 Ma based on constraints within the basin (Zhou et al., 2013).

Mafic Volcanism on the Southern Puna Plateau

On the southern Puna Plateau, the most recent phase of mafic volcanism is represented by widely distributed, late Miocene monogenetic cinder cones and lava flows (e.g., Allmendinger et al., 1997; Kay et al., 1994). The volcanic rocks have experienced little erosion and are associated with crustal extension, as described already. The oldest post–mid-Miocene monogenetic mafic volcanism clusters around the Antofagasta-Antofalla region (~26°S; Fig. 8A; Risse et al., 2008; Schoenbohm and Carrapa, 2015; Zhou et al., 2013), with ages of up to 8.7 ± 0.04 Ma (Schoenbohm and Carrapa, 2015). The Pasto Ventura region is located in the southern-most Puna Plateau, and our newly dated samples in this study agree with the results from previous studies in the area (Table 2; Fig. 8A; Ducea et al., 2013; Risse et al., 2008; Zhou et al., 2013; this study). The post–8.7 Ma mafic magmatism on the southern Puna has been linked to lithospheric foundering, notably because of a lack of geochemical evidence for subduction (described as intraplate volcanism; Kay et al., 1994). Recent

Figure 8. Mafic volcanism on the southern Puna Plateau. (A) Geochronology for post–mid-Miocene monogenetic mafic volcanism on the Puna Plateau (Risse et al., 2008; Ducea et al., 2013; Kay et al., 1994; Schoenbohm and Carrapa, 2015; Zhou et al., 2013; this study). Circles denote the geographic distributions for all existing geochemical analyses for this volcanic group. (B–F) Selected geochemical plots, denoting contrasts between NF samples and post–mid-Miocene monogenetic mafic volcanics (compiled from Ducea et al., 2013; Drew et al., 2009; Kay et al., 1984, 1999; Knox et al., 1989; Risse et al., 2013; Murray et al., 2015). REE—rare earth elements. Plots in Figs. E, F are primitive mantle normalized (Sun and McDonough, 1989).
geochemical studies indicate that the melts for the small-volume, monogenetic mafic volcanics were derived primarily from pyroxenites located in the lower parts of the lithosphere (Ducea et al., 2013; Murray et al., 2015), and they seem to have become progressively hotter within a short, ~1.7 m.y. time window (Ducea et al., 2013). This model suggests that the foundered lower lithospheric blocks were likely small in size (~50 km; Ducea et al., 2013; Murray et al., 2015).

The 11.7 Ma unit Nfl lava in this study represents a distinct group of voluminous mafic volcanics on the southern Puna Plateau that are overlain by sedimentary units and that were incorporated in upper-crustal deformation (Figs. 2A and 5). They outcrop mostly along the southern margin of the Puna Plateau (Fig. 2A; Roy et al., 2006; Schnurr et al., 2006) and are older than late Miocene (this study). They are geochemically distinct from the post–mid-Miocene monogenetic mafic volcanics (Fig. 8). The lower MgO and Mg# indicate that they are more evolved than post–mid-Miocene monogenetic lavas. The less depleted HREEs in the Nfl samples imply that the melts were unlikely to have been sourced from a garnet-bearing source indicative of lithospheric foundering, such as eclogite (Lee et al., 2006; van Westrenen et al., 2001), as were the younger mafic volcanics in the region. While the origin of Nfl lavas is not resolvable in this study, they are not readily explained by the same models invoked for the post–mid-Miocene monogenetic magmas. The less depleted HREEs in the Nfl samples are older than late Miocene (this study). They are geochemically distinct from the post–mid-Miocene monogenetic mafic volcanics and are more evolved than post–mid-Miocene monogenetic lavas (Fig. 8). The lower MgO and Mg# indicate that they are more evolved than post–mid-Miocene monogenetic lavas. The less depleted HREEs in the Nfl samples imply that the melts were unlikely to have been sourced from a garnet-bearing source indicative of lithospheric foundering, such as eclogite (Lee et al., 2006; van Westrenen et al., 2001), as were the younger mafic volcanics in the region. While the origin of Nfl lavas is not resolvable in this study, they are not readily explained by the same models invoked for the post–mid-Miocene monogenetic mafic volcanism (from the same region (Drew et al., 2011). This model emphasizes flexural compensation of the success of this model based on the onset of sedimentation. Thus, although we cannot exclude this model based on available data, we argue that other models, which incorporate wedge dynamics and lithospheric foundering, have the potential to explain more of the observations.

A second set of models characterizes the central Andes as a classic orogenic wedge, highlighting west-to-east propagation of a deformation front and initiation of Cenozoic sedimentation, albeit at an unsteady pace (e.g., Arriagada et al., 2006; Carrapa et al., 2011; DeCelles et al., 2011; Horton and DeCelles, 2001; McQuarrie, 2002a). In the northern Puna (~24°S–25°S), the succession of sedimentary units across the eastern Puna Plateau and adjacent lowlands has been interpreted to result from an eastward-migrating foreland basin system, and it is thought to be the southern extension of the foreland basin system in the Altiplano, despite the striking difference in deformation style as compared to the Altiplano (DeCelles et al., 2011). This model emphasizes flexural compensation of the lithosphere due to end loading and involves age reinterpretation of the Santa Barbara Subgroup, which was proposed to be a product of thermal

Figure 9. Schematic illustration of multiple processes operating in the southern Puna Plateau (not to scale). Dynamic processes for the Puna Plateau are listed in the lower panel.
subsidence following the demise of the Salta Rift (Marquillas et al., 2005). In this model, the Western Cordillera was deforming by early Paleogene time (e.g., Arriagada et al., 2006; Claudouhos et al., 1994; Mpdodiz et al., 2005; Kennan et al., 1995). Deformation migrated quickly across the plateau but stalled in the Eastern Cordillera from ca. 25 to 19 Ma (e.g., Deeken et al., 2006). Deformation has been propagating east through the Subandes and Santa Barbara Ranges since ca. 15–8 Ma (Fig. 3; e.g., Echavarría et al., 2003; Marrett et al., 1994; McQuarrie, 2002a; Pearson et al., 2013). However, for the southern Puna Plateau, especially for regions south of ~26°S, a similar pattern is difficult to establish given the sparse data (Carrapa et al., 2011; DeCelles et al., 2011). Around the Salar de Antofalla (~26°S; ANT, Fig. 3), ~130 km NW of the Pasto Ventura region, several studies have suggested a foreland-like sedimentary record containing Upper Eocene upward-coarsening units with growth strata (Adelmann, 2001; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002). Although the record is incomplete in the Pasto Ventura basin, the older, fine-grained, paleosol-bearing depositional unit Ns-4 might reflect forebulge deposition. It likely predates felsic volcanic activity in this region, which began in the early Miocene (Kay et al., 2010). After 11.7 Ma, the Pasto Ventura basin was filled with fluvialuviclastite-eolian, mostly medium- to fine-grained sediments and was accompanied by syndepositional deformation. In a foreland system, wedge-top sediments are commonly associated with syndepositional deformation due to the active thrust faulting in the deformation front, but they are typically coarser than we observe in the Pasto Ventura basin, given their alluvial and fluvial origins in proximity to high topographic relief (DeCelles and Giles, 1996). Moreover, if the southern Puna was part of the large-scale, N-S–striking foreland basin system in the central Andes (e.g., DeCelles et al., 2011), the deformation front during late Eocene–late Oligocene time should already have been located in the Eastern Cordillera, east of the Pasto Ventura region, as is seen in the northern Puna and the Altiplano (e.g., Carrapa et al., 2011; Horton, 2012). The post–11.7 Ma basin formation and deformation events in the Pasto Ventura region therefore need an alternative explanation.

Other models emphasize out-of-sequence deformation, describing thick-skinned, post-Miocene deformation as nonsystematic and pure shear–like (e.g., Allmendinger and Gubbels, 1996). A study in the Eastern Cordillera at ~23°S suggests the development of hinterland basins by fold-and-thrust deformation and fault reactivation on the eastern flank of the Puna Plateau during the Miocene (Siks and Horton, 2011). Another recent model characterizes the Eastern Cordillera at ~25°S as a broken foreland, where deformation and relief development took place along steeply dipping inherited faults from the Salta Rift, which do not necessarily show any directional gradients in timing (Hain et al., 2011). Although different authors highlight various specific mechanisms, these irregular basement uplift and deformation models agree on the important role of inversion of preexisting Salta Rift structures (e.g., Carrapa et al., 2014; Cristallini et al., 1997; Deeken et al., 2006; Hong et al., 2007; Insell et al., 2012; Riller et al., 2012; Riller and Oncken, 2003). However, it is not clear if the southern Puna Plateau inherited any structures from the Salta rift during the Cretaceous, which is underscored by the lack of Cretaceous sediments or exhumation (Carrapa et al., 2014; Löbens et al., 2013; Marquillas et al., 2005), meaning these models may not be readily applicable to our study area.

Out-of-sequence deformation may also be a result of lithospheric foundering. A key prediction for the formation and detachment of a lithospheric drip is that during drip formation, the upper crust would experience contraction accompanied by sedimentary basin formation, and after drip detachment, it would experience extension, basin inversion, and ignimbritic and mafic volcanism (e.g., DeCelles et al., 2009, 2015; Elkinstanton, 2007; Göğüş and Pyškyłwec, 2008; Kay et al., 1994; Krystopowicz and Currie, 2013; Schoenbohm and Carrapa, 2015). Kay et al. (1994) first proposed that the lithosphere beneath the Puna Plateau may have thinned since the mid-Miocene through delamination, based on geochemical signatures of mafic volcanism. Mineralogical and geochemical studies also reveal that ignimbrite complexes such as Agua Calientes and the Cerro Galan may be related to melting due to asthenospheric upwelling in the wake of foundering lithosphere (Kay et al., 2010, 2011). Geophysical studies have supported this hypothesis, documenting thinned crust and lithosphere (Tassara et al., 2006; Yuan et al., 2002) and potential foundered blocks in the asthenosphere (Bianchi et al., 2013; Calixto et al., 2014; Heit et al., 2014). In the northern Puna region, structural data suggesting out-of-sequence deformation and basin formation in the Salar de Arizaro region have been interpreted to reflect lithospheric foundering (DeCelles et al., 2015). In the southern Puna, the spatio-temporal pattern and geochemistry of small-volume mafic volcanics, and surficial extensions are consistent with the formation of small (~50-km-diameter) lithospheric drips (Fig. 8A; Ducea et al., 2013; Marrett et al., 1994; Murray et al., 2015; Risse et al., 2008; Schoenbohm and Carrapa, 2015; Zhou et al., 2013).

A small-scale, late Miocene lithospheric dripping event beneath the southern Puna Plateau may be responsible for the late Miocene upper-crust deformation in the Pasto Ventura region. Subsidence and contraction of the basin since 11.7–10.5 Ma and until 7.8 Ma may reflect the formation of a drip. After the drip detached, the contractional deformation was replaced by upper-crustal extension, evidenced by transtensionally reactivated older thrust faults and the eruption of monogenetic mafic lavas (Allmendinger et al., 1989; Schoenbohm and Streek, 2009; Zhou et al., 2013). Evidence exists to support ongoing upper-crustal Quaternary extension in the Pasto Ventura region (Zhou et al., 2013; this study). Additionally, geochemical signatures from deformed basin-floor unit Nfl are distinct from those of recent monogenetic mafic lavas, which presumably resulted from lithospheric foundering (Fig. 8; e.g., Drew et al., 2009; Ducea et al., 2013; Kay et al., 1994; Murray et al., 2015), and may suggest alternative origins for 11.7 Ma Nfl lavas, prior to the detachment of the late Miocene lithospheric drip. The center of this drip may be located beneath the northern part of the Pasto Ventura region, around ~26°S, which is evidenced by the oldest post–mid-Miocene monogenetic mafic volcanism (Schoenbohm and Carrapa, 2015) and by an azimuthal change in recent extension of the southern Puna Plateau (Zhou et al., 2013). This inferred location is also close to the Cerro Galan, which erupted 6.6–2.06 Ma and is thought to be a result of lithospheric foundering (Kay et al., 2010).

Lithospheric foundering (Kay et al., 1994) may be a key element in controlling upper-crustal deformation in the central Andes, acting together with other important geodynamic processes (Fig. 9; e.g., Allmendinger et al., 1997; DeCelles et al., 2011; Horton, 2005; Riller and Oncken, 2003; Sobel et al., 2003; Streek et al., 2007), all of which have been supported by geological observations from various locations, and all of which may contribute to the development of the Puna Plateau. Particularly in the Pasto Ventura region, the nature and timing of basin sedimentation and deformation, when combined with other geophysical and chemical data, are supportive of models for a small-scale lithospheric drip beneath the southern Puna Plateau during the late Miocene (Schoenbohm and Carrapa, 2015).

CONCLUSIONS

Sedimentary basins on the Puna Plateau, although isolated and with only limited exposure, serve as important recorders of upper-crustal deformation and geodynamic processes in the central Andes. This work documents a basin formation and deformation event in the Pasto Ventura region of the southern Puna Plateau. Most recent sediment accumulation within the basin started after ca. 11.7 Ma (the age of the underlying basaltic trachyandesite flow, unit Nfl) and before ca. 10.5 Ma (the oldest volcanic
ash from unit Ns-3) and consists of fluvial/alluvial, lacustrine, and eolian sedimentary rocks. The youngest volcanic ash is dated at ca. 7.8 Ma from unit Ns-1, and no outcrop of younger basin fill is observed in this region. The Early Miocene basin likely reflects formation and detachment of a lithospheric drip in the central Andes: Evidence from sedimentary rock provenance and apatite fission track thermochronology in the Fiambalá Basin, southernmost Puna Plateau margin (NW Argentina): Earth and Planetary Science Letters, v. 247, p. 82–100, doi:10.1016/j.epsl.2006.04.010.


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MANUSCRIPT RECEIVED 19 MAY 2014
REVISED MANUSCRIPT RECEIVED 23 JANUARY 2015
MANUSCRIPT ACCEPTED 23 FEBRUARY 2015
Printed in the USA