Regional Deformation and Dynamic Processes of the
Southern Puna Plateau, Central Andes

by

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for the degree of Doctor of Philosophy

Department of Earth Sciences
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Abstract

The southern Puna Plateau of the central Andes offers an excellent opportunity to (1) advance our understanding in orogenic models for non-collisional mountain belts, (2) test Earth’s upper-crustal responses to lower lithospheric processes, and (3) explore uses of low-temperature thermochronology in studying landscape evolution. This thesis provides important constraints on regional deformation and landscape evolution for the southern Puna Plateau since the late Paleozoic, and focuses on reconstructing its Cenozoic history in detail, using field, laboratory and numerical methods.

Sedimentary data, sandstone modal composition, detrital zircon U-Pb geochronological and detrital apatite fission-track thermochronological data suggest that the oldest sedimentary units on the SE (southeastern) and SW (southwestern) Puna Plateau belong to upper Eocene to lower Oligocene, defining a regional sedimentary basin across the plateau region during ~38-28 Ma. Provenance analysis suggests that this basin was sourced from active orogenic front to the west and re-activated relict landscape to the east. During the late Oligocene (~25 Ma) to mid-late Miocene (~8 Ma), the southern Puna Plateau started to be compartmentalized by several major, ~N-S-trending intervening ranges, such as the Sierra de Calalaste and Sierra Laguna Blanca, documented by bedrock apatite fission-track and (U-Th-Sm)/He thermochronology. Field
mapping, structural analysis, zircon U-Pb and whole-rock $^{40}$Ar/$^{39}$Ar dating reveal that the southernmost Puna Plateau experienced upper-crustal contraction during the late Miocene (~12-8 Ma). Before the late Eocene, the region of the southern Puna Plateau was home to several episodes of exhumation and related topographic highs. Using bedrock apatite fission-track and zircon and apatite (U-Th-Sm)/He thermochronology and modeling, this thesis documents evidence for major topographic highs during the late Paleozoic and the late Cretaceous, and refines paleogeographical models for the southern central Andes. Particularly, the late Cretaceous exhumation related to the development of rift shoulders has caused larger amounts of exhumation compared to that introduced by Cenozoic mountain building processes.
致我的父亲和母亲
I am grateful for having an opportunity to live in Toronto where I spent the most exciting years in my 20s that I would never trade for anything. I would have not possibly done this without love and understanding from my family on the other side of the globe.

I heartily thank Lindsay Schoenbohm for being the best advisor that I could ever ask for (and arguably the best advisor I have ever heard of). Lindsay holds broad interests in tectonics and has served as a superb intellectual source. My project has greatly benefited from the freedom, advice and support provided by Lindsay. I would have not accomplished this degree without her encouragement and patience. My life would be quite different if Lindsay were not my advisor.

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Appendix D: Supplementary materials for Chapter 3, including sandstone modal composition data, detrital and bedrock U-Pb geochronology data, results for the K-S test of U-Pb ages, results of apatite fission-track thermochronology (single-grain ages and track-length measurements)

Appendix E: Supplementary materials for Chapter 4, including apatite fission-track data (single-grain ages and track-length measurements), AFT age-D_{par} plots, inverse and forward modeling results, time-temperature plots for forward models.

Appendix F: Supplementary materials for Chapter 5, including data for U-Pb geochronology, Ar/Ar geochronology, and volcanic samples. This item is identical to GSA (Geological Society of America) Data Repository Item 2015112, available at www.geosociety.org/pubs/ft2015.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.
CHAPTER 1

INTRODUCTION
1 Introduction

1.1 Background and Motivation

1.1.1 Orogenic models in the central Andes

The Andes in South America are Earth’s highest and longest mountain belts associated with ocean-continent plate convergence. They have evolved along the western margin of the South America plate that is underlain by the subducting Nazca plate (Fig. 1). At ~15-27°S, the central Andes host the Altiplano-Puna Plateau (the Central Andean Plateau), the world’s second largest continental plateau after the Tibetan Plateau. The Altiplano-Puna Plateau is internally-drained with an elevation over 3 km (Allmendinger et al., 1997; Isacks, 1988; Strecker et al., 2007) (Fig. 1A). Divided at ~22.5°S latitude into the Altiplano Plateau to the north and the Puna Plateau to the south, the entire plateau is bounded by higher, more rugged Western and Eastern Cordilleras; at its southern end it transitions into the Sierras Pampeanas (Fig. 1A). The central Andes vary structurally along-strike. The Altiplano Plateau, the adjacent Eastern Cordillera and the Subandes are thin-skinned, with fold-and-thrust belts developing mostly within the sedimentary cover and evolving along a regional décollement (e.g., McQuarrie et al., 2005). The Puna Plateau and its adjacent Eastern Cordillera is thick-skinned, with ~N-S-trending bedrock ranges bounded by deep-seated faults (e.g., Strecker et al., 2012). To the south, the Sierras Pampeanas is also thick-skinned, but is characterized with widely-spaced bedrock ranges bounded by thrust faults that possibly cut the entire crust (e.g., Jordan and Allmendinger, 1986) (Fig. 1).

In the central Andes, studies of regional kinematics, exhumation, sedimentation and dynamic processes have been conducted for decades, leading to the development of several classes of orogenic models that are supported by case studies throughout the central Andes (Fig. 1C) and are under active debate. The purpose of such debates is not simply to distinguish which model is “correct”, but rather to gain insight into the complex orogenic processes operating in the central Andes at various spatial scales and during different time periods.
A first class of models views the evolution of the central Andes as a result of an easterly-propagating foreland basin system starting during the early Cenozoic, analogous to a retro-arc orogenic wedge (e.g., DeCelles et al., 2011; Horton 2005). Proponents of this model suggest that the orogenic front initiated in the western Altiplano-Puna Plateau at ~50-60 Ma and propagated to the east through time (e.g., Carrapa et al., 2011; DeCelles et al., 2011; Echavarria et al., 2003; McQuarrie et al., 2005). As the orogenic front evolved, the lithosphere flexed and accommodated a ~N-S-striking foreland basin system, with sedimentary records documented in locations in the eastern Altiplano Plateau, the Eastern Cordillera and the northern Puna Plateau (e.g., DeCelles et al., 2007; Horton, 2005; Reynolds et al., 2001). The changing 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, as predicted in wedge dynamics (e.g., Dahlen, 1990; Garzione et al., 2006; Kay et al., 1994). Furthermore, formation of a regional ~N-S-trending foreland basin system suggests that the lithosphere flexed similarly in response to crustal load,
despite the change from thin- to thick-skinned structural conditions along-strike (e.g., Allmendinger and Gubbels, 1996; DeCelles et al., 2011).

A second class of models views the present-day Puna Plateau, its adjacent Eastern Cordillera and the Sierras Pampeanas as thick-skinned, characterized by compressional basin-and-range morphology, with sedimentary basins bounded by ~N-S-oriented bedrock mountain ranges (Fig. 1). This thick-skinned tectonic style may reflect reactivation of structures inherited from the complex extensional and transtensional tectonism in northwestern Argentina, such as the Salta rift that evolved in Cretaceous-Paleogene time (~150-60 Ma) 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). The availability of pre-existing weaknesses may lead to out-of-sequence deformation, with early Cenozoic deformation notably taking place hundreds of kilometers inland of the plate boundary (e.g., Carrera et al., 2006; Grier et al., 1991; Hongn et al., 2007; Kley et al., 2005). Such models favor inversion tectonics (e.g., Hongn et al., 2007) and related “broken foreland” dynamics for the Puna Plateau and Eastern Cordillera (Sobel and Strecker, 2003; Sobel et al., 2003; Strecker et al., 2012). The broken foreland landscape, which consists a set of variably connected, laterally restricted depocenters, isolated by bedrock mountain ranges, evolves with distributed range uplift and limited flexural accommodation space (Hain et al., 2011; Strecker et al., 2012).

A third class of models suggests an alternative dynamic control for the thick-skinned system in the Sierras Pampeanas (Fig. 1). Although the Sierras Pampeanas also hosts widely-spaced bedrock ranges that are bounded by thick-skinned structures, this region is characterized by a much broader areal extent: basement ranges are broadly distributed for hundreds of kilometers from the coast. This region is not associated with recent magmatism and lies above a low angle segment of the Nazca plate. The evolution of the Sierras Pampeanas is therefore argued to be controlled by low-angle (or “flat”) subduction, which drives distributed range uplift due to mechanical coupling between underlying and overriding plates (e.g., Jordan et al., 1983). Because the location of low-angle subduction changes through time (e.g., Ramos and Folguera, 2009), it may exert an important dynamic control on tectonic and magmatic activities for a broad region in the central Andes, as emphasized by this class of models (e.g., Kay and Coira, 2009; Ramos and Folguera, 2009).
1.1.2 Surficial diagnostics of lower lithospheric processes

Because the Earth's subsurface lies hidden from view, we are left with indirect observations to study deep-seated processes. Lower lithosphere foundering is one such process, in which the lower part of the lithosphere, mainly the lithospheric mantle, ‘sinks’ into the asthenosphere. There are two major mechanisms hypothesized for lower lithospheric foundering: (1) delamination and (2) lithospheric dripping through formation of a viscous Rayleigh-Taylor instability (Fig. 2). During delamination, the cold and dense mantle lithosphere peels away as a coherent slice from the crust along the Moho, and the removed slice of mantle lithosphere is replaced by hot and buoyant asthenosphere (Bird, 1979) (Fig. 2). Lithospheric dripping refers to viscous convective removal of the mantle lithosphere, where some or all of the mantle lithosphere may be removed as cold dense “drips” as a Rayleigh-Taylor gravitational instability (Fig. 2) (Houseman et al., 1981; Houseman and Molnar, 1997).

Figure 2. Surficial expression of lower lithospheric foundering events derived from numerical and analog models (DeCelles et al., 2015, from Göğüş and Pysklywec, 2008; Krystopowicz and Currie, 2013; Molnar and Houseman, 2014; Wang et al., 2015).

Fingerprinting lower lithospheric foundering events on geological timescales is difficult. Geophysical imaging is useful for detecting recent or on-going lower lithospheric foundering events (e.g., Bianchi et al., 2013; Zandt et al., 2004), but is inadequate on longer timescales
because of the recovery of the lower lithosphere through time. Geochemical investigations have also played a vital role in exploring lower lithospheric foundering on geological timescales, because the upwelling of asthenosphere after the foundering event may result in distinct, mantle-derived volcanism preserved on the Earth’s surface (e.g., Ducea et al., 2013; Elkins-Tanton, 2007; Kay and Kay, 1993; Kay et al., 1994).

Recent advances in numerical and analog modeling offer specific predictions for upper-crustal behaviors involving different lithospheric foundering scenarios (Fig. 2) (e.g., Göğüş and Pysklywec, 2008; Krystopowicz and Currie, 2013; Molnar and Houseman, 2014; Wang et al., 2015). Therefore, geological events observed on the Earth’s surface, such as basin subsidence and inversion, upper-crustal contraction and extension, may be used to test various lithospheric foundering events in geological timescales, leading to an exciting, relatively new direction in structural geology and tectonics.

Lower lithospheric foundering has been argued to be a cause for deforming Earth’s upper crust, resulting in contractional/extensional events, or subsidence/uplift of Earth surface. For example, wholesale lower lithospheric foundering beneath the Altiplano plateau may have caused surface uplift during the late Miocene (~11-7 Ma) due to post-foundering buoyancy (Garzione et al., 2008). Recent, ongoing surficial extension of the Puna Plateau may reflect this recent lower lithospheric foundering event as well (Kay et al., 1994; Schoenbohm and Strecker, 2009; Zhou et al., 2013).

1.1.3 Using low-temperature thermochronology to study landscape evolution

Low-temperature thermochronology utilizes temperature-dependent retention of radiation products, such as fission tracks and radiogenic $^4$He in apatite and zircon. Above a particular temperature, decay products are not preserved over geological timescales and thermochronological ages are reset to zero (Fig. 3). When cooled through a range of temperatures, known as the partial annealing zone (PAZ) for fission-track systems and the partial retention zone (PRZ) for (U-Th-Sm)/He systems, decay products start to accumulate within minerals. Typically, the PRZ for zircon (U-Th-Sm)/He system extends from $\sim$200 to 130 °C (Wolfe and Stockli, 2010); the PAZ for apatite fission-track system extends from $\sim$140-120 to 60 °C (Ketcham et al., 1999); and the PRZ for apatite (U-Th-Sm)/He system extends from $\sim$85 to
40°C (Ehlers and Farley, 2003). A measured thermochronological age reflects the time and rate of the sample’s cooling path through the PAZ or PRZ. Assuming a monotonic and steady cooling history, the calculated thermochronological age can be viewed as the time since the sample cooled through a particular closure isotherm, defined as the effective closure temperature (e.g., Dodson, 1973; Ehlers and Farley, 2003; Reiners and Brandon, 2006). The effective closure temperate lies within the corresponding PAZ or PRZ and is influenced by the cooling rate, with the faster cooling rate resulting in a higher effective closure temperature in the same thermochronological system (e.g., Dodson, 1973; Ehlers and Farley, 2003; Reiners and Brandon, 2006). Given a modest-fast, 10 °C/Myr cooling rate, the effective closure temperature for zircon (U-Th-Sm)/He, apatite fission-track and apatite (U-Th-Sm)/He systems are calculated to be approximately 183 °C, 116 °C and 67 °C, respectively (Reiners and Brandon, 2006).

![Figure 3. A summary of geo-/thermochronometers with their ranges of closure temperatures.](image)

By constraining a sample’s temperature history, low-temperature thermochronological data provide important information regarding the sample’s exhumation history with certain assumptions. For example, when assuming a geothermal gradient of 20-30 °C/km and vertical exhumation pathways, the zircon (U-Th-Sm)/He age for a sample reflects exhumation from ~8-12 km in depth, driven by tectonic or erosional processes. For bedrock samples that were once fully reset, measured thermochronological ages reflect the timing and the rate of their in situ exhumation history (e.g., Reiners and Brandon, 2006). For a detrital sample that has not been
deeply buried and re-heated, single-grain thermochronological ages are a sum of the exhumation histories of source regions.

Thermochronological data can be further interpreted to constrain landscape evolution, because low-temperature thermochronometers are sensitive to subsurface thermal field changes induced by modifications of Earth’s landscape. A suite of samples obtained from a region or a vertical transect can thereby play an important role in quantitatively reconstructing the landscape history, particularly when combined with landscape evolution models that incorporate principles of heat conduction and advection (e.g., Braun et al., 2012; Ehlers, 2005).

In many cases, samples may experience complex, non-monotonic cooling histories. For apatite fission-track system, the fission-track length distribution serves as an independent constraint on the sample’s thermal history, in addition to the measured age (e.g., Armstrong, 2005; Gleadow et al., 1986). For example, if a rock is subjected to elevated temperatures, the existing fission tracks are shortened progressively and eventually erased by the thermal recovery (i.e., annealing) (Fleischer et al., 1975), and the resulted shortened fission track length distribution is a function of heating time and temperature. In general, for fast-cooling basement sample (e.g., non-volcanic sample that cooled from subsurface condition with temperature higher than the PAZ), one would expect a dominant amount of tracks being 12-13 µm with a range of ~10-16 µm (Gleadow et al., 1986). Very short tracks of <10 µm are considered as a product at the high-temperature range of the PAZ (e.g., Tagami and O’Sullivan, 2005). Therefore the presence of short tracks is evidence that the sample has spent a period of time at the high-temperature range of the PAZ or has been subsequently heated to that temperature range. Annealing kinematics of apatite fission tracks have been studied with both laboratory and geological samples; models have been developed to solve for the apatite fission-track length distribution and the expected AFT ages for a given thermal history (e.g., Carlson et al., 1999; Donelick et al., 2005; Green et al., 1986; Ketcham, 2005). Conversely, one can also use the observed apatite fission track-length distribution and AFT ages to invert for plausible thermal histories, which contribute to geological interpretations using other geological lines of evidence.

Complications in interpreting thermochronological ages also lie in zircon and apatite (U-Th-Sm)/He thermochronology, as many factors other than cooling events may affect calculated ages, including grain geometry and aspect ratio (e.g., Beucher et al., 2013; Brown et al., 2013; Farley
et al., 1996; Reich et al., 2005; Wolf et al., 1996), U-rich grain-boundary phases (e.g., Murray et al., 2014), U and Th zoning (e.g., Farley et al., 2011), U-rich fluid inclusions or radiation damage (e.g., Flowers et al., 2009; Guenthner et al., 2013; Shuster et al., 2006).

Based upon studies of apatite, $^4$He diffusivity appears to decrease with increasing damage, because the radiation damage may result in preferential partitioning (“trapping”) of $^4$He in damage zones and impede diffusion of $^4$He (Flowers et al., 2009; Shuster et al., 2006). The amount of radiation damage is quantified by the effective uranium concentration (eU), calculated as $[\text{eU}]=[\text{U}]+0.235[\text{Th}]$ (Flowers et al., 2009). In apatite, radiation damage tends to manifest as a positive correlation of AHe ages and their respective eU concentrations for a given suite of samples that experience a similar time-temperature (t-T) history, as described by the RDAAM Model (the radiation damage accumulation and annealing model) (Flowers et al., 2009). In a case reported by Fox and Shuster (2014) from the Grand Canyon, long-term accumulation of radiation damage in apatite may have resulted in significant retention of $^4$He in apatite, leading to an elevated effective closure temperature that is higher than that of the AFT system.

Studies show that radiation damage in zircons can be observed as both positive and negative correlations between ZHe ages and eU concentrations, depending on the dose of damage (Guenthner et al., 2013). Radiation damage may impede diffusion of $^4$He and lead to positive correlation between eU and ZHe ages when the radiation damage is low (quantified by the alphas doses of $1.2\times10^{16}$ to $1.4\times10^{18}$ α/g) (Guenthner et al., 2013). However, when the radiation damage is higher than $\sim2\times10^{18}$ α/g (empirically it may correspond to zircons with over hundreds of ppm of eU), $^4$He diffusivity increases dramatically due to interconnection and consequential shrinking of the effective diffusion domain size (Guenthner et al., 2013). Under such a condition, zircons are ‘leaky’ and ZHe ages from a same sample would display negative correlation with their perspective eU concentrations.

1.2 Aims and Scope

In this thesis, I present multidisciplinary studies from the southern Puna Plateau in northwestern Argentina where I document the provenance and deformation of two sedimentary basins, and the exhumation history of one major bedrock mountain range. I analyze my data with existing
documentation from the southern central Andes in order to reconstruct regional deformation and landscape evolution, and to explore the underlying dynamic processes.

The regional deformation history of the southern Puna Plateau is key to understanding orogenic models for the central Andes. There are several reasons for this. First, although the southern Puna Plateau is home to distributed basement-cored ranges that are bounded by deep-seated faults, Cretaceous deposits are absent in this region, implying that it may have not been affected by the Salta rift (e.g., Marquillas et al., 2005; Salfity and Marquillas, 1994). Therefore, it provides an opportunity to test different orogenic models for the central Andes independent from the inversion of pre-existing structures. Second, documented Cenozoic deformation, including that related to the late Oligocene exhumation of the Sierra de Calalaste in the southwestern Puna Plateau (Carrapa et al., 2005) and late Miocene basin deformation of the Pasto Ventura region on the southeastern Puna Plateau (Zhou and Schoenbohm, 2015), cannot readily be correlated to the north, as has been recently proposed for the central Andes foreland system (DeCelles et al., 2011). In this regard, the southern Puna Plateau has lacked any geological data or measurements by which to test such models. Third, the southern Puna Plateau grades into the Sierras Pampeanas to the south, which is argued to reflect low-angle subduction-driven basement uplift. A good knowledge of the southern Puna Plateau is key to understanding how these two tectonic domains are genetically connected. This study explores ages and provenances of sedimentary strata preserved throughout the plateau region to reconstruct a regional deformation history, using sedimentary, petrological, structural, geochronological and thermochronological data.

The southern Puna Plateau has previously been the focus of pioneering studies regarding lower lithospheric foundering (Kay and Kay, 1993; Kay et al., 1994). It is now widely accepted that the lithosphere in this region has been thinned through lithospheric foundering during the late Cenozoic, as first proposed by Kay et al. (1994). However, the timing, location and mechanism for the lithospheric foundering event(s) are poorly understood. A recent hypothesis argues that the loss of lower lithosphere beneath the southern Puna Plateau may have taken place through multiple, small-scale (~50-100 km in diameter) drips (Ducea et al., 2013; Schoenbohm and Carrapa, 2015). This study provides a test for this hypothesis by constraining deformation of a sedimentary basin on the southernmost Puna Plateau and interpreting it with new and published geochronological, geochemical and geophysical data.
The geological history in the region of the present-day southern Puna Plateau reflects a complex interplay of extensional and contractional tectonism along a long-lived retro-arc region of an ocean-continent subduction system (e.g., Allmendinger 1983; Coira et al., 1982). There, bedrock mountain ranges that have witnessed intense deformation since the Paleozoic are suitable candidates for low-temperature thermochronological studies. On the one hand, thermochronological data can provide important constraints on tectonics and landscape evolution history for the study region. On the other hand, because of their complex history, they provide an opportunity to further advance and reveal the limits of low-temperature thermochronological data in landscape evolution studies. A multi-system thermochronological study is presented here for a major bedrock range on the southern Puna Plateau, using zircon and apatite (U-Th-Sm)/He and apatite fission-track thermochronometers. Thermal histories of samples are reconstructed with single-sample time-temperature modeling and explore the regional landscape evolution with landscape-thermochronology modeling.

1.3 Thesis Unfolded

This thesis has been written in manuscript style with four main chapters \((Chapters\ Two\ to\ Five)\), each of which is a single manuscript intended for a peer-reviewed journal. All four manuscripts are coauthored, but I am the first and corresponding author. In preparing each manuscript, I led the project design, conducted (most of the) field and laboratory work, directed data interpretation and wrote the text. This thesis also contains an introduction chapter \((Chapter\ One)\), which describes the overall goals and links among the main chapters, and a concluding chapter \((Chapter\ Six)\), which summarizes the major findings and proposes anticipated work.

Because of the chosen manuscript-style format, inevitable overlaps exist especially regarding sections describing the geological background and methodologies described within this thesis. Yet, each main chapter focuses one or two aspect(s) among many scientific questions regarding southern Puna Plateau and contains an independent set of data.

\textit{Chapter Two} explores the age and provenance of sedimentary strata preserved on the SE (southeastern) Puna Plateau. Sandstone modal composition and detrital zircon U-Pb and apatite fission-track data from Cenozoic strata indicate basin accumulation started during the late Eocene to early Oligocene (~38-28 Ma). With provenance analysis, this chapter proposes that the
southern Puna Plateau was occupied by a regional sedimentary basin during the late Eocene to early Oligocene (~38-28 Ma), and then was compartmentalized into small-scale intramontane basins by the growth of basement-cored ranges, leading to the formation of the modern plateau morphology by the late Miocene (~12-8 Ma).

Chapter Three elaborates on the regional deformation model proposed in Chapter Two and explores along-strike (~N-S) correlation with features and data from the Antofagasta de la Sierra region on the SW (southwestern) Puna Plateau. Sedimentary data, detrital zircon U-Pb geochronology and detrital apatite fission-track thermochronology data constrain the ages and provenance of these strata. Regional sedimentary correlation and statistical analysis of detrital zircon U-Pb data supports a regional basin covering the most of the southern Puna Plateau during the late Eocene to early Oligocene (~38-28 Ma). Apatite thermochronological data and models, and structural relationships suggest that the intervening bedrock ranges on the southwestern Puna Plateau began forming in the late Oligocene to early Miocene (~25-20 Ma), with increasing magmatic activity. Kinematic differences between the northern and southern Puna Plateau suggest that the central Andean foreland basin was initiated with the eastern side bounded by reactivated relict landscape, highlighting the importance of pre-existing crustal heterogeneities in influencing Cenozoic orogeny.

Chapter Four uses multiple thermochronometers, including zircon and apatite (U-Th-Sm)/He systems and apatite fission-track systems, to explore the low-temperature history of the Sierra Laguna Blanca, the major bedrock mountain range on the south Puna Plateau. Single-sample time-temperature modeling reveals that this range experienced a cooling event between the late Cretaceous to early Cenozoic (~90-50 Ma) times and a recent cooling event during the late Miocene (~20-10 Ma). An older cooling event is also present during the late Paleozoic (~300-250 Ma), but is less constrained by our data. The bedrock exhumation history defined in this chapter is supportive of the Cenozoic basin evolution as proposed in Chapter Two and Chapter Three, and also refines Paleozoic and Cretaceous paleogeographical models for the southern central Andes.

Chapter Five documents late Miocene (~12-8 Ma) basin formation and deformation in the southern Puna Plateau, using field mapping, deformation analysis and U-Pb and $^{40}$Ar/$^{39}$Ar geochronology. This chapter shows that major basin formation started ~11.7-10.5 Ma and
continued until at least ~7.8 Ma. The basin underwent syn-depositional faulting and folding during ~10-8 Ma. Contractional deformation in the Pasto Ventura basin ended between ~7.3 and 4 Ma, based on the onset of regional horizontal extension. This chapter highlights the role of the formation and detachment of a late Miocene lithospheric drip in shaping the upper crustal deformation on the southern Puna Plateau.

As of September 2015, Chapter Two is under revision in *Earth and Planetary Science Letters*; Chapter Three is under revision in *Geological Society of America Bulletin*; Chapter Four is under preparation for submission to *Tectonics*; Chapter Five has been published in *Lithosphere*. 
CHAPTER 2

SEDIMENTARY RECORD OF REGIONAL DEFORMATION AND DYNAMICS OF THE THICK-SKINNED SOUTHERN PUNA PLATEAU, CENTRAL ANDES (26-27°S)
2 Sedimentary record of regional deformation and dynamics of the thick-skinned southern Puna Plateau, central Andes (26-27° S)†

2.1 Abstract

The Puna Plateau, adjacent Eastern Cordillera and the Sierras Pampeanas of the central Andes are largely characterized by thick-skinned, basement-involved deformation. The Puna Plateau hosts ~N-S trending bedrock ranges bounded by deep-seated reverse faults and sedimentary basins. We contribute to the understanding of thick-skinned dynamics in the central Andes by constraining regional kinematics of the poorly understood southern Puna Plateau through a multidisciplinary approach. On the southeastern plateau, sandstone modal composition and detrital zircon U-Pb and apatite fission-track data from Cenozoic strata indicate basin accumulation during the late Eocene to early Oligocene (~38-28 Ma). Provenance analysis reveals the existence of a regional-scale foreland basin covering the southern Puna Plateau during late Eocene to early Oligocene time (~38-28 Ma) that was sourced from both the active orogenic front located on the western plateau and from re-activated, inherited topographic relicts located on the eastern plateau margin. Our petrographic and detrital zircon U-Pb data reveal erosion of proximal western and eastern sources after ~12 Ma, mid-late Miocene time. This indicates that the regional basin was compartmentalized into small-scale depocenters by the growth of basement-cored ranges continuing into the late Miocene (~12-8 Ma). We suggest that thick-skinned, distributed deformation in the southern Puna Plateau is the result of both an east-propagating foreland basin system and the reactivation of inherited structures.

† This chapter is under revision in Earth and Planetary Science Letters.

2.2 Introduction

Many of the world’s orogens are entirely or partially controlled by thick-skinned deformation: basement-cored mountain ranges, such as the Tien Shan, north African Atlas, southern Pyrenees, Laramide, northern and central Andes and Sierras Pampeanas, are confined by orogen-parallel, bivergent, deep-seated reverse faults (e.g., Carrera et al., 2006; Jordan et al., 1983; Macaulay et al., 2013; Nemcok et al., 2013). Their orogen-scale deformation histories have been used as indicators of underlying dynamic processes in models for thick-skinned tectonics. Most models, based on observation of regionally distributed deformation, envision the kinematics of thick-skinned orogens as primarily controlled by inversion tectonics, low-angle subduction and/or strong inter-plate coupling (e.g., Carrera et al., 2006; Hongn et al., 2007; Iaffa et al., 2013; Jordan et al., 1983; Kley et al., 1999; Nemcok et al., 2013).

The Puna Plateau of the central Andes has the morphology of a typical thick-skinned orogenic system, but its evolution history and dynamic processes remain enigmatic, especially its early Cenozoic to Miocene evolution. On the one hand, because this region inherited significant, deep-seated faults from Cretaceous rifting (Marquillas et al., 2005), Cenozoic mountain building triggered distributed deformation and range uplift, with early Cenozoic deformation notably taking place hundreds of kilometers inland of the plate boundary along reactivated rift structures (e.g., Carrera et al., 2006; Hongn et al., 2007; Kley et al., 1999). This phenomenon has led to the development of models favoring inversion tectonism and related “broken foreland” dynamics for the Puna Plateau and Eastern Cordillera (e.g., del Papa et al., 2013; Hain et al., 2011; Hongn et al., 2007; Sobel et al., 2003; Strecker et al., 2012). Studies have suggested that the modern Eastern Cordillera has experienced distributed range uplift deformation since the late Miocene, which has been argued to provide an analog to the earlier history for the interior of the Puna Plateau (e.g., Hain et al., 2011). However, other studies have argued that deformation and exhumation migrated from west to east within the thick-skinned Eastern Cordillera (e.g., Carrapa et al., 2011). Case studies in the Eastern Cordillera and western Altiplano-Puna Plateau, combined with regional compilations, argue for regional sequential deformation across the plateau region since the late Paleocene to middle Miocene, involving both thin- and thick-skinned domains (e.g., Carrapa et al., 2011; DeCelles et al., 2011, Horton, 2005; DeCelles and Carrapa, 2015). The central Andes developed along the Nazca subduction plate-boundary system
and may have experienced a number of orogenic cycles (the Cordillera cycle, DeCelles et al., 2015). Cyclicities in Cordilleran-type orogenic systems, featuring episodic high-flux magmatism in the arc and detachment of dense lower lithospheric roots (DeCelles et al., 2015, 2009; Paterson and Ducea, 2015), lead to pulsed advance of the orogenic front on the back-arc side of the orogenic system, analogous to a supercritical orogenic wedge (e.g., DeCelles et al., 2015, 2009; Reiners et al., 2015). The late Paleocene (~45-40 Ma) to middle Miocene (~20-12 Ma) in-sequence deformation in the Puna Plateau is in line with the predictions of this model (Carrapa et al., 2011; DeCelles et al., 2015, 2009). However, little is known about the regional deformation history and related dynamic processes in the southern Puna Plateau. The southern Puna Plateau is particularly important because it is a region that apparently did not experience significant Cretaceous rifting, evidenced by the lack of Cretaceous rift deposits (e.g., Marquillas et al., 2005), and thus has the potential to provide information on the dynamics of thick-skinned deformation independent of inversion tectonics (e.g., Carrapa et al., 2013).

We aim to explore the dynamics of the southern Puna Plateau and decipher its relationship with regional in-sequence or broadly distributed (out-of-sequence) deformation. We evaluate the onset of sedimentation and provenance change in the Pasto Ventura region, host to the only exposed sedimentary units in the southeastern portion of the plateau, through the use of sedimentary data, sandstone modal composition, and detrital zircon U-Pb and apatite fission-track data. We then reconstruct the paleolandscape and regional kinematics of the southern Puna Plateau and evaluate our results in the context of the larger scale dynamics of the thick-skinned orogenic development in the central Andes.

2.3 Background

The Altiplano-Puna Plateau (the Central Andean Plateau) is the dominant physiographic zone of the central Andes, with an average elevation of 3.9 km (e.g., Isacks, 1988) (Fig. 1A). It has relatively low relief, internal drainage, and is bounded by the higher, more rugged Eastern and Western Cordilleras. Volcanic rocks are widely distributed throughout the central Andes, mostly concentrated along the Western Cordillera and the western Altiplano-Puna Plateau, where the modern active volcanic arc is located. Well-documented Miocene-present volcanic rocks in the Puna Plateau range from large ignimbrite complexes to intermediate lavas to back-arc monogenetic mafic cinder cones (e.g., Kay and Coira, 2009). An important volcanic hiatus at
~38-27 Ma has been identified in the Puna Plateau during which virtually no volcanic activity took place (Kay and Coira, 2009; Trumbull et al., 2006). This lull has been attributed to low-angle subduction of the Nazca plate (e.g., Kay and Coira, 2009) or to the low-flux stage in the Cordillera cycle (e.g., DeCelles, et al., 2015; 2009).

The Altiplano Plateau and Eastern Cordillera of Bolivia largely developed through thin-skinned thrust-belt deformation that involves variably metamorphosed Paleozoic deposits but little basement rocks, along a gently westerly dipping regional basal detachment; the deformation front generally advanced to the east through time (e.g., Horton, 2005; McQuarrie et al., 2005; Uba et al., 2009). Sedimentary records suggest that a regional scale foreland basin system developed during Paleocene time and migrated 800 to 1000 km to the east, accompanied by concomitant eastward propagation of the orogenic front (e.g., DeCelles and Horton, 2003; Horton, 2005).

In contrast, the Puna Plateau and Eastern Cordillera of NW Argentina is a thick-skinned orogenic system, with basement rocks exposed through east and west-vergent, deep-seated reverse faults (e.g., Pearson et al., 2013). In the eastern Puna Plateau and adjacent Eastern Cordillera, the thick-skinned tectonic style is thought to be directly linked to the Salta rift that evolved through Cretaceous-Paleogene time (~150-60 Ma) in NW Argentina and resulted in accumulation of rift deposits covering the region of the present-day Santa Barbara system, parts of the Eastern Cordillera and the western margin of the Puna Plateau (Marquillas et al., 2005) (Fig. 2A). The Proterozoic-Paleozoic Puncoviscana Formation is widely distributed in the Puna Plateau mostly consist of the, yielding 450-650 and ~800-1350 Ma detrital zircon U-Pb ages (e.g., DeCelles et al., 2011). In the southern plateau interior, timing of topographic establishment of the major bedrock ranges and erosion is poorly known. The only documentation available is for the Sierra de Calalaste (Fig. 2A), which was exhumed during the late Oligocene (Carrapa et al., 2005).

Currently, the Puna Plateau is compartmentalized and hosts modern basins (e.g. salars) that are separated by basement-cored ranges. The earlier part of the sedimentary accumulation history, especially the early Cenozoic, is poorly constrained. Some consensus exists that the oldest Cenozoic sedimentary record on the Puna Plateau is represented by Eocene (~40 Ma) continental coarse facies described as the continental red beds, such as the Geste Formation in the northeastern Puna Plateau (e.g., Carrapa and DeCelles, 2008; Jordan and Alonso, 1987; Kraemer
et al., 1999). However, the distribution, age, and provenance of basin-fill strata across the plateau region, especially for the earliest Cenozoic sedimentary units, remain poorly documented. To the west in the southwestern Puna Plateau, the sedimentary record starts in the late Eocene, represented by the Quínoas Formation in the Salar de Antofalla region (e.g., Adelmann, 2001; Kraemer et al., 1999; Voss, 2002) (Fig. 2A). To the east, the Pasto Ventura region of NW Argentina hosts the only exposed Cenozoic sedimentary units in the southeastern Puna Plateau (Fig. 2A). The Pasto Ventura region has recently been mapped and the stratigraphic framework is now well-documented (Zhou and Schoenbohm, 2015) (Fig. 1B, 3A). Volcanic ashes are only present in the upper sections and are dated to 10.5 Ma to 7.9 Ma, providing good age constraints for the late Miocene sedimentary history for the southeastern Puna Plateau (Zhou and Schoenbohm, 2015). The exposed units also contain an older sedimentary unit that directly overlies the plateau basement, but its age was unknown prior to this study.

2.4 Methods and Results

2.4.1 Pasto Ventura sedimentary record

We studied the provenance history and the onset of sedimentation in the Pasto Ventura region, adopting the units defined by Zhou and Schoenbohm (2015) (Fig. 3A). Sedimentary facies in the central part of the basin, from youngest to oldest are typical of eolian (Ns-1), fluvial (Ns-2) and shallow lacustrine (Ns-3) environments, and these units contain abundant airfall and reworked ash horizons that range in age from 10.5 to 7.9 Ma (Zhou and Schoenbohm, 2015) (Fig. 3A; Fig. A1). Paleocurrent reconstruction from imbricated conglomerates and trough cross-stratification for units Ns-1 to -3 suggests that local paleo-flow directions vary but are generally eastward (Fig. 3A). Unit Ns-4 in the western part of the basin is deposited directly on basement rocks and is ash-free; it contains crude laminated and massive conglomerates, as well as massive sandstones and mudstones, with paleosol horizons and common carbonate nodules and burrows (Zhou and Schoenbohm, 2015). We sampled four medium grain sandstones, one from each unit, for detrital zircon U-Pb and detrital apatite fission-track analysis, and thirty-four sandstones from throughout the section for modal composition analysis.
2.4.2 Sandstone modal composition

We performed sandstone modal composition analysis on samples obtained throughout the sedimentary columns in the Pasto Ventura region. Thirty-four standard sandstone petrographic thin sections from units Ns-1 to -4 were stained for Ca- and K-feldspar and were point-counted (~450 counts per slide) according to the Gazzi-Dickinson method. We followed grain classes used in DeCelles et al. (2011) (Fig. 3; Table A2), and also calculated L’ and Lt’ by excluding volcanic lithic grains from L and Lt, respectively (Fig. 3; Table A3).

All samples from Ns-1 to -3 plot in the magmatic arc fields (dissected arc, transitional arc and undissected arc) in the QFL diagrams (Dickinson et al., 1983) (Fig. 3D) and contain abundant volcanic grains (on average 28% and up to 55% of total grains identified). In contrast, samples from Ns-4 contain few volcanic grains and plot away from volcanic arc zones (Fig. 3D, E). By excluding volcanic grains, which may not reflect erosion of proximal source areas but rather distal arc magmatism, and using Qt/L’ and Lm/Ls, we are able to minimize the influence of abundant ash fall derived grains in Ns-1 to -3 in our analysis. We find that Qt/L’ and Lm/Ls are higher in Ns-4 compared to the other units. In particular, the Lm/Ls (metamorphic lithic grains to sedimentary lithic grains) ratio is about an order of magnitude higher in Ns-4 than in Ns-1 to -3 (Fig. 3B, C), suggesting a relatively abundant contribution from sedimentary rocks in the source region during deposition of Ns-1 to -3.

2.4.3 Detrital zircon U-Pb geochronology

We dated ~100 detrital zircons grains from each of four samples, one from each sedimentary unit, by laser ablation ICP-MS at the University of Toronto (Fig. 4A, B). Zircon grains were randomly picked and mounted. We used cathodoluminescence and backscattered electron images to avoid cracks and target specific crystal domains when conducting laser ablation. Details are included in the data repository. Because $^{206}\text{Pb}^{207}\text{Pb}$ ages are most reliable for old zircons, we use $^{206}\text{Pb}^{238}\text{U}$ ages for analyses younger than 1000 Ma and use its $^{206}\text{Pb}^{207}\text{Pb}$ age if the $^{206}\text{Pb}^{238}\text{U}$ ages are older than 1000 Ma in histograms and probability density function plots (Fig. 4) (e.g., Gehrels, 2014).

Detrital zircon U-Pb age populations from Ns-1 to -3 (samples PVT53, PVT61 and PVT51) are similar to each other, characterized by overlapping age clusters at ~450-550, ~900-1350, and
2100-2000 Ma, with scattered ages from ~200-300 Ma and Cenozoic ages (Fig. 4A, B). The ~200-300 Ma grains mostly likely derived from distal western sources located in the Western Cordillera and the western Puna Plateau that host the Permian-Triassic intrusive rocks formed along the proto-Pacific margin (e.g., Breitkreuz and Zeil, 1994). The ~900-1300 Ma ages overlap with the age of the Precambrian Antofalla Terrane that comprises the basement beneath the vast part of the western-central Puna Plateau (Ramos, 2008). The Precambrian plateau basement experienced multiphase metamorphism during late Proterozoic to early Ordovician time, forming widespread metamorphic units such as the Proterozoic-lower Cambrian Puncoviscana Formation (e.g., Bahlburg and Breitkreuz, 1991; DeCelles et al., 2011; Ramos, 2008). The oldest detrital zircon U-Pb ages, ~1600-2000 Ma, may reflect detrital contribution from the eastern Puna Plateau where Paleoproterozoic bedrock outcrops (e.g., Bahlburg and Breitkreuz, 1991; DeCelles et al., 2011). Sample PVS1 from unit Ns-4 differs in that it lacks the ~200-300 Ma and ~1750-1900 Ma ages, and most notably, does not contain any Cenozoic grains (Fig. 4A). In units Ns-1, -2 and -3, Cenozoic ages have a wide range of 8-37 Ma, but are concentrated around ~9-13 Ma (Fig. 4B).

2.4.4 Detrital apatite fission-track thermochronology

We analyzed apatite grains for fission-track thermochronology at the Universität Potsdam using the external detector method (Gleadow, 1981; Hurford and Green, 1983). AFT ages and Chi-squared (χ^2) values were calculated using the Trackkey program (Dunkl, 2002), following the procedures of Galbraith (1981) and using the ζ calibration method (Hurford and Green, 1983) with a ζ value of 370.1±12.6 (R. Zhou).

We treated an aliquot of sample PVS1 with heavy ion irradiation at the Materials Research Department of the GSI Helmholtzzentrum (Darmstadt, Germany) in order to enhance yields of measurable confined tracks. We measured horizontal confined fission tracks in PVS1 from as many c-axis parallel grains as possible; only track-in-tracks were measured (Donelick et al., 2005). The length of apatite fission tracks decreases when the sample passes a geologically significant amount of time in the hotter portion of the apatite partial annealing zone (~140-120 to 60 °C) (e.g., Donelick et al., 2005). Typically, for the case of rapid initial cooling followed by a prolonged period staying close to the present-day temperature, the fission-track lengths are dominantly long (>~14 μm). For a bedrock sample that was fully reset before the latest cooling
event, the fission-track length distribution is indicative of its most recent cooling history. Yet, for a detrital AFT sample that was not thermally reset, the track-length distribution is a combination of track length distributions from all source regions.

We obtained over 100 ages each from samples PVS1 (Ns-4), PVT51 (Ns-3) and PVT61 (Ns-2). However, PVT53 (Ns-1) only yielded 20 dateable grains. We used the BinomFit program (Brandon, 2002) under the auto-mode to deconvolve the component ages for detrital samples (Table 2; Fig. 5). PVS1 from unit Ns-4 yields a large spread of AFT ages, from ~ 38-140 Ma, with four age populations calculated as $38.1^{+4.4}_{-4.0}$, $55.8^{+3.0}_{-2.8}$, $92.9^{+6.1}_{-5.7}$ and $139.5^{+9.8}_{-9.2}$ Ma (1σ) (Fig. 5D, Table 2). Samples from units Ns-1 to -3 all contain a population of ~16 Ma grains (Table 2), which most likely reflects input from volcanic fields on the southern Puna Plateau. They also yield early Cenozoic ages at $52.7^{+3.2}_{-3.0}$ and $64.7^{+3.0}_{-2.8}$ Ma (1σ) (PVT51 and PVT61) and early Cretaceous ages of $106.4^{+9.5}_{-8.7}$ and $110.4^{+5.4}_{-5.2}$ Ma (PVT61 and PVT53) (Table 2; Fig. 5). For sample PVS1 in unit Ns-4, 259 track-length measurements yield a mean track length of $11.67\pm1.74$ µm, with a wide range from 6 to 16 µm (Fig. 5F).

2.5 Discussion

2.5.1 Onset of sedimentation in the southeastern Puna Plateau

The depositional age of the earliest Cenozoic sedimentary unit across the Puna Plateau is a critical element in testing plateau development models, especially during the early Cenozoic. Unit Ns-4 was deposited directly on bedrock and therefore represents a proxy for the minimum age of uplift and exhumation of the source region, which must have preceded deposition. Several lines of evidence suggest that, although it is not directly dated, unit Ns-4 is late Eocene to early Oligocene in age (~38-28 Ma). First, sandstone modal composition, detrital zircon U-Pb data and detrital apatite fission-track data suggest that the unit Ns-4 was deposited during a regional volcanic lull. Cenozoic zircons and volcanic lithic grains are absent from unit Ns-4, and the unit contains no ashes. Further, apatite fission-track analysis for sample PVS1 from unit Ns-4 did not reveal distinguishing euhedral shapes from the youngest, Cenozoic population, supporting that those grains were not likely derived from air-fall volcanics. Although volcanic activities along the central Andes vary spatiotemporally, the southern Puna Plateau and the adjacent Western and Eastern Cordilleras have been volcanically active since late Oligocene, resulting in widely
distributed volcanic rocks (Kay and Coira, 2009; Trumbull et al., 2006), implying that Ns-4 must at least predate the late Oligocene (~28 Ma). Second, given the lack of volcanic input, detrital apatites in Ns-4 must have derived from eroded source terrains. AFT ages therefore reflect exhumation of source terrains or burial heating or a combination of both. We rule out significant post-depositional heating and hence AFT age reduction based on the wide dispersion of single-grain AFT ages and preservation of long tracks (> ~14 µm) in sample PVS1 (Fig. 5F). The shortened mean track lengths we do observe most likely reflect inheritance from the sediment source region; several bedrock ranges on the present-day Puna Plateau bear apatites with shortened track lengths because of slow exhumation through the partial annealing zone (e.g., Carrapa et al., 2013; Zhou et al., 2014). Therefore, the youngest AFT population from Ns-4, 38.1 +4.1/-4.0 Ma (1 σ) (Fig. 5D), must be older than deposition of PVS1. In short, we argue that the age of the oldest sedimentary unit (Ns-4) from the Pasto Ventura region, and therefore the onset of sedimentation in the southeastern Puna Plateau, dates to late Eocene to early Oligocene with an age range of ~38-28 Ma.

2.5.2 Late Eocene to early Oligocene (~38-28 Ma) basin initiation

The existence of upper Eocene to lower Oligocene strata (Ns-4) in Pasto Ventura is interesting in a regional context. To the west in the southwestern Puna Plateau, the first sedimentary units above the basement, the Quiñoas Formation in the Salar de Antofalla region (Fig. 2A), are also dated as late Eocene to early Oligocene (e.g., Adelmann, 2001; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002). Exhumation of the two major intervening bedrock ranges, the Sierra de Calalaste and the Sierra Laguna Blanca (Fig. 2A), is argued to have occurred later, during the late Oligocene (~29-24 Ma) (Carrapa et al., 2005) and late Miocene (~15 Ma) (Zhou et al., 2014), respectively. Therefore both ranges were likely submerged below sedimentary strata during the late Eocene to early Oligocene, leading us to argue that the southern Puna Plateau was occupied by a large, uninterrupted, regional sedimentary basin at this time (Fig. 2C).

This regional sedimentary basin was sourced from both the east and the west (Fig. 2C). The eastern source was located approximately at the present-day southeastern margin of the Puna Plateau (Fig. 2A, C), where the Sierra de Chango Real yields apatites with 38-29 Ma AFT ages (Coutand et al., 2001). Similar ages were obtained from unit Ns-4 in the Pasto Ventura region (the P1 age component, Fig. 5D), supporting this inference. Additionally, although unit Ns-4
does not contain paleocurrent indicators, it contains layers of coarse-facies sediments (Fig. 3A), suggesting the existence of a relatively proximal source. To the west, this regional sedimentary basin received sediments from western bedrock mountain ranges such as the Sierra Quebrada Honda (Fig. 2A, C). Paleocurrents and conglomerate composition data demonstrate that the Sierra Quebrada Honda was the primary source for the upper Eocene to lower Oligocene sedimentary rocks in the Salar de Antofalla region (e.g., Adelmann, 2001; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002) (Fig. 2A). The lack of intervening ranges between the Pasto Ventura region and the Salar de Antofalla region (Fig. 2A), as suggested by our paleographic reconstruction, implies that sediments derived from this western source would also have reached the Pasto Ventura region.

Because unit Ns-4 is relatively thin (<0.5 km) and contains paleosol horizons, burrows and carbonate nodules indicating prolonged subaerial exposure, it is not likely to represent the basin depocenter. We suggest that the depocenter was instead located in the present-day Salar de Antofalla and Antofagasta de la Sierra regions to the west (Fig. 1B, 2A), where thick (1-3 km) late Eocene and early Oligocene sedimentary units are preserved (e.g. Adelmann, 2011; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002) (Fig. 2C).

### 2.5.3 Late Miocene (~12-8 Ma) basin rejuvenation

There is no recorded deposition between Ns-4 (~38-28 Ma, this study) and 11.7-7.8 Ma strata (Ns-1, -2 and -3) in the Pasto Ventura region (Zhou and Schoenbohm, 2015), suggesting a phase of non-deposition and/or erosion before the mid-Miocene (Fig. 3A).

Permian-Triassic detrital zircon U-Pb ages (~200-300 Ma) are absent in Ns-4, but are present in Ns-3 to -1, arguing for post-mid-Miocene input from the western plateau and Western Cordillera, which hosts same-aged intrusive rocks related to the proto-Pacific margin (e.g., Breitkreuz and Zeil, 1994). Lower Qt/L’ and Qm/Lt’ for Ns-2 and -3 suggest that the source region was composed of relatively more sedimentary and/or low-medium-grade metamorphic rocks compared to unit Ns-4 (Fig. 3B, C). Lower Lm/Ls ratios in Ns-1 to -3 compared to Ns-4 also imply a significant source of sedimentary rocks (Fig. 3B, C). Collectively, we interpret these signatures to be the result of recycling of exhumed sedimentary units in the Antofagasta de la Sierra region, east of the Sierra de Calalaste (Fig. 1B, 2A). The exhumation of this section took...
place as early as the late Oligocene (~29-24 Ma), when rapid cooling of the Sierra de Calalaste began (Carrapa et al., 2005). The presence of 1600-2000 Ma zircon U-Pb ages in Ns-1, -2, and -3 leads us to also consider a detrital contribution from the eastern plateau where Paleoproterozoic bedrock outcrops (e.g., Bahlburg and Breitkreuz, 1991; DeCelles et al., 2011). Study of a vertical transect from the Sierra Laguna Blanca documents AFT ages of 45-65 Ma (Zhou et al., 2014), which are consistent with age populations obtained from PVT61 in unit Ns-2 (52.7 $^{+3.2/-3.0}$ Ma, 1 $\sigma$) and PVT51 in unit Ns-3 (64.7 $^{+4.4/-4.1}$ Ma, 1 $\sigma$) (Fig. 5). Therefore, we argue that post-mid Miocene sedimentation on the southeastern Puna Plateau took place in a small-scale depocenter bounded by topographically high, thrust or reverse fault-bounded ranges to the northwest and northeast (Fig. 2B). The bounding thrusts were likely active during basin formation, consistent with documented mid-late Miocene syn-depositional deformation in the region (Schoenbohm and Carrapa, 2015; Zhou and Schoenbohm, 2015).

2.5.4 Dynamic models for the thick-skinned southern Puna Plateau

When testing dynamic models for the Puna Plateau, previous studies have focused on discriminating between eastward younging (in-sequence) and widely distributed (out-of-sequence) deformation across the Puna Plateau and the adjacent Eastern Cordillera (e.g., Carrapa et al., 2011; DeCelles et al., 2011; del Papa et al., 2013; Hain et al., 2011; Hongn et al., 2007). The central argument is that regional in-sequence deformation and the formation of a regional, ~N-S-trending foreland basin supports supercritical wedge-like dynamics (e.g., Carrapa et al., 2011; DeCelles et al., 2011); widely distributed deformation and the formation of isolated depocenters, on the other hand, supports dynamic models such as inversion tectonics or broken foreland dynamics, which reflect the influence of pre-existing crustal heterogeneities (e.g., del Papa et al., 2013; Hain et al., 2011; Hongn et al., 2007), or are a result of the subcritical state of the central Andean orogenic wedge (e.g., Pearson et al., 2013).

We propose that the late Eocene to early Oligocene development of the southern Puna Plateau represent the formation of a regional, uninterrupted basin, supporting the model of a wedge front located in the western plateau (V in Fig. 6). Cenozoic shortening simultaneously caused the reactivation of an older relict landscape >200 km east of the deformation front, separated by a vast region apparently lacking deformation (Fig. 2C). Our argument for this revised, hybrid orogenic model for the southern Puna Plateau is as follows.
We argue that the regional late Eocene-early Oligocene basin in the southern Puna Plateau (Fig. 2C) represents the southern continuation of the central Andean foreland basin system (DeCelles et al., 2011). First, as discussed above, the depocenter of this basin was located to the west and the late Eocene-early Oligocene strata were thicker to the west and thinner to the east (Fig. 2C), documenting a W-E asymmetry consistent with a foreland basin geometry (e.g., DeCelles and Giles, 1996; Flemings and Jordan, 1989). Second, in the Pasto Ventura region, unit Ns-4 yields detrital AFT ages of ~35-120 Ma, but the eastern source we propose for this basin (the Sierra de Chango Real; Coutand et al., 2001, Fig. 1B) does not account for the full range of detrital AFT ages observed in this unit. Therefore, its western source, such as the Sierra Quebrada Honda, must have yielded the ~35-120 Ma detritalapatites (AFT ages). Indeed, such ages are nearly identical to detritalapatites dated in the Geste Formation in the northern Puna Plateau (north of ~25-26°S). Structural, sedimentary and geo-/thermochronological data suggest that the Geste Formation was derived from proximal sources that were associated with upper-crustal deformation (Carrapa and DeCelles, 2008). Deposition of the Geste Formation suggests that the deformation front was active along the eastern margin of the Puna Plateau during late Eocene time (Carrapa and DeCelles, 2008) (IV in Fig. 6). We argue that the deformation front south of ~25-26°S in the Puna Plateau was located in the Sierra Quebrada Honda region during the late Eocene-early Oligocene, marking a mapview deflected orogenic front for the southern central Andes (Fig. 6). The orogenic front served as the primary source and also as a crustal load, leading to flexural subsidence and the formation of a foreland basin system that is traceable from the northern Puna Plateau and the Altiplano Plateau, encompassing both thin- and thick-skinned domains (e.g., Carrapa et al., 2011; DeCelles et al., 2011) (Fig. 6).

At the same time, the late Eocene-early Oligocene basin was also bounded and received sediments from an eastern source (Fig. 2C). Apatites from the Sierra de Chango Real, hundreds of kilometers east of the orogenic front, have relatively long (~13-14 µm) mean fission-track lengths (Coutand et al., 2001), suggesting active exhumation during late Eocene time (Fig. 2C). Although it has not been previously thought to lie within the Cretaceous Salta rift (Marquillas et al., 2005), the Sierra de Chango Real is close to a range that was exhumed in the Cretaceous (La Quebrada, Carrapa et al., 2013), suggesting it could have formed a local topographic high, possibly as a rift shoulder, prior to the formation of the late Eocene-early Oligocene basin. Thus, we argue that the eastern sedimentary source for this regional basin was a result of reactivation
of relict topography and/or faults by Cenozoic shortening, similar to what is proposed for the northern part of the eastern flank of the Puna Plateau and the adjacent Eastern Cordillera, where direct field evidence is preserved (e.g., del Papa et al., 2013; Hongn et al., 2007; Pearson et al., 2013). Interestingly, our analysis highlights the lack of evidence for late Eocene to early Oligocene deformation between the western and eastern sedimentary sources (Fig. 2C), suggesting deformation was likely confined to the vicinity of the Salta rift, underlining the role of pre-existing structural discontinuities in allowing deformation to occur far inboard of the wedge-related deformation front.

The existence of this eastern source may have influenced the formation of the regional basin, highlighting that both lithospheric flexure related to orogenic wedge dynamics and the reactivation of preexisting structures have played an important role in the evolution of the southern Puna Plateau. The oldest, ~38-28 Ma strata in the Pasto Ventura region (unit Ns-4) are thin (<0.5 km) and contain paleosol horizons, burrows and carbonate nodules indicating prolonged subaerial exposure, typical of a forebulge setting. However, this unit also contains coarse facies that are not typical of a forebulge (DeCelles and Giles, 1996). These conflicting observations suggest the presence in this location of an ephemeral forebulge within a system that was mostly overfilled due to the activation of the eastern source region. Alternatively, the late Eocene-early Oligocene basin could have been overfilled by spill-over from the distal foredeep and/or because of the high flexural rigidity which might be expected during the volcanic lull.

Since late Oligocene time the regional basin was compartmentalized by internal deformation and bedrock-cored range uplift (e.g., Carrapa et al., 2005; Zhou and Schoenbohm, 2015; Zhou et al., 2014; this study), leading to the formation of a landscape similar to the present-day broken foreland in the Eastern Cordillera. Activation of intervening bedrock ranges represents an important shift of dynamic processes in the southern Puna Plateau from large-scale lithospheric flexure to local, small-scale dynamic factors. First, because the orogenic front had migrated to the Eastern Cordillera by the Miocene (e.g., Carrapa et al., 2011), distributed deformation within the southern Puna Plateau after the late Oligocene may reflect wedge-top dynamics. Alternatively, the upper-crustal deformation could be controlled by formation of a lithospheric drip. For example the late Miocene basin rejuvenation in the Pasto Ventura region has been argued to be a result of a small-scale lithospheric foundering event beneath the southern Puna Plateau (Schoenbohm and Carrapa, 2015; Zhou and Schoenbohm, 2015). Third, once the thick-
skinned compressional basin-and-range system was established, the oscillating basin infill and excavation caused by shifts of orographic precipitation may have controlled the uplift of individual bedrock ranges (e.g., Sobel et al., 2003). Furthermore, we find that the formation of a regional foreland basin took place during the volcanic lull, while the start of bedrock-cored range uplift coincided with the recent episode of volcanic activity, implying a genetic link between upper-crustal evolution and arc activity as envisioned in the Cordillera-type orogenic cyclicity (DeCelles et al., 2015, 2009).

Although many of the world’s thick-skinned orogens are dominated by distributed deformation localized by reactivation of pre-existing structures or basement anisotropies (e.g., Carrera et al., 2006; Iaffa et al., 2013; Jordan et al., 1983; Macaulay et al., 2013; Nemcok et al., 2013), our results support the view that the southern Puna Plateau (26-27°S) belongs to the central Andean foreland basin system, which propagated eastward through time as a Cordilleran-type orogenic dynamic system and encompassed both thin-skinned and thick-skinned structural domains (e.g., DeCelles et al., 2015). A broken foreland-style landscape, which may be promoted by reactivation of inherited crustal heterogeneities, such as those inherited from the Salta rift, has evolved in the southern Puna Plateau as early as the late Oligocene, following the formation of a regional basin.

2.6 Conclusion

We conclude that the development of the southern Puna Plateau was achieved by the formation of a regional-scale sedimentary basin during late Eocene to early Oligocene time, which was then compartmentalized by the growth of ~N-S-trending bedrock ranges starting as early as the late Oligocene (~29-24 Ma) (Carrapa et al., 2005) until after the late Miocene (~12-8 Ma) (this study). The early, late Eocene to early Oligocene (~38-28 Ma) regional basin was sourced from both an active orogenic front to the west and reactivated topographic relicts to the east. The southern Puna Plateau can be viewed as part of the central Andes orogenic system, formed by a wave of east-propagating deformation and a regional foreland basin since Paleogene time. Since the late Oligocene, the southern Puna Plateau was characterized by distributed range uplift, similar to the late Neogene broken foreland in the Eastern Cordillera. We find that the regional deformation history in the southern Puna Plateau supports both plate-boundary-scale Cordillera-type orogenic dynamics and regional-scale inversion tectonics.
### TABLE 1. RESULTS FOR DETRITAL APATITE FISSION-TRACK THERMOCHRONOLOGY

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<th>NI</th>
<th>$pD^\dagger$</th>
<th>ND#</th>
<th>$P (\chi^2)$ (%)</th>
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<th>±1σ</th>
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*pS and $pI$ stand for density of spontaneous tracks and density of induced tracks, respectively.
†NS and NI stand for number of actual counted spontaneous tracks and induced tracks, respectively.
§$pD$ stands for density of induced tracks to the CN5 dosimetry glass.
ND stands for the number of actual counted tracks for determining $pD$.
**We report central ages for all samples (bold italics) because they all fail the $\chi^2$ test.
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Figure 1. (A) Overview of the central Andes. Blue lines encompass the internal drainage area (plateau region). Red lines are 3-km contours. (B) Geological map for the southern Puna Plateau (modified from Zhou and Schoenbohm, 2015). Detailed logged sections are included in Data Repository.
Figure 2. (A) Map of the southern Puna Plateau (after Allmendinger et al., 1989; Martinez, 1995; Schnurr et al., 2006) indicating major geological units and thrust faults. (B) and (C) Proposed schematic cross-sections of the southern Puna Plateau at 26-27 °S.
Figure 3. (A) Generalized stratigraphic column for the Pasto Ventura region from measured sedimentary sections (Fig. A1). Ages for volcanic ashes (*Zhou and Schoenbohm, 2015; **Schoenbohm and Carrapa, 2015), and detrital zircon and apatite samples are indicated. Arrows in circles indicate paleocurrent reconstructions. (B, C) Selected ratios for sandstone modal composition parameters. (D, E) Ternary diagrams illustrating recalculated modal petrographic data from Pasto Ventura region. Provenance fields are from (Dickinson et al., 1983).
Figure 4. (A) Detrital zircon U-Pb ages for Pasto Ventura samples. (B) Histogram for Cenozoic detrital zircon U-Pb ages.
Figure 5. (A to D) Radial plots for detrital AFT ages. (E) Lag-time plots for detrital AFT age populations. AFT age populations are listed in Table 2. (F) Track-length distribution for PVS1.
Figure 6. Regional compilation of the locations of the deformation front through time (Arriagada et al., 2006; Carrapa and DeCelles, 2008; Carrapa et al., 2009; DeCelles and Horton, 2003; McQuarrie, 2002; McQuarrie et al., 2005). Black lines outline area above 3 km in elevation. Blue lines outline internally-drained region, and indicate extent of the Altiplano-Puna Plateau.
CHAPTER 3

LATE EOCENE TO LATE MIOCENE (~38-8 MA) BASIN EVOLUTION ON THE SOUTHERN PUNA PLATEAU: NEW CONSTRAINTS ON OROGENIC MODELS IN THE CENTRAL ANDES
3 Late Eocene to late Miocene (~38-8 Ma) basin evolution on the southern Puna Plateau: new constraints on orogenic models in the central Andes‡

3.1 Abstract

Sedimentary units within an orogenic system provide important constraints in reconstructing the regional deformation history and understanding underlying dynamic processes. On the southwestern Puna Plateau, sandstone modal composition, detrital zircon U-Pb ages and modeled bedrock apatite fission-track data suggests that the accumulation of sediments took place in (~38-28 Ma) in the Antofagasta de la Sierra region. Provenance data suggest a western source that probably lies in the Sierra Quebrada Honda region. Apatite fission-track and (U-Th-Sm)/He data suggest that the exhumation of the intervening range, the Sierra de Calalaste between the Sierra Quebrada Honda region and the Antofagasta de la Sierra region, occurred later during the late Oligocene to early Miocene (~25-20 Ma). In the southeastern Puna Plateau, we find that the major bedrock range, Sierra Laguna Blanca, was not present during the late Eocene to early Oligocene based on newly-documented and published detrital and bedrock zircon U-Pb data. Taken together, this study supports the existence of a regional basin covering the southern Puna Plateau, including (from west to east) the Salar de Antofalla, Antofagasta de la Sierra, Sierra Laguna Blanca and Pasto Ventura regions during the late Eocene to early Oligocene (~38-28 Ma). Starting in the late Oligocene (~29-24 Ma), the regional basin was compartmentalized by uplift of intervening bedrock ranges, leading to the formation of the modern, compressional basin-and-range morphology by no later than the late Miocene (~12-8 Ma). The southern Puna Plateau was incorporated into the regional east-propagating central Andes foreland basin starting as early as late Eocene, driven by lithospheric flexure and orogenic wedge dynamics. We

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observe a mapview deflection of the orogenic front in the southern Puna Plateau and suggest a potential influence of the pre-existing structures in modulating Cenozoic orogenic propagation in the Puna Plateau.

3.2 Introduction

The Altiplano-Puna Plateau is an important example of orogenic processes along a non-collisional plate boundary. Current models, conceived from observations at various locations and scales, suggest a great degree of diversity in geodynamic processes within the central Andes. In the Puna Plateau and Eastern Cordillera of the NW Argentina, pre-existing crustal heterogeneities, such as deep-seated faults inherited from Cretaceous rifts, were reactivated during Cenozoic shortening, leading to distributed range uplift and formation of isolated depocenters (e.g., Hongn et al., 2007; Sobel et al., 2003; Strecker et al., 2012). Other models emphasize along-strike, ~N-S correlation of geologic data throughout the central Andes. Periodic build-up and foundering of dense roots beneath the arc, as evident from patterns in volcanism and plutonism (e.g., DeCelles et al., 2009; Paterson and Ducea, 2015), may lead to alternation in the orogenic system between subcritical and supercritical states (DeCelles, et al., 2015b). During the supercritical state, the back-arc deformation front advances to the east (cratonic side), leading to in-sequence deformation that can be ~N-S correlated (e.g., Carrapa et al., 2011; McQuarrie et al., 2005; Uba et al., 2009) and the formation of a regional ~N-S-striking foreland basin due to lithospheric flexure in front of the orogenic load (e.g., DeCelles et al., 2011; Horton, 2005), contrasting with out-of-sequence deformation and isolated depocenters predicted in the broken foreland model.

The southern Puna Plateau (~26-27°S) is poorly understood, but is key to deciphering orogenic growth and dynamic processes in the central Andes. The southern Puna Plateau hosts distributed basement-cored ranges that are bounded by deep-seated faults, similar to the Eastern Cordillera, where inversion tectonics has been invoked (e.g., del Papa et al., 2013; Hongn et al., 2007). However, the thick rift-related Cretaceous deposits in the Eastern Cordillera are absent in the southern Puna Plateau (Marquillas et al., 2005; Salfity and Marquillas, 1994), suggesting that the Salta Rift did not extend into the plateau region. Therefore the origin of any pre-existing weaknesses is enigmatic and the importance of inversion tectonics is uncertain.
The southern Puna Plateau is also a region that is immediately to the south of the proposed, ~N-S-trending foreland basin system (DeCelles et al., 2011), so it might be expected that deposition and deformation correlate. However, current knowledge regarding sedimentation and deformation for the southern Puna Plateau is still limited, but existing document, including the late Miocene deformation in the Pasto Ventura region (Zhou and Schoenbohm, 2015) and late Oligocene exhumation of the Sierra de Calalaste (Carrapa et al., 2005), seems to indicate out-of-sequence deformation, inconsistent with the ~N-S-trending regional foreland basin.

Resolving the viability of these models is important for understanding foreland basin systems in regions of complex inherited geology in the southern Puna Plateau and around the world. We provide new insights into those intertwining orogenic processes by reconstructing regional deformation using multidisciplinary approaches. This study explores genetic links among widely distributed but poorly exposed sedimentary units and intervening bedrock ranges on the plateau, combining new and existing sedimentary, provenance, structural and geo-/thermochronological data. The analysis enables us to explore the growth and dynamics of the southern Puna and to gain insights into the orogenic dynamics of the central Andes. We find that the regional deformation of the Puna Plateau may be explained by a combination of eastward foreland basin propagation and inversion tectonics. The different types of pre-existing structures in the cratonic side may lead to different amounts of Cenozoic shortening, exerting a control on the propagation of the Cenozoic orogenic front.

### 3.3 Background

The Altiplano-Puna Plateau of the central Andes is the world’s second largest continental plateau after the Tibetan Plateau. It is the dominant physiographic zone of the central Andes, internally drained and with an average elevation of over 3 km (Allmendinger et al., 1997; Isacks, 1988; Strecker et al., 2007) (Fig. 1). It is divided at ~22.5°S latitude into the Altiplano Plateau to the north and the Puna Plateau to the south, and is bounded by higher, more rugged Western and Eastern Cordilleras (Fig. 1).

Volcanic rocks are widely distributed throughout the central Andes, mostly concentrated along the Western Cordillera and the western Altiplano-Puna Plateau, where the modern active volcanic arc is located. Volcanic rocks on the Puna range from large ignimbrite complexes to
intermediate lavas to back-arc monogenetic mafic cinder cones (e.g., Kay and Coira, 2009). An important volcanic hiatus at ~38-27 Ma has been identified for the Puna Plateau during which virtually no volcanic activity took place (Trumbull et al., 2006; Guzmán et al., 2014; Kay and Coira, 2009). This lull has been attributed to low-angle subduction of the Nazca plate (e.g., Trumbull et al., 2006; Guzmán et al., 2014; Kay and Coira, 2009) or to a low-flux stage in the Cordillera cycle (DeCelles et al., 2009; DeCelles et al., 2015).

Estimates of paleoaltimetry from paleobotanic and stable isotope-based proxies ($\delta^{18}$O, $\delta$D and clumped isotope $\Delta$47) suggest that, although the timing of surface uplift of the Altiplano Plateau and its adjacent Eastern Cordillera varies, the major surface-elevation gain occurred during the Miocene (Gregory-Wodzicki, 2000; Garzione et al., 2008, 2014; Leier et al., 2013). Such estimates are supported by sedimentary and geomorphic constraints along the plateau margins (e.g., Alván et al., 2015; Fox et al., 2015; Schildgen et al., 2010). In contrast, in the Puna Plateau, stable isotope-based paleoaltimetric proxies, such as $\delta^{18}$O and $\Delta$47 from pedogenic carbonate nodules and $\delta$D from devitrified volcanic glass, suggest that the western part of the plateau and the Western Cordillera were high, similar to present-day elevations, by ~34 Ma (Canavan et al., 2014; Quade et al., 2015), and the eastern Puna Plateau and its adjacent Eastern Cordillera achieved their modern elevation at ~21-14 Ma (Canavan et al., 2014; Carrapa et al., 2014; Quade et al., 2015). Thermochronological and geomorphologic studies have demonstrated the existence of high topography along the southern margin of the Puna by middle-late Miocene (~18-9 Ma) (Carrapa et al., 2006; Montero-López et al., 2014).

The present-day morphology of the Puna Plateau and its adjacent Eastern Cordillera and Sierras Pampeanas is characterized by compressional basin-and-range morphology, with sedimentary basins or modern salt flats (salars) bounded by ~N-S-oriented bedrock mountain ranges composed mainly of Paleozoic metasedimentary and intrusive rock and Precambrian basement (e.g., Schnurr et al., 2006) (Fig. 1). In the Eastern Cordillera and the eastern Puna Plateau, this tectonic style is thought to be directly linked to the Salta rift that evolved through Cretaceous-Paleogene time (~150-60 Ma) in NW Argentina and left rift deposits covering the present-day Santa Barbara system, parts of Eastern Cordillera and the western margin of the Puna Plateau (Marquillas et al., 2005; Salfity and Marquillas, 1994) (Fig. 1).
Outcrops of Cenozoic sedimentary rocks are limited, especially in the southern Puna Plateau where they are largely obscured by young volcanic and Quaternary colluvial materials. The oldest Cenozoic sedimentary record on the Puna Plateau is Eocene (~40 Ma) in age and consists of continental coarse facies described as ‘continental red beds’ (e.g., Carrapa and DeCelles, 2008; Jordan and Alonso, 1987; Kraemer et al., 1999). Consensus exists that since at least the Miocene (~24 Ma), possibly the late Oligocene (~29-24 Ma), sedimentation on the Puna took place in separated intramontane basins (Adelmann, 2001; Alonso et al., 1991; Carrapa et al., 2005; Jordan and Alonso, 1987; Kraemer et al., 1999; Vandervoort et al., 1995; Voss, 2002).

Deformation and exhumation events documented throughout the Puna Plateau and its adjacent Eastern Cordillera (see Barnes and Ehlers, 2009, and Zhou and Schoenbohm, 2015, for summaries), have led to the development of two classes of plateau-scale kinematic models. The first class emphasize that the thick-skinned structural style in the Puna is strikingly similar to that of both the Eastern Cordillera to the east and the Sierras Pampeanas to the south, and argues for distributed basement-cored range uplift across the region (e.g., Hongn et al., 2007; Jordan and Allmendinger, 1986; Sobel et al., 2003; Strecker et al., 2012). Particularly, the well documented Neogene Eastern Cordillera, a “broken foreland,” is argued to be an analog for the Puna Plateau, where the shortening drives distributed, deep-seated reverse thrust faulting that preferentially reactivates pre-existing rift structures and may occur irregularly throughout the region, rather than progressing across the region over time (e.g., del Papa et al., 2013; Hain et al., 2011; Hongn et al., 2007; Pingel et al., 2014; Sobel et al., 2003; Strecker et al., 2012). The alternative class of model emphasizes along-strike correlations, in particular the similar ages of the earliest Cenozoic deformation within the Altiplano-Puna Plateau, and proposes a ~N-S striking foreland system that occupied the majority of the central Andes with a deformation front that migrated eastward since the Paleogene (e.g., Carrapa et al., 2011; DeCelles et al., 2011). This model suggests that the central Andes behaved as an orogenic wedge, with the deformation front propagating eastward during a supercritical stage in the Cenozoic driving the formation of a regional, flexural basin (DeCelles et al., 2009; DeCelles et al., 2015).
3.4 Methods

3.4.1 Study sites

We present new sedimentary, structural, detrital zircon geochronological and detrital and bedrock apatite thermochemical data from the Antofagasta de la Sierra region (the ANT region, \( \sim 26^\circ29'\text{S}; \sim 67^\circ56'\text{W} \)). Here, thick sedimentary units are exposed immediately to the east of the Sierra de Calalaste (the Calalaste range), a major \( \sim \text{N-S oriented bedrock mountain range} \) in the southern Puna Plateau.

In addition, we present new bedrock zircon U-Pb geochronological data from the Sierra Laguna Blanca (the LB range), a \( \sim \text{N-S bedrock mountain range located in the southeastern Puna Plateau} \) that is composed of Precambrian basement rocks, and further interpret detrital zircon U-Pb geochronological data from the Pasto Ventura region (the PV region) that is adjacent to the LB range and hosts sedimentary strata as old as late Eocene (Zhou et al., in revision; Chapter 2).

3.4.2 Geological mapping and sedimentary logging in the Antofagasta de la Sierra region

In the ANT region, exposed sedimentary units are mapped using an ASTER (Advanced Space-borne Thermal Emission and Reflection Radiometer) satellite image base through both field and remote mapping, aided by aerial photographs (Instituto Geografico Militar, Argentina), satellite images and digital elevation models (DEMs) including Landsat, Google Earth and SRTM (Shuttle Radar Topography Mission) data (Fig. 2A). We supplement the western part of our mapping with existing geological maps (e.g., Schnurr et al., 2006). Three sedimentary sections were logged in the field and imbricated pebbles were measured as paleocurrent indicators (Fig. 3).

3.4.3 Sandstone modal composition

Sixteen standard sandstone petrographic thin sections were prepared and stained from the ANT sections for Ca- and K-feldspar and performed point-counting (\( \sim 450 \) counts per slide) according to the Gazzi-Dickinson method (Ingersoll et al., 1984) (Fig. 4). We follow the grain classification used by DeCelles et al. (2011) (Data Repository) and plot the results in the Dickinson diagrams (Dickinson and Suczek, 1979; Dickinson et al., 1983).
3.4.4 Zircon U-Pb geochronology

Eight detrital samples from the ANT region and four bedrock samples from the LB range were analyzed with zircon U-Pb geochronology by laser ablation ICP-MS (a VG Series 2 Plasmaquad ICP-MS and a 213-nm New Wave laser system) at the Jack Satterly Geochronology Laboratory at University of Toronto. Zircon grains were mounted in epoxy and polished, then imaged with cathodoluminescence (CL) and backscattered electrons (BSE) using a JEOL JSM6610-Lv scanning electron microscope. CL and BSE images were used to avoid cracks and target specific crystal domains when conducting laser ablation. Details are included in the Data Repository.

We focus on analyses that yield less than ±15% discordance, but also report analyses with discordance ranging from ±15-40% (Fig. 6). Because $^{206}\text{Pb}/^{207}\text{Pb}$ ages are most reliable for old zircons, we use $^{206}\text{Pb}/^{238}\text{U}$ age for an analysis when younger than 1000 Ma and use its $^{206}\text{Pb}/^{207}\text{Pb}$ age if the $^{206}\text{Pb}/^{238}\text{U}$ age is older than 1000 Ma in histograms and probability density function plots (e.g., Gehrels, 2014).

3.4.5 Apatite fission-track thermochronology

Apatite fission-track (AFT) thermochronology were performed on six detrital and two bedrock samples from the ANT region at the Universität Potsdam using the external detector method (Table 1) (Gleadow, 1981; Hurford and Green, 1983). Apatite grains were mounted on glass slides with epoxy, ground, polished and etched with 21°C, 5.5 N HNO₃ for 20 seconds (Carlson et al., 1999; Donelick et al., 2005). We then attached a low-U mica sheet to each sample and irradiated them in the Oregon State University Radiation Center together with CN5 dosimeters. Following irradiation, the mica external detectors were etched with 21°C, 40% hydrofluoric acid (HF) for 45 minutes. We also treated an aliquot of apatite grains from each bedrock sample (ABD2 and ABD22) with heavy ion irradiation at the Materials Research Department of the GSI Helmholtzzentrum (Darmstadt, Germany), in order to enhance yields of measurable confined tracks (Jonckheere et al., 2007). Heavy ion-treated slides were etched under the same condition as the regular grain-mount slides.

C-axis parallel apatite grains were analyzed with reflected and transmitted light at 1250x magnification under a Leica DMRM microscope with drawing tube located above a digitizing tablet and a Kinetek computer-controlled stage driven by the FTStage program (Dumitru, 1994).
AFT ages and Chi-squared ($\chi^2$) values were calculated using the Trackkey program (Dunkl, 2002) following the procedures of Galbraith (1981). We used the $\zeta$ calibration method (Hurford and Green, 1983) with a $\zeta$ value of 370.1±12.6 (R. Zhou). We measured horizontal confined fission tracks from as many c-axis parallel grains as possible; only track-in-tracks were measured (Donelick et al., 2005). $D_{par}$ values (the etch figure length parallel to c-axis, Donelick et al., 2005) were used to parameterize the kinetic properties for grains that were either counted or containing track lengths. We averaged at least four measured $D_{par}$ from each grain and report corrected values following Sobel and Seward (2010) using a factor of 0.88 (R. Zhou) (Fig. 7). We used the program BinomFit (Brandon, 2002) under the auto-mode to deconvolve the component ages for detrital samples (Fig. 7).

3.4.6 Single-grain apatite (U-Th-Sm)/He thermochronology

Apatite grains from two detrital samples (ADR5 and ADR6) and one bedrock sample (ABD22) were analyzed with apatite (U-Th-Sm)/He thermochronology (AHe) (Table 2).

Clear apatite grains without apparent inclusions and other impurities were selected and packed individually in platinum tubes under a binocular microscope. We analyzed four to seven single-grain aliquots from each sample and used grain dimensions and number of terminations to calculate the $F_T$ correction factor (Farley et al., 1996). Sample-containing tubes were loaded into a 25-spot laser chamber of an ASI Alphachron He extraction and analysis system at Universität Potsdam, equipped with a 30W Coherent 978 nm diode laser (FAP-98-30C-800-B) and a Pfeiffer Prisma 200 Quadrupole mass spectrometer. Blank tubes and age standards (Durango apatite) were routinely run together with samples. Samples were heated by the laser system at 8 amps (~3.5W) for 5 minutes to release all He from the apatite crystals. The released gas was purified by exposure for 1 minute to a hot getter (SAES AP10N), designed to remove chemically active gas species. The amounts of $^4$He in the purified gas were determined by isotope dilution using a $^3$He tracer, calibrated against a manometrically determined $^4$He standard. Each sample was re-extracted and analyzed a second time to make sure that the grain was degassed entirely in the first step. In this study, average $^4$He blanks are ca. 0.76 fmol (0.017 ncc) and sample-to-blank ratios are averaged to 19. Sample-specific sample-to-blank ratios averaged over all aliquots were 1, 23 and 27 for samples ADR6, ADR5 and ABD22, respectively.
After degassing, the samples were recovered from the laser chamber, transferred to a clean lab at GFZ Potsdam, and prepared for analysis of U, Th, and Sm by isotope dilution. Samples and their Pt wraps were placed in 3 ml Savillex PFA screw-cap vials, spiked with a HNO₃-based ²³⁵U-²³⁰Th spike and a HNO₃-based ¹⁴⁹Sm spike, and dissolved with ~0.5 ml 7N HNO₃. The spikes are calibrated against NIST-traceable, Certified Reference Material ICP concentration standards. To ensure complete sample dissolution and isotopic homogenization between sample and spikes, the vials were placed for at least 24 h on a hotplate at ~100°C. No significant amounts of Pt are dissolved during this process. The solution was then evaporated to dryness and re-dissolved for another 24 h in 1.5 ml 2% HNO₃. The solution was then analyzed for U, Th and Sm isotopic composition on a Thermo Element 2 XR ICP-MS instrument at GFZ Potsdam, equipped with a CETAC ASX-520 auto-sampler system, and run in low-resolution mode to maximize transmission of ions. Beside ²³⁸U, ²³⁵U, ²³²Th, ²³⁰Th, ¹⁴⁷Sm and ¹⁴⁹Sm, we also analyzed mass 234, which is used to detect potential Pt-Ar isobaric interferences on the U mass spectrum. Such interferences were generally found to be negligible. Instrumental mass fractionation was monitored by repeated analysis of ¹⁴⁹Sm/¹⁴⁷Sm ratios of naturally occurring Sm, and of the NIST SRM material U-500. Total procedural blanks were <0.005 pmol for ²³²Th, <0.0006 pmol for ¹⁴⁷Sm, and <0.003 pmol for ²³⁸U. Due to total blank levels commonly considerably lower than the above values and the high variability of total blank levels, no useful blank correction can be applied to the analytical data. Sample results with potentially significant contributions of the procedural blank on the overall U, Th, Sm abundances were discarded. Reproducibility (2SD) of standard solution concentration data is in the range of 0.4% for U, 1.3% for Th, and 0.6% for Sm.

Ages are calculated following Meesters and Dunai (2005) using U, Th, Sm abundances, blank-corrected He abundances, the Fₜ correction factor and the alpha-particle stopping distance described by Ketcham et al. (2011). The mean age determined for 66 Durango apatite aliquots in the Potsdam labs (excluding outliers) is 30.82±0.54 Ma (1SD), with a standard error (2SE) of ±0.21 Ma and reproducibility (2SD weighted error/mean age) of ±3.5%. This mean age is in good agreement with the ⁴⁰Ar/³⁹Ar reference age (31.44±0.18 Ma) and related (U-Th-Sm)/He ages for the Durango apatite (mean of 31.02±1.01 Ma, 1σ; McDowell et al., 2005). For samples, we report a weighted error which weights the uncertainty of the isotopic abundance by the relative contribution to the total helium production and also includes the uncertainty on the
blank-corrected measured $^4$He. We report concentrations based on measured abundances and a mass calculated from grain dimensions converted to an equivalent spherical radius (ESR) and an assumed apatite density of 3.15 gm/cc (Table 2).

### 3.4.7 Inverse modeling of apatite thermochronological data

We performed time-temperature (t-T) inverse modelling with AFT and AHe data from two bedrock samples (ABD2 and ABD22) using the HeFTy program (Ketcham, 2005). AFT and AHe data were modeled using the Ketcham (2007) annealing model and the RDAAM model (the radiation damage accumulation and annealing model, Flowers et al., 2009), respectively.

Because single-grain AFT ages from both samples failed the $\chi_2$ test, we modeled a subset of each dataset, grouped based on Dpar values (Ketcham et al., 2007). All AFT lengths were modeled with c-axis projected length values (Ketcham et al., 2007) and the 252Cf-irradiation option was used.

Models were started at 180-140 °C, temperatures high enough to fully reset the AFT system, and finished at 10±10 °C at t=0 Ma, without any t-T constraints imposed (Fig. 9). We divided the possible t-T range into two or three t-T boxes in order to facilitate model search in as much t-T space as possible (Fig. 9). All t-T histories are presented with acceptable and good fits of the input thermochronological data. From a statistical point of view, the so-called best-fit model does not necessarily represent the most plausible t-T history, but rather a means to check how well the models reproduce the observed thermochronological data.

### 3.5 Antofagasta de la Sierra Region

#### 3.5.1 Sedimentary units and sandstone modal composition

We logged three sections: S3 (oldest, 552 m), S2 (1263 m) and S1 (1062 m, youngest). 500-meter separation is estimated between the top of S2 and the base of S1 from dip of the strata and the measured surface distance from satellite images. Therefore the ANT region hosts at least a ~3.4 km sedimentary succession, which we divide into five units based on their sedimentary facies and distinct colors on satellite images.

N-5 is the lowest member and is deposited directly onto bedrock. It is characterized by fine-grained sandstone and mudstones. It contains structure-less, well sorted, 10-20-cm thick
conglomerates and massive, laminated and large-scale cross bedded sandstones. N-4 is dominantly composed of thick (> 50 cm) conglomerate layers that are matrix-supported and structure-less. N-3 contains alternating mud-siltstone to medium-coarse sandstone layers typically 3-5 meters thick. The sandstone layers lack cross-beddings and are dominantly massive. Discontinuous conglomerate (~2-5 cm) strips are occasionally present in sandstone layers. N-2 is an alluvial fan complex with widespread cross-bedded conglomerate and sandstone, with alluvial channel structures. Units N-5 to N-2 are conformable. Paleocurrent indicators from N-5 to N-2 provide overwhelming easterly flow directions (Fig. 3). In contrast, paleocurrents from N-1 are westerly-directed (Fig 2B; Fig. 3). We infer a regional angular unconformity between the N-1 and underlying N-2 to N-5, based on different paleo-flow directions and contrasting colors from satellite images. The top of N-1 is overthrust by the major basin-bounding fault (Fig. 2) and only several tens of meters of N-1 are preserved. N-1 is mainly composed of trough cross-bedded conglomerate and medium sandstone. It also contains massive conglomerate and massive medium sandstone layers, with thin (less than 50 cm) mudstone that contains pedogenic carbonate nodules.

Sandstone samples from the ANT sections show little compositional variation, lack volcanic lithic grains and cluster within the mixed zone among the recycled orogenic, cratonic interior and basement uplift sources in the Dickinson diagrams (Fig. 4). The sandstone compositions are similar to those of the Quiñoas Formation from the Salar de Antofalla region to the west (Adelmann, 2001; Carrapa et al., 2005) and to unit Ns-4 in the PV region to the east (Zhou et al., in revision; Chapter 2).

Since the late Oligocene, the Puna Plateau and its Western and Eastern Cordilleras have been volcanically active, as recorded in widely distributed volcanic rocks (Kay and Coira, 2009; Kay et al., 2010). This signature manifests itself in the sedimentary record through an increasing volcanic input signal in sandstone modal compositions (Fig. 4) or preserved air-fall volcanic materials, such as volcanic ash layers, in younger sedimentary strata. Therefore, we argue that the studied sedimentary sections in the ANT region are no younger than late Oligocene based on the lack of volcanic input, and are time-equivalent with the upper Eocene to lower Oligocene Quiñoas Formation in the Salar de Antofalla region to the west based on their similarities in lithologic characteristics.
3.5.2 Basin inversion and basement-involved deformation

Deformation in the ANT region is dominantly controlled by a series of ~N-S-trending structures and is thick-skinned (Fig. 2A). The study region is bound to the west by a major W-dipping, basin-bounding reverse fault, which carries bedrock (Puncoviscana formation) over the sedimentary section. The fault trace on the surface is clearly evident by a distinct color change in satellite images and therefore it enables us to measure a ~50-60° dip based on mapped fault trace and topography (Fig. 2A, lowest stereonet). Field measurements of associated minor faults suggest that this fault curves locally but overall strikes ~N-S strike dips relatively steeply (Fig. 2A). A major N-S-striking syncline lies in the footwall of the thrust, affecting the majority of the mapped strata. The syncline has a sub-vertical axial plane, is ~S-plunging and is truncated by the basin-bounding reverse fault (Fig. 2A). To the north, the reverse fault extends into bedrock and the major syncline dies out, replaced by a basement-cored anticline in the immediate footwall of the fault (Fig. 2A). Further east, sedimentary units have been folded into a ~N-S-striking, open anticline-syncline pair, segmented by local reverse faults that strike NW-SE (Fig. 2A). Additionally, Quaternary terraces in the north are displaced by several ~N-S-striking normal faults which are related to recent plateau-scale extension (Dortch et al., in prep.) and are not within the scope of this study.

3.5.3 Magmatic intrusion into sedimentary strata

The sedimentary strata in the ANT region were intruded by mafic sills after deposition (Fig. 5). The intrusions are usually parallel to but sometimes cut at a low angle through sedimentary beds, and are traceable for hundreds of meters on the surface (Fig. 5). Therefore it is plausible to infer that intrusive bodies are sheet-like with tapered fronts and widespread beneath the surface. The outcropping sills are typically 2-6 meters thick and fine-grained with chilled margins (Fig. 5K). Thin-section observations reveal a mafic composition and diabasic texture, with fine-grained, needle-like plagioclase crystals and visible olivine grains in a highly degraded matrix. This post-depositional magmatic intrusion would have heated the sedimentary basin and perturbed the geothermal field in three dimensions, particularly if aided by basin-fluid circulation.

3.5.4 Detrital zircon U-Pb geochronology

Eight detrital samples yielded similar zircon U-Pb ages, despite being sampled across the entire ~3.4 km thick section, suggesting rapid deposition or existence of stable source regions. Three
major age groups are observed: ~200-300 Ma, 450-650 Ma and ~900-1300 Ma (Fig. 5). The ~900-1300 Ma ages overlap with the age of the Precambrian Antofalla Terrane that is basement to the vast part of the western-central Puna Plateau (Ramos, 2008, 2010). The Precambrian plateau basement experienced multiphase metamorphism during late Proterozoic to early Ordovician time, producing widespread metamorphic units such as the Proterozoic-lower Cambrian Puncoviscana Formation (e.g., Aceñolaza et al., 1988). The observed 450-650 Ma ages from our detrital samples are consistent with this distributed metamorphic bedrock source (Fig. 12). Core-rim structures on CL images of zircons are supportive of this interpretation, which shows ~900-1300 Ma core ages and 450-650 Ma rim ages (Fig. 12). As for the ~200-300 Ma zircons, CL images reveal distinct oscillatory zoning, typical of magmatic zircon (Fig. 12) (Corfu et al., 2003), suggesting a source in the Permian-Triassic plutonic and volcanic rocks in coastal Chile on the western margin of the Puna Plateau (e.g., Lopez-Gamundi and Breitkreuz, 1997; Breitkreuz and Zeil, 1994).

Detrital samples also yield a limited number of Cenozoic ages; we found four grains from three samples that are 38-39 Ma (Fig. 4). Although we do not rely on these four dates to determine the age of the studied sections, they may represent a maximum deposition age and are consistent with our upper Eocene to lower Oligocene age inference for the ANT sections.

### 3.5.5 Apatite thermochronology for bedrock samples

Sample ABD2 was sampled in the hanging wall of the W-dipping, basin-bounding reverse fault (Fig. 2A). It yields an AFT central age of 88.8±7.6 Ma with significantly shortened track length (MTL, mean track length: 9.72±2.22 μm). The track-length histogram contains a single peak around 10 μm (Fig. 8), suggesting that the sample has experienced partial resetting or spent a prolonged period of time within the partial anneal zone (PAZ) prior to final exhumation (e.g., Armstrong, 2005; Gleadow et al., 1986). Our HeFTy model further delineates its possible thermal history. As evidenced by the oldest track at 138 Ma, AFT data for ABD2 provide constraints for its thermal history since the Cretaceous. The sample cooled below ~80 ºC as early as ~130 Ma and stayed at this relative low temperature until the late Eocene (~35-30 Ma) (Fig. 9A). The sample may have experienced an increase of ~10-20 ºC, reaching ~90 ºC during late Eocene to late Oligocene (~35-25 Ma) and cooled to the surface temperature starting at ~25-20 Ma (Fig. 9A).
Sample ABD22 was collected from the hinge zone of the basement-cored anticline in the north of the ANT region (Fig. 2). It yields an AFT central age of 19.9±1.5 Ma with relatively long track length (MTL: 12.63±2.38 µm) (Fig. 8). The track-length histogram has a peak around 14-16 µm but notably also distributed short, ~4-12-µm tracks, implying relatively fast exhumation after a protracted period spent in the partial anneal zone (PAZ) (e.g., Armstrong, 2005; Gleadow et al., 1986). We analyzed seven apatite grains from the ABD22 sample with AHe thermochronology. Because the AHe ages cluster despite of a wide range of eU, we argue that the sample experienced rapid cooling from a temperature higher than the PRZ (partial retention zone) and calculate a mean AHe age of 19.9±2.0 Ma (Fig. 8). We modeled AFT and AHe data of this sample with HeFTy (Ketcham, 2005). The AFT data do not provide constraints for the sample’s thermal history before formation of the oldest track at 30-Ma; therefore randomly-distributed t-T paths occupy the t-T space before ~30 Ma. Nevertheless, the model suggests monotonic cooling towards the surface since ~25-20 Ma (Fig. 9B).

The t-T histories of samples ABD2 and ABD22 indicate that the basement rocks in the ANT region experienced a rapid cooling from at least ~80 °C at ~25-20 Ma. The dissimilarities between the two samples imply that they were located at different structural depths prior to the most recent phase of exhumation, with ABD22 likely being lower (higher temperature). In both cases, the samples may have remained at a subsurface temperature of ~80 °C around 25 Ma (Fig. 9), which is consistent with a late Eocene to early Oligocene (~38-28 Ma) deposition in the ANT region.

3.5.6 Apatite thermochronology for detrital samples

Most detrital grains yield AFT ages from ~20 to 200 Ma (Fig. 7). However, two detrital samples (ADR5 and ADR6) from the middle-lower part of the total sedimentary column yield abnormally young ages at ~13-14 Ma that are younger than samples at lower and higher stratigraphic positions (Fig. 7). The single-grain AFT ages within the same samples are statistically consistent and result in similar pooled ages of 14.2±1.0 Ma and 13.6±1.2 Ma in ADR5 and ADR6 respectively. The mean track lengths from ADR5 and ADR6 are 13.41±1.89 µm and 13.01±2.70 µm respectively; these are longer than from the rest of the detrital samples, which are ~10-11 µm (Fig. 7). These observations suggest that the ANT units experienced a post-depositional thermal event(s) that was able to significantly, if not totally, reset the AFT system in the middle-lower
part of the section. This would require a rise of temperature to at least higher than \(~100-110 \, ^\circ\text{C}\), the high-temperature boundary of the PAZ. We argue that the intruded sills served as the required heat source and draw caution for influence of magmatic activities in arc basins in detrital thermochronological studies.

AHe data from ADR5 and ADR6 support our inference of a post-depositional thermal event(s). ADR6 yields four consistent AHe ages averaged to 5.9±2.1 Ma, with low eU concentrations in the narrow range of 24 to 47 ppm (Table 2; Fig. 10). For ADR5, four of five analyzed grains yielded clustered ages ranging from 16.7 to 23.7 Ma, with high eU concentrations of 111 to 277 ppm, while the remaining grain yielded an older age of 49.4±1.3 Ma with notably high eU concentration of 419 ppm (Table 2; Fig. 10). In ADR6, AFT and AHe ages may represent one or two post-depositional thermal events. One possibility is that a \(~6 \, \text{Ma}\) thermal event fully reset the AHe system but only partially reset the AFT system. Alternatively, the \(~14 \, \text{Ma}\) AFT age and \(~6 \, \text{Ma}\) AHe age could be the result of two distinct thermal events, with the younger event only hot enough to reset the AHe system. In ADR5, the AFT age (14.2±1.0 Ma) is younger than any AHe ages (Table 2; Fig. 10) and all measured apatite grains in AHe analysis have eU concentrations that are one order of magnitude higher than average apatites. Very high eU concentrations and a crude correlation between eU and AHe ages (Fig. 10) lead us to consider that the AHe ages in ADR5 are mostly likely a result of incomplete annealing of radiation damage (e.g., Flowers et al., 2009).

Because the effect of post-depositional thermal event(s) on the other samples higher or lower in the section is not readily to be constrained, a further interpretation for the detrital AFT population, such as a classic lag-time analysis (e.g., Coutand et al., 2006), is not grounded. Indeed, the lack of long (\(~14 \, \mu \text{m}\)) track lengths in the rest of the detrital AFT samples supports the interpretation that they have been affected by post-depositional thermal event(s). The two samples likely least affected by post-depositional thermal-resetting are ADR1 and ADR2, based on their larger dispersal of AFT ages (Fig. 7). The youngest AFT populations from ADR1 and ADR2 are 27.9±4.7 Ma and 32.9±2.9 Ma, identical within error (Fig. 7). They are in line with the inferred age for the studied sedimentary sections because the youngest detrital AFT population should be no younger than the depositional age of the host rocks.
3.6 Sierra Laguna Blanca and Pasto Ventura region

3.6.1 Bedrock zircon U-Pb geochronology for the Sierra Laguna Blanca

U-Pb ages for zircon from the bedrock samples of the LB range yield ages ranging from ~450 to over 2000 Ma. We calculated concordia ages (Ludwig, 1998) from the youngest data clusters for all samples, yielding ages of 510.4±4.1 Ma (LB5369), 517.1±3.3 Ma (LB4992), 510.7±5.2 Ma (LB4621) and 474.4±4.9 Ma (LB4025) (Fig. 13), indicating the timing for the partial melting events. Scattered ages as old as 2000 Ma indicate that rocks of the LB range are a result of partial melting of Precambrian basement (Fig. 13; Data Repository).

3.6.2 Interpretation of existing detrital zircon U-Pb ages for the Pasto Ventura region

Detrital zircon U-Pb ages from the sedimentary units in the PV region were reported by Zhou et al. (in revision; Chapter 2). The oldest Cenozoic sedimentary unit is deposited directly above bedrock and is estimated to be upper Eocene to lower Oligocene (~38-28 Ma) in age (Zhou et al., in revision) (Fig. 10). After an inferred sedimentary hiatus, sedimentary units were deposited from 11.7-7.8 Ma (Zhou and Schoenbohm, 2015). We further interpret these data by incorporating them into the K-S test (the Kolmogorov-Smirnov test; e.g., Berry et al., 2001; DeGraaff-Surpless et al., 2003; Guynn and Gehrels, 2010; Press et al., 1986) with data from the ANT region and the LB range (Fig. 14). The K-S test functions to reject, rather than to realize, a null hypothesis that the two distributions are the same or came from the same parent population (e.g., Guynn and Gehrels, 2010). Therefore two samples that fail the K-S test were likely not derived from the same source, but passing the test does not prove correlation.

Because the four bedrock samples were taken from the bottom to the top from the LB range, well-representing the spectrum of derived detrital zircons, we are able to use their ages in the K-S test to test the LB range as a source terrane for sedimentary strata in the ANT and PV regions. The K-S test rejects the possibility of the LB range as the source for the lowest sample from the PV region and samples N-2 to N-5 from the ANT region, all of which were deposited during the late Eocene to early Oligocene (Zhou et al., in revision; Chapter 2; this study). This suggests that the LB range was not exposed at the surface at least prior to early Oligocene.
3.7 Development of the regional landscape of the southern Puna Plateau

3.7.1 Late Eocene to early Oligocene: formation of the regional basin

We document thick (>3.4 km) sedimentary units in the ANT region that mostly accumulated during the upper Eocene to lower Oligocene, with eastward oriented paleocurrent indicators (for N-2 to Ns-5). They were likely continuous with the Quiñoas Formation to the west of the Calalaste range, the oldest known Cenozoic unit in the Salar de Antofalla region. This formation also has an upper Eocene to lower Oligocene age, constrained by radiometric ages of volcanic tuffs (Adelmann, 2001; Kraemer et al., 1999) and also shows eastward oriented paleocurrent data (Adelmann, 2001; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002). The two intervening bedrock ranges were exhumed after this time; 24-29 Ma for the Calalaste range (Carrapa et al., 2005) and ~22-20 Ma for the bedrock in the hanging wall of the thrust immediately west of the ANT region (this study). The ANT section was also likely continuous with the oldest unit in the PV region, which is of similar age (Zhou et al., in revision; Chapter 2).

We therefore propose the existence of a regional-scale sedimentary basin on the southern Puna Plateau (~26-27° S) during the late Eocene to early Oligocene that covered the present-day Salar de Antofalla region, the Calalaste range, the ANT region, the LB range and the PV region, extending at least ~150 km from west to east. The maximum thickness for the Quiñoas Formation to the west is ~1.4 km, consisting of playa mudflat, alluvial and sheet flood and distal braided river deposits (e.g., Adelmann, 2001). To the east in the PV region, only ~400 meters of the lowest unit is exposed, but it contains coarse cross-bedded red sandstone and siltstone, pedogenic carbonate nodules and paleosol horizons, indicating slow accumulation and prolonged subaerial exposure in the late Eocene to early Oligocene, with possible periods of erosion or at least sedimentary hiatuses (Zhou and Schoenbohm, 2015; Zhou et al., in revision; Chapter 2).

We suggest that this early southern Puna basin was W-E asymmetric, with its depocenter located in the ANT region where the thickest (>3.4 km) upper Eocene to lower Oligocene units are found.

Provenance studies support this model. The Quiñoas Formation was sourced from the western, adjacent Sierra de la Quebrada Honda (Adelmann, 2001; Kraemer et al., 1999; Voss, 2002), which is composed of Ordovician low-grade metamorphic rocks in greenschist-facies (e.g.,
Alonso et al., 1984; Zimmermann et al., 1996). In the ANT region, our data from detrital zircon U-Pb geochronology, sandstone modal composition and paleocurrents also point to a source region in the Sierra de la Quebrada Honda. In the PV region, Zhou et al. (in revision; Chapter 2) data show that the late Eocene to early Oligocene unit had both western and eastern sedimentary sources. The western source is consistent with the Sierra de la Quebrada Honda and the eastern source is inferred to be the present SE plateau margin, including the Sierra Chango Real (Zhou et al., in revision; Chapter 2).

In short, during late Eocene to early Oligocene time, the southern Puna Plateau (~26-27º S) was occupied by one regional basin extending from the western Salar de Antofalla to the present-day southeastern plateau margin. This basin was bounded by higher topography to the west (e.g., Sierra de la Quebrada Honda) and the east (e.g., Sierra Chango Real), was E-W asymmetric and had a depocenter located in the ANT region.

### 3.7.2 Late Oligocene to late Miocene: basin compartmentalization and magmatic invasion

Since late Oligocene to late Miocene, the early southern Puna basin was compartmentalized by uplift of distributed ~N-S bedrock ranges, forming small-scale basins, some of which continued to accumulate sediment while others experienced sediment-starvation. At the same time, they were subject to magmatic intrusion as the volcanic arc became increasingly active.

On the southwestern Puna Plateau, in the Salar de la Antofalla region, the Quiñoas Formation is overlain by the Chacras Formation above a regional angular unconformity; a tuff close to the base of the Chacras Formation yields an age of 24.3±0.9 Ma (Adelmann, 2001). Paleocurrents from the Chacras Formation are variable but show some clusters of westward directions (Adelmann, 2001; Carrapa et al., 2005). Carrapa et al. (2005) documents the exhumation for the Calalaste range at 24-29 Ma and argues that the Chacras Formation received sediments from the uplifted Calalaste range based on paleocurrent and petrographic data. Since then, the Salar de la Antofalla region has been internally drained and has accumulated sediments more or less continuously to the present-day (Adelmann, 2001; Carrapa et al., 2005; Kraemer et al., 1999; Voss, 2002). We find that unit N-1 in the ANT region was deposited above an inferred angular unconformity and has westward paleocurrents as well, suggesting its probable equivalence with the Chacras Formation. The major basin-bounding thrust therefore is required to be active no
earlier than the deposition of the Chacras Formation, consistent with recorded ~25-20 Ma bedrock exhumation ages from this study (Fig. 7). The wide range of 29-20 Ma onset of exhumation documented from across the Calalaste range (Carrapa et al., 2005 and this study) implies that the uplift may have been accommodated through a series of range-parallel faulting imbrications. Alternatively, the 29-20 Ma onset of exhumation may reflect multiphase reverse faulting and sub-range uplift within the Calalaste range.

On the southeastern Puna Plateau, in the Pasto Ventura region (the PV region), the sedimentary record is missing from late Oligocene (~28 Ma) to 11.7 Ma, suggesting sediment-starvation or erosion (Zhou and Schoenbohm, 2015; Zhou et al., in revision). Sedimentation resumed in the PV region after 11.7 Ma and continued until shortly after 7.9 Ma, accompanied by syndepositional deformation (Zhou and Schoenbohm, 2015). Thermochronological data suggest that the LB range northwest of the PV region experienced its most recent exhumation in the middle Miocene (Zhou et al., 2014). Therefore, the southeastern part of the Puna Plateau, represented by the PV region and the LB range, was actively shortening during the late Miocene. Similarly, bedrock along the southern margin of the Puna Plateau underwent rapid exhumation during ~25-15 Ma (Carrapa et al., 2006).

Because the central Andes hosts an active volcanic arc, sedimentary basins located within it are prone to magmatic intrusion, as demonstrated in this study. We find that two detrital samples in the lower part of the section have abnormally young AFT and AHe ages which may be explained by heating at least to ~110 ºC during ~18-6 Ma (Fig. 9). This age is younger than the exhumation event found in bedrock samples from the same region and the lowest detrital sample (ADR21) shows less thermal imprint, leading us to exclude burial heating as a cause. Although they do not outcrop close within tens of meters of our sampled sections, sills may be widely distributed and have a complicated subsurface geometry, affecting a broad region. We draw awareness of post-depositional magmatic intrusion in arc basins for future detrital thermochronological studies in similar settings.

3.8 Growth and Dynamic Models for the Puna Plateau

Based on our findings for the southern Puna Plateau (~26-27° S) of a wide Eocene to early Oligocene basin with a westerly-located depocenter and post-late Oligocene
compartamentalization, we explore different plateau growth and dynamic models and along-strike similarities and dissimilarities in the central Andes.

In the northern Puna Plateau (~24.5-25.5º S), the sedimentary strata record a regional foreland basin during late Eocene to early Oligocene time, with an elevated, deforming wedge front and wedge-top depozone in the eastern Puna, and a foredeep depozone in the western Eastern Cordillera (e.g., Carrapa and DeCelles, 2008; Canavan et al., 2014; DeCelles et al., 2007, 2011; Quade et al., 2015), although this interpretation is controversial (e.g., del Papa et al., 2013; Hongn et al., 2007). Recent isotopic dating indicates that strata previously thought to be Eocene in the Salar de Arizaro region in fact belong to the lower-middle Miocene Vizcachera Formation (DeCelles et al., 2015a), highlighting that the Eocene strata in the northern Puna are concentrated in the eastern plateau margin (the Geste Formation of the Pasto Grande region) and the adjacent Eastern Cordillera (the Quebrada de los Colorados Formation of the Angostaco region) (Fig. 11). Provenance, sedimentary, structural and geo-/thermochronological studies suggest that the Geste Formation was derived from the bedrock nearby and associated with upper-crustal shortening, suggesting a wedge-top origin (e.g., Carrapa and DeCelles, 2008; DeCelles et al., 2007). In this context, the Quebrada de los Colorados Formation is interpreted to be located in the foredeep depozone, receiving sediments from the elevated region to the west (e.g., DeCelles et al., 2011) (Fig. 11). Stable isotope-based paleoaltimetric estimates from the Geste Formation and the Salar de Arizaro region, which represent active wedge front or top-wedge depozone, yield elevation close to present-day as early as late Eocene (Canavan et al., 2014; Quade et al., 2015). The deformation front propagated to the east through time (e.g., Carrapa et al., 2011; McQuarrie et al., 2005; Uba et al., 2006). In the late Eocene, the deformation front was located at the western plateau margin in both the northern Puna Plateau and the Altiplano Plateau (e.g., Carrapa et al., 2011). Sedimentary and deformation record along-strike to the north in the Altiplano correlates as well (e.g., DeCelles et al., 2011; Horton, 2005).

We argue that, during the late Eocene to early Oligocene, the southern Puna Plateau (~26-27ºS) was the southern continuation of this central Andes foreland system. First, the Geste Formation in the Pasto Grande region and Ns-4 in the PV region yield highly similar detrital AFT ages of ~35-120 Ma, suggesting their source region, likely the Sierra de la Quebrada Honda to the west, was likely undergoing upper-crustal shortening (Carrapa and DeCelles, 2008; Zhou et al., in revision; Chapter 2), consistent with its being in a wedge front position. Second, although most
of the southern Puna Plateau lacks paleoaltimetric data, the existing data from the western Salar de Antofalla region and indicate elevation as high as the present-day during late Eocene (Canavan et al., 2014; Quade et al., 2015), in line with development of a wedge-top depozone. Therefore, third, we infer that the thick (> 3 km) and westerly-derived units N-2 to N-5 in the ANT region reflect deposition in the foredeep depozone. The foreland basin may have been overfilled and therefore lacks a distinguishable forebulge. Fourth, the oldest, ~38-28 Ma strata in the PV region are thin (<0.5 km) and contain paleosol horizons, burrows and carbonate nodules indicating prolonged subaerial exposure, representing an ephemeral forebulge (Zhou et al., in revision; Chapter 2). Overfilling of the distal foredeep in the ANT region, the activation of the eastern source region and/or the high flexural righty may together lead to an overall overfilling of the regional foreland basin in the southern Puna Plateau. Additionally, in line with this reconstruction, the ANT region and the PV region may not have become elevated until after the late Eocene to early Oligocene.

Our reconstruction requires that the Cenozoic deformation front in the southern Puna Plateau (~24.5-25.5° S) was deflected to the west in the mapview relative to that in the northern Puna Plateau and the Altiplano Plateau (Fig. 15). We suggest that this deflection reflects larger amounts of shortening in the northern Puna Plateau and smaller amounts in the southern Puna Plateau, correlating with the different types of pre-existing structures and deposits in each segment. We suggest a potential role of inherited cratonic structures in modulating propagation of the Cenozoic orogenic front (Fig. 15). The Cretaceous rift developed mostly in the north northern Puna Plateau and its adjacent Eastern Cordillera, covering both regions with thick Cretaceous rift deposits (Marquillas et al., 2005). Rift structures are prone to reactivation during Cenozoic shortening (e.g., Carrera et al., 2006; Hongn et al., 2007). In the Eastern Cordillera adjacent to the northern Puna Plateau, shortening may have activated the inherited Cretaceous rift-related faults (e.g., Hongn et al., 2007); syn-depositional deformation in the Quebrada de los Colorados Formation is in line with this inference (del Papa et al., 2013). Although Cretaceous rift deposits are not present in the southern Puna Plateau (e.g., Marquillas et al., 2005), thermochronological data document Cretaceous cooling of some of the ranges on the eastern margin (e.g. La Quebrada, Carrapa et al., 2013; Sierra Laguna Blanca, Zhou et al., 2014), suggesting that the rift may have extended to the plateau. For example, rapid exhumation of the Sierra Chango Real (38-29 Ma AFT ages and relatively long ~13-14 µm mean track lengths;
Coutand et al., 2001) might be a result of inversion of rift-related faults. Alternatively, the southern Puna Plateau may lack significant rift structures, but may instead dominantly host relict Paleozoic topography, supported by some observed Paleozoic zircon (U-Th-Sm)/He ages from the bedrock ranges on the southern Puna Plateau and adjacent Sierras Pampeanas (Carrapa and DeCelles, 2015; Reiners et al., 2015, Zhou et al., 2014). Lacking rift-related faults may place the southern Puna Plateau less favored for accommodating large amounts of Cenozoic shortening, impeding the propagation of the Cenozoic orogenic front for the southern Puna Plateau (Fig. 15).

Formation of a regional depositional basin in the southern Puna Plateau during the late Eocene to early Oligocene, correlative with basins to the north, supports that, at the largest-scale, the central Andes behave as an orogenic wedge with a corresponding foreland basin created by lithospheric flexure (e.g., DeCelles et al., 2011). We find, however, that preexisting geological conditions, including inherited Cretaceous rift structures and relict Paleozoic topography, may exert a control in Cenozoic orogenic processes, modulating the growth of the central Andes along strike, slowing it in the south where pre-existing crustal heterogeneities are more pronounced (Fig. 15). Additionally, the late Oligocene to late Miocene evolution of the southern Puna suggests that, once the deformation front swept past, orogenic shortening may have continued to trigger distributed range uplift in the interior of the plateau (e.g., Carrapa et al., 2005; Zhou and Schoenbohm, 2015; Zhou et al., 2014; this study). Such a scenario resembles the well-documented Neogene Eastern Cordillera deformation history (e.g., Hain et al., 2011), implying that broken foreland dynamics plays a role in controlling evolution of the compressional basin-and-range system at regional scale in the southern Puna Plateau since the late Oligocene. Although we do not document paleoaltimetry directly, our provenance analysis suggests a probable low elevation for the southern Puna Plateau before early Oligocene time. Therefore, the elevation gain and development of the plateau morphology in the southern Puna Plateau would have taken place during the late Oligocene to late Miocene. The landscape change from a regional foreland basin to a “broken foreland” implies a change of dominant lithospheric behavior from flexure to local isostasy, which we argue to take place during the early to late Oligocene in the southern Puna Plateau.
3.9 Conclusion

The southern Puna Plateau is a key region for understanding along-strike variation and underlying dynamics of the central Andes. We present new sedimentary, structural, detrital zircon U-Pb ages, bedrock and detrital apatite fission-track data from the Antofagasta de la Sierra region on the southwestern Puna Plateau, and new bedrock zircon U-Pb data from the Sierra Laguna Blanca on the southeastern Puna Plateau. We find that the sandstone modal composition, detrital zircon U-Pb ages and modeled bedrock apatite fission-track data are consistent with the onset of sedimentary accumulation in the late Eocene to early Oligocene. Analysis of sedimentary data across the southern Puna Plateau reveals that the whole region was occupied by a regional sedimentary basin during the late Eocene to early Oligocene, bounded by topographic highs to the west in the Sierra Quebrada Honda and to the east in the Sierra Chango Real. This basin was compartmentalized by basement-cored range uplift starting in the late Oligocene. We find that the exhumation the Sierra Calalaste was 29-20 Ma (Carrapa et al., 2005; this study). We also document magmatic intrusion in the mid-late Miocene and may have influenced the detrital apatite fission-track studies in this and other arc basins.

Incorporation of the southern Puna Plateau into the regional east-propagating central Andes foreland basin starting as early as late Eocene highlights the potential for flexural behavior of the lithosphere, despite inherited crustal heterogeneities. Along-strike comparison suggests that the different types of inherited structures in the northern and southern Puna Plateau may explain the observed mapview deflection of the orogenic front for the late Eocene to early Oligocene Puna Plateau (Fig. 15).

The late Oligocene to late Miocene history of the southern Puna Plateau is characterized by distributed range uplift, similar to the observed history for the Neogene Eastern Cordillera. This suggests that the broken foreland dynamics, which involves isolated depocenters created in-between distributed ranges, governs the post-late Oligocene evolution of the southern Puna Plateau. We also find that the high elevation and low relief landscape in the southern Puna Plateau may be mainly established during this period of time because the deposition of the late Eocene to early Oligocene foredeep in the Antofagasta de la Sierra region was unlikely to have taken place at high elevation. This study reiterates that, in a complicated orogenic system such as
the central Andes, one must use different models to characterize orogenic processes in different time periods at different scales.
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<tr>
<td>ABD2 19</td>
<td>3.49</td>
<td>426</td>
<td>6.28</td>
<td>767</td>
<td>8.7694</td>
<td>3679</td>
<td>4.0</td>
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<tr>
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<td>496</td>
<td>28.56</td>
<td>4077</td>
<td>8.8172</td>
<td>3679</td>
<td>1.5</td>
<td>19.8</td>
<td>1.2</td>
<td>19.9</td>
<td>1.5</td>
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</table>

*pS and pl stand for density of spontaneous tracks and density of induced tracks, respectively.

†NS and NI stand for number of actual counted spontaneous tracks and induced tracks, respectively.

§pD stands for density of induced tracks to the CN5 dosimetry glass.

ND stands for the number of actual counted tracks for determining pD.

**We report (bold italic) pooled ages when P (χ^2)>5% and central ages when P (χ^2)<5%.

↑MTL=mean track length
## TABLE 2. RESULTS FOR APATITE (U-TH-SM)/HE THERMOCHRONOLOGY

<table>
<thead>
<tr>
<th>Aliquot ID</th>
<th>Age (Ma)*</th>
<th>±2σ</th>
<th>U (ppm)</th>
<th>Th (ppm)</th>
<th>147Sm (ppm)</th>
<th>eU (ppm)</th>
<th>Th/238U</th>
<th>He (nmol/g)</th>
<th>$F_T$</th>
<th>ESR (µm)**</th>
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<td>ABD22-a</td>
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<tr>
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<tr>
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<tr>
<td>ADR6-d</td>
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<td>47</td>
<td>7.73</td>
<td>1.24</td>
<td>0.75</td>
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</tr>
</tbody>
</table>

*$F_T$ corrected age.

†Weighted 2σ analytic error.

§eU=effective uranium concentration, a parameter that weights the decay of the two parents for their alpha productivity, by [U]+ 0.235*[Th] (Flowers et al., 2009).

#$F_T$ factor (Farley et al., 2006).

**ESR=equivalent spherical radius, recalculated from measured crystal dimensions.
Figure 1. (A) Topographic map for the central Andes based on SRTM DEM data. White line represents 3-km topographic contour. Blues line encompasses internal drainage region. Patched area denotes regions covered by syn-rift sediments related to the Cretaceous Salt rift (Marquillas et al., 2005). (B) Geological map for the Puna Plateau, compiled from Martinez, 1995; Schnurr et al., 2006 and Roy et al., 2006.
Figure 2. (next page). (A) Geological map of the Antofagasta de la Sierra region of NW Argentina. (B) Generalized stratigraphic column of Antofagasta de la Sierra. (C to G) Selected field photos of the sedimentary units
Figure 3. Detailed logged sections in the Antofagasta de la Sierra region of NW Argentina. Paleocurrent data were obtained by measure >10 of imbricated pebbles in the field. See Figure 2 for section locations.
Figure 4. Dickinson diagrams for sandstone modal compositions of sedimentary units on the southern Puna Plateau. See Figure 3 for sample locations.
Figure 5. (next page) Maps (A, B, D, F) and sketches (C, E, G) of magmatic intrusions (sills) in the studied strata. Maps are based on Google Earth images (http://www.earth.google.com). (H to K) Field photos of the sills. (J) Close-look at the sill, showing concentric weathering. (K) Contact between the sill and the host rock, showing chill margin.
Figure 6. (next page) Detrital zircon U-Pb ages, plotted with histogram and probability density function plots. Insets: Cenozoic single-grain ages from ADR1, ADR6 and ADR7. Number of ages plotted are shown in black (15% discordance) and in grey (40% discordance). See Figure 3 for sample locations.
Figure 7. Radial plots, histograms (with peak ages), track-length data and age-$D_{par}$ diagram for detrital apatite fission-track samples. See Figure 3 for sample locations.
Figure 8. (left) Apatite fission-track age data for bedrock samples from the Antofagasta de la Sierra region. (middle) Apatite fission-track length data for ABD2 and ABD22. (right) Apatite (U-Th-Sm)/He age and effective uranium concentrations (eU) for sample ABD22. See Figure 2 for sample locations.
Figure 9. HeFTy time-temperature (t-T) modeling results for Apatite fission-track data for bedrock samples from the Antofagasta de la Sierra region. See Figure 2 for sample locations. Grey, black and white lines represent acceptable, good and best-fit models, respectively.
Figure 10. AHe ages and effective U (eU) concentrations for apatites in detrital samples from the Antofagasta de la Sierra region. See Figure 3 for sample locations.
Figure 11. Generalized stratigraphic columns for the Puna Plateau. (upper) topographic swath profile for the southern Puna Plateau, extracted from a 100-km wide swath box using SRTM DEM data. Topographic line is shown in Fig. 1A
Figure 12. CL (cathodoluminescence) images for representative detrital zircons from the ANT region, from ADR4 and ADR2 (A), and ADR1 and ADR25. See Figure 3 for sample locations.
Figure 13. Zircon U-Pb geochronological results for the samples from the Sierra Laguna Blanca (the LB range). See Figure 1B for sample locations.
Figure 14. Results for K-S test for zircon U-Pb ages from the ANT region, PV region and the LB range. Grey boxes indicate that the corresponding two samples fail the K-S test. Detailed results are included in Data Repository.
Figure 15. Block diagram for the late Eocene to early Oligocene Puna Plateau region, illustrating the east-propagating foreland basin system and pre-existing structures/landscape to the east. N. Puna represents region of the Puna Plateau at ~24.5-25.5° S; S. Puna represents the region at ~26-27° S.
CHAPTER 4
TECTONICS AND LANDSCAPE EVOLUTION OF THE SOUTHERN PUNA PLATEAU, CENTRAL ANDES, SINCE THE LATE PALEOZOIC REVEALED BY MULTIPHASE BEDROCK COOLING HISTORY
4 Tectonics and landscape evolution of the southern Puna Plateau, central Andes, since the late Paleozoic revealed by multiphase bedrock cooling history

4.1 Abstract

Using multiple thermochronometers, we document three cooling events for the Sierra Laguna Blanca, the major bedrock mountain range on the southern Puna Plateau, finding that the range cooled during the late Paleozoic, the late Cretaceous and the mid-late Miocene. We use single-sample time-temperature modeling and landscape-thermochronological modeling to explore landscape change during those cooling events. The exhumation event during the late Cretaceous is likely associated with relief development, but the generated relief was reduced before the mid-late Miocene exhumation through surface processes (erosion and sedimentation). Our study supports initiation of a foreland basin during the late Eocene to early Oligocene which was compartmentalized since the late Oligocene by major, ~N-S-trending bedrock ranges such as the Sierra Laguna Blanca and Sierra de Calalaste. The relief of the Sierra Laguna Blanca established after ~20-10 Ma. The multi-phase evolution of the southern Puna Plateau reflects landscape evolution controlled by lithospheric flexure during the initial phase of Cenozoic shortening that was replaced by broken foreland dynamics by the mid-late Miocene, characteristic of the Neogene evolution of the present-day Eastern Cordillera as well. Additionally, the Salta rift plays an important role in defining total exhumation in the southern Puna Plateau and its margins, complicating the uses of low-temperature thermochronology in studying Cenozoic mountain building processes.

§ This chapter is under preparation for submission to Tectonics.

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4.2 Introduction

Low-temperature thermochronology has served as an important tool in deciphering mountain belt deformation and landscape development history for its abilities to record rock’s exhumation history since several kilometers on Earth surface. In regions with multiphase tectonics, utilities of thermochronology is challenging because low-temperature thermochronological system may be only partially reset or inherited signatures from prior tectonic events, complicating or even misleading its interpretation.

In long-lived subduction systems, such as the Nazca-South America system, the back-arc region experiences alternating extensional and contractional tectonism, offering an opportunity to advance utilities of low-temperature thermochronometers. Along the western South America system, the central Andes formed during crustal shortening during the Cenozoic, but this region is also home to rifting events in the Paleozoic and the late Cretaceous whose areal extends is much less constrained.

It has been suggested that the pre-Cenozoic geological conditions, such as deep-seated faults and relict high topography, may exert a control on the geometry and growth history for the Cenozoic orogens (Allmendinger et al., 1983; Carrera and Muñoz, 2008; Kley et al., 1999). On the one hand, the southern Puna Plateau currently resembles the first-order geometry, the compressional basin-and-range landscape, with the Eastern Cordillera, where prior extensional event, the Salta rifting, developed. Therefore, models suggest that the southern Puna Plateau may have also inherited preexisting structures that were randomly activated during the shortening, leading to the distributed range uplift throughout the southern Puna Plateau during the Cenozoic (e.g., Hain et al., 2011). On the other hand, the southern Puna Plateau is the southern continuation of the proposed Central Andean foreland basin (e.g., DeCelles et al., 2011), which hosted an eastward-propagating wave of deformation and exhumation during the Paleogene to the Miocene and does not host distributed range uplift for the early history during the Cenozoic.

We measure multi-system thermochronometers from the 6-km high Sierra Laguna Blanca, a preeminent bedrock mountain range on the southern Puna Plateau. We use zircon (U-Th-Sm)/He, apatite fission-track and apatite (U-Th-Sm)/He thermochronometry to explore its cooling and exhumation history since ~200 °C. Our goal is to use the exhumation history of the Sierra
Laguna Blanca to test the geologic and landscape models of the southern Puna Plateau. We also evaluate the uses of low-temperature thermochronometry in studying deformation of region of the southern central Andes at different times.

4.3 Background

The central Andes is composed of several tectonomorphic zones (Fig. 1A), including the Altiplano-Puna Plateau (the Central Andean Plateau), the world’s second largest continental plateau after the Tibetan Plateau. The Altiplano-Puna Plateau is an internally-drained with an elevation of over 3 km (Allmendinger et al., 1997; Isacks, 1988; Strecker et al., 2007) (Fig. 1A). It is divided at ~22.5°S latitude into the Altiplano Plateau to the north and the Puna Plateau to the south, is bounded by higher, more rugged Western and Eastern Cordilleras, and translates into the Sierras Pampeanas to the south and southeast (Fig. 1A).

Both the southern Puna Plateau and the adjacent Sierras Pampeanas are characterized by ~N-S-trending bedrock mountain ranges that are composed of crystalline basement and metamorphosed sedimentary rocks. They are bounded by deep-seated reverse faults and intervening sedimentary basins with various thicknesses ranging in age from the Paleozoic to the Cenozoic. Bedrock exposed in the southern central Andes are as old as >1000 Ma, representing the southern extent of the Antofalla terrane, one of the major continental cores of the South American craton (e.g., Ramos, 2008). During the early Paleozoic, the area of northwestern Argentina and northern Chile has been located in west-facing active margin (e.g., Bahlburg and Hervé, 1997), resulting in deposition of widespread, dominantly Ordovician marine, siliciclastic rocks and the formation of related metamorphic rocks (e.g., Allmendinger et al., 1983; Coira et al., 1982; Bahlburg and Hervé, 1997; Zimmermann and Bahlburg, 2003). After the Paleozoic, the region frequently experienced back-arc conditions along the long-lived subduction zone, and therefore experienced multiphase magmatism (e.g., Adams et al., 2011). Therefore, the southern Puna Plateau and the Sierras Pampeanas host distributed exposed plutonic rocks, notably belonging to the Cambrian-Ordovician and the Cretaceous.

The sedimentary strata exposed in the Serra Pampeanas region are generally older than those in the southern Puna Plateau and its adjacent Eastern Cordillera. The record of basin formation in the Sierras Pampeanas region dates to late Carboniferous (e.g., Aquino et al., 2014; Limarino and
Spalletti, 2006) and is characterized by strata produced in multi-generation extensional-transtensional basins (e.g., Coira et al., 1982). The Cretaceous sedimentary record is relatively well-documented, with extensional basins forming in the Triassic in the southern Sierras Pampeanas, progressing to the north through time (e.g., Legarreta and Uliana, 1996). The Salta rift evolved during the late Cretaceous in the southern Eastern Cordillera region, leaving thick sedimentary strata (e.g., Marquillas et al., 2005) (Fig. 1). The sedimentary strata associated with the Salta rift are absent from the southern Puna Plateau, where Cenozoic sedimentary rocks are mostly present. The earliest Cenozoic sedimentary record from the interior of the southern Puna Plateau is late Eocene in age (Adelmann, 2001; Kraemer et al., 1999; Zhou et al., in revision; Chapter 2 and Chapter 3), and was argued to have been deposited in a regional-scale basin (Zhou et al., in revision; Chapter 3).

The Sierra Laguna Blanca is a prominent ~N-S-trending bedrock mountain range composed of Paleozoic granite and Precambrian metamorphic rocks. The range is approximately W-E symmetric, rising from ~3.25 km at its base more than 2.5 km to its peak at 6012 m. The formation of this range is enigmatic. This range is characterized with steep slopes at two sides (Fig. 2B, C), implying a relatively young formation age. However, bedrock ranges close to the Sierra Laguna Blanca along the southeastern margin of the Puna Plateau exhumed in various times, from the late Cretaceous to the Eocene (e.g., Carrapa et al., 2013; Coutand et al., 2001).

4.4 Multi-thermochronometric vertical transect from Sierra Laguna Blanca

We obtained ten samples between ~3.5 and 5.5 km elevation from the eastern flank of the Sierra Laguna Blanca (Fig. 2). We performed zircon (U-Th-Sm)/He, apatite (U-Th-Sm)/He and apatite fission-track analysis on these samples.

4.4.1 Zircon (U-Th-Sm)/He thermochronology

Zircons (U-Th-Sm)/He thermochronological analysis were performed on five samples (elevations 5578, 4793, 4233, 3783, and 3596) at the University of Taxes at Austin (Table 1). Four single-grain aliquots were analyzed for each sample, except for 3783 in which three were analyzed. Single-aliquot ages overlap with each other within the 2σ error, enabling us to
calculate a weighted mean age for each sample (Table 1, Fig. 3). These ages range from early Carboniferous to early Triassic (247±27 Ma to 347±44 Ma), three of which yield similar ages at 270 Ma (elevations 5578, 4793 and 4233) (Table 1; Fig. 3A). The effective uranium concentration (eU) calculated from all aliquots ranges from ~80 to 350 ppm, relatively low compared to typical concentrations of uranium and thorium in natural zircon (5-4000 ppm and 2-2000 ppm, respectively) (Speer, 1980), and varies little within the each sample. There is no correlation between eU and ZHe ages among the five samples, or within single-grain aliquots from a single sample (Table 1; Fig. 3B).

4.4.2 Apatite fission-track thermochronology

Apatite fission-track (AFT) thermochronological analysis were conducted using the external detector method (Gleadow, 1981; Hurford and Green, 1983) at the Universität of Potsdam. We also prepared an aliquot from each sample with heavy ion irradiation at the Materials Research Department of the GSI Helmholtzzentrum (Darmstadt, Germany), in order to enhance yields of measurable confined tracks (Jonckheere et al., 2007). Detailed laboratory procedures are identical to those used in Zhou et al. (in revision; Chapter 3) and are included in the Data Repository. For each sample, we analyzed up to 48 grains and calculated AFT ages and Chi-squared ($\chi^2$) values using the Trackkey program (Dunkl, 2002), following the procedures of Galbraith (1981) (Fig. 4). We used the $\zeta$ calibration method (Hurford and Green, 1983) with a $\zeta$ value of 370.1±12.6 (R. Zhou). We also measured horizontal confined fission tracks from as many as possible c-axis parallel grains; only track-in-tracks were measured (Donelick et al., 2005). $D_{par}$ values (the etch figure length parallel to c-axis, Donelick et al., 2005) were measured and used to parameterize the kinetic properties for grains that were either counted or contain measured track lengths (Data Repository). All samples yield similar $D_{par}$ values ranging from ~1.4-1.8 µm and we do not observe significant $D_{par}$ variations among samples (Data Repository), implying that the measured apatite grains are mono-compositional and have a homogeneous closure temperature.

We report the pooled age from three samples (elevations 5578, 5369 and 3596) that pass $\chi^2$ test and report the central age for the rest of samples (Table 2; Fig. 4). The observed AFT ages range from 50.7±2.5 Ma to 67.1±3.6 Ma, without a display strong age-elevation relationship, as the youngest samples are located in the middle of the section (Fig. 5).
Mean track lengths display a unimodal distribution with concentrated track lengths at ~11-12 µm, but with a broad distribution from <8 µm to~14-15 µm tracks (Fig. 4). We suggest that the samples have experienced a complicated, non-monotonic cooling history: they may have spent an extended period of time in the PAZ in order to produce the short track length.

AFT ages suggest the possibility of a fault between the top seven and bottom three samples (Fig. 5), as sample 4025 yields old age of 64 Ma that is comparable to the top two samples (Table 2). Because this fault would have placed the top seven samples in a relatively higher structural position (lower temperature) compared to the block of lower three samples, the fault is either a west-dipping reverse fault or an east-dipping normal fault. Further, the location of this inferred faults coincide with a abrupt surface slope change (Fig. 2B, C), which either reflects a lithologic change due to faulting or recent activities of this fault.

### 4.4.3 Apatite (U-Th-Sm)/He thermochronology

Apatite (U-Th-Sm)/He thermochronological analysis (AHe) was performed jointly at the Universität Potsdam and Helmholtz-Zentrum Potsdam Deutsches GeoForschungsZentrum. Lab procedures are identical to those described in Zhou et al. (in revision; Chapter 3) and included in the Data Repository. We analyzed four to eight single-grain aliquots for each sample (Table 3; Fig. 5). Apatite grains were carefully hand-picked to avoid apparent fluid inclusion or internal crystal structures.

AHe ages vary significantly in most of the samples, ranging from ~30 Ma to 110 Ma. Except for the top two samples (elevation 5578 and 5369), all samples yield single-grain AHe ages that are older than the AFT age in the same sample (Fig. 5). Notably, the analyzed apatites have high (>100 ppm) eU concentrations, including some with outstandingly high eU concentrations (several hundreds of ppm) (Fig. 6). Analyses with eU<100 ppm often yield the youngest ages within each sample (Fig. 5). Our data also show a crude correlation between eU and AHe ages (Table 3; Figure 6A, B), suggesting control by radiation damage on the dispersed old AHe ages.

### 4.5 Thermal history for the Sierra Laguna Blanca

Our results from multi-system thermochronology constrain the <~200 °C time-temperature (t-T) history of the Sierra Laguna Blanca. We first interpret its t-T history based on observed
thermochronological ages. We then conduct single-sample inverse modeling and refine the t-T history using the program HeFTy (Ketcham, 2005).

We argue against the influence of radiation damage on the ZHe-dated zircons because of their modest U and Th concentrations and poor eU-ZHe age correlation (Table 1; Fig. 3). We also rule out the possibility of significant resetting of the ZHe system by subsequent temperature rise into the ZHe PRZ because the observed AFT ages are notably younger than ZHe ages, implying that the subsequent thermal event(s) that reset the AFT system did not affect the ZHe system. Therefore the measured ~350-250 Ma ZHe ages reflect the cooling of the Sierra Laguna Blanca before the early Triassic (~250 Ma). Dominant Permian ZHe ages at ~280-240 Ma and their constancy with elevation favor a fast cooling event during mid-late Permian time that brought up at least ~2-km of basement rock through the effective closure temperature. Since then, the measured samples stayed at least above their corresponding ZHe effective closure temperature.

Because the AFT ages range ~50-68 Ma and are associated with shortened track-lengths, a younger cooling event that brought all samples from the AFT PAZ to the surface temperature is required after the observed AFT ages. The variability in our AHe ages demonstrates poor resolution for the lowest, ~85 to 40°C temperature history of the range, possibly because of accumulated radiation damage evidenced by very high eU concentration and a correlation between AHe ages and eU concentrations. Therefore, young (~20-30 Ma) AHe ages that have low eU (less than 100 ppm) may better reflect the most recent cooling event, implying late Oligocene to early-mid-Miocene cooling of the Sierra Laguna Blanca since cooling through the AHe PRZ (~85 to 40°C).

In order to better refine t-T histories, we conducted inverse modeling of individual samples with HeFTy (Ketcham, 2005) using the following strategies. For a sample that has ZHe data (Fig. 7A, C, E), models start at 340-350 Ma with temperatures at 200-220 °C. This allows full annealing of all three systems (AHe, AFT and ZHe) prior to 340 Ma (e.g., Reiners and Brandon, 2006). The models are constrained at 20-160 °C from 340 Ma to present, consistent with observed ZHe ages. For a sample that does not have ZHe data (Fig. B, D), models start at 140-150 Ma with temperatures at 200-220 °C, and are constrained at 20-160 °C from 140 Ma to present. In order to fully explore the t-T space, we place several t-T boxes covering 20-160 °C (Fig. 7). For all samples with ZHe data, we use the aliquot that is the closest to the weighted mean age as the
model input. For a sample that yields AFT ages with $P(\chi^2) > 5\%$, all AFT age and length data are used. When $P(\chi^2) < 5\%$, a subset of AFT ages and length data, grouped by their Dpar values (Ketcham, 2005), are used. For AHe ages, we use only measured aliquots if eU <100 ppm. The ZHe, AFT and AHe data are modeled using thermochronological kinetic models reported by Guenthner et al. (2013), Ketcham (2005) and Flowers et al. (2009), respectively.

The t-T models reveal three cooling events for the Sierra Laguna Blanca (Fig. 7). First, as evidenced by models with ZHe data input (e.g., sample 5578, Fig. 7), the Laguna Blanca range remains at <160-140 °C since at least 250-200 Ma. The modeled oldest apatite fission tracks in all samples are dominantly younger, at 100-60 Ma (except for one model, 4025-B, which yields the oldest preserved track at 260 Ma) (Fig. 7). Therefore, the amount of cooling before the early Cretaceous (~250-200 Ma) is not constrained by our data. Second, temperatures are confined to > ~80 °C during the late Cretaceous (~100-60 Ma) and all samples are cooler than ~80-60 °C by 50 Ma, documenting a late Cretaceous event that cooled the samples for at least 20-40 °C. Notably, the t-T paths occupy a wide temperature range of ~80-60 °C to ~20 °C, which permit that this cooling event brought the samples to the surface. However, we caution that the amount of cooling, <~60 °C, the low-temperature boundary of the AFT PAZ, cannot be constrained without AHe data. Third, the final cooling event takes place at ~15-10 Ma and brings all samples to the surface temperature (Fig. 7). Finally, the amount of cooling during the late Paleozoic and the Cretaceous remains less constrained, as we can only constrain the minimum temperature decrease at each cooling event. This unknown factor is critical because it is directly related to whether or not a subsequent temperature increase is required prior to the next cooling event (Fig. 7). We attempt to address this issue in the landscape-thermochronogical modeling in the following section.

4.6 Topographic development for the Sierra Laguna Blanca

We further investigate the landscape development history of the Sierra Laguna Blanca region by exploring the timing of the major relief generation and testing potential burial during the late Eocene to early Oligocene (e.g., Zhou et al., in revision; Chapter 2) (Fig. 8A). Because our data do not provide strong t-T constraints cooler than the ZHe PRZ and hotter than the AFT PAZ, we restrict our modeling to the post-mid-Cretaceous landscape evolution (Fig. 8A). We use Pecube (Braun, 2003), a finite-element code that solves the three-dimensional heat transport equation in
an evolving crustal/lithospheric block. We first use the inverse mode in Pecube to search for best-fit values for landscape parameters. Then we conduct forward modeling, using the searched parameter values, in order to track the t-T history of the range.

Several assumptions are made to facilitate our modeling. First, we treat the modeled mountain range as a fault-free block and only use the top seven samples in order to minimize complications caused by faulting. Second, the observed cooling events are represented by vertical tectonic uplift that takes place uniformly across the entire modeled landscape. Because the purpose of the modeling is to better refine the relief change history, our models do not constrain or provide insights into absolute elevation change of the modeled region.

We model two tectonic events at 75-52 Ma (Rock uplift 1) and 16-4 Ma (Rock uplift 2) based on general trends in our single-sample t-T models, with the rock uplift rate at the two time intervals varying from 0-1 mm/yr (Fig. 8A). The topographic conditions at 75, 52, 16 and 0 Ma are constrained (Fig. 8A). At each time, the topography is parameterized with base elevation (B) and relief (R), scaling the modern Sierra Laguna Blanca topography, which currently has a relief of 2.75 km (Fig. 8A). At 75 Ma, the relief is set to zero, representing no relief before the Cretaceous exhumation. At 52 Ma, the model is allowed to generate a portion of the modern relief (0-2.75 km, 0-100% of R0, Fig. 8), which is then totally leveled by a higher (B1>R2) surface. Because there is no tectonic event (rock uplift) between 52 and 16 Ma, topographic development during this time must take place through erosion and deposition (Braun, 2003). The modern (t=0 Ma) topography is represented by a 2.75-km relief (Fig. 8A).

In each model run, one value is picked based on the Neighbourhood Algorithm (NA) (Sambridge, 1999), from each parameter range, in order to calculate AFT ages and AFT track-length distributions for each sample location. A misfit value is then calculated by comparing the modeled and observed ages. ZHe data are not used in the inverse model because we focus on testing the Cretaceous-Cenozoic history of the region. Because Pecube does not include radiation damage models for AHe system (Braun, 2003), we do not model the AHe ages either.

We obtained 754 models that produce the topographic set-up (Data Repository). We use the sets of parameter values that yield the lowest misfit (<0.04, n=11) to run forward models in order to track the modeled t-T histories that yield the best-fit AFT ages (Fig. 8B; Data Repository). The
purpose of forward models is to track t-T histories at the sample locations under different topographic conditions, and to explore the amount of cooling, particularly in the Cretaceous which is unconstrained from single-sample t-T models. Because rock particles in models move vertically and there is no change of relative positions between samples, the modeled AFT ages along the vertical transect do not reflect geological processes such as faulting or block rotation. The following observations are made from the modeling results (Fig. 8B). First, at 75 Ma, the top portion of our modeled section lies within the bottom part of the AFT PAZ while the lower portion is hotter than the PAZ, and is totally annealed. Second, the cooling event driven by the late Cretaceous rock uplift brings the top of the section to as low as < 50 °C, and the bottom of the section to ~80-60 °C, within the PAZ. Third, during the tectonic quiescent period from 52-16 Ma, heating of the section may take place, but the amount of heating is constrained to be less than ~30°C (Fig. 8B). Before the final cooling, the samples may stay at the upper part of the AFT PAZ (~80°C) (Fig. 8B).

4.7 Tectonics and landscape evolution of the southern Puna Plateau

Our multi-system thermochronological data constrain the t-T history and landscape evolution of the Sierra Laguna Blanca; these constraints in turn allow us to evaluate models for the tectonic and landscape evolution of the southern Puna Plateau since the late Paleozoic (Fig. 9).

We find that the earliest cooling event recorded at the Sierra Laguna Blanca took place in the mid-Permian (~270-280 Ma) and cooled the bedrock samples from >220°C to <180°C (Fig. 7). This is consistent with documented Carboniferous and Permian cooling from K-feldspar ⁴⁰Ar/³⁹Ar data along the eastern margin of the Puna Plateau and the western Eastern Cordillera (Insel et al., 2012). Further, during the Permian, the Paganzo basin to the south experienced subsidence and received sediments from the elevated Pampean arc immediately to the north (Fig. 9D) (Enkelmann et al., 2014). To the west of the Sierra Laguna Blanca, Permian-aged strata are present in the Salar de Antofalla region, suggesting active subsidence there as well (Adelmann, 2001; Kraemer et al., 1999). Therefore, we suggest that the southern Puna Plateau during the late Paleozoic was composed of two parts; the eastern, elevated region was likely the northern continuation of the Pampean arc (Fig. 9D), whereas the western part was occupied by active
depocenters which may be part of the Navidad-Arizaro basin (e.g., Limarino et al., 2006; Limarino and Spalletti, 2006) (Fig. 9D).

Our single-sample t-T models reveal that the cooling of the Sierra Laguna Blanca from ~130 °C (the high-temperature boundary of the AFT PAZ) took place as early as ~100 Ma to 70-60 Ma, and cooled the samples to at least 40-60 °C by ~50 Ma (Fig. 7), representing a minimum 2-3 km exhumation assuming a 20-30 °C/km geothermal gradient. This late Cretaceous to early Cenozoic event most likely reflects cooling related to the Salta rift. The Salta rift evolved in northwestern Argentina in the Cretaceous to the early Cenozoic and formed several depocenters with irregularly distributed topographic highs (e.g., Marquillas et al., 2005; Salfity and Marquillas, 1994). Bedrock ranges from the southeastern margin of the Puna Plateau, such as the Laguna Brava and La Quebrada ranges yield Cretaceous AFT ages (Carrapa et al., 2013) (Fig. 1B). Similar cooling is also obtained by inverse modeling of apatite fission-track age and length data (Mortimer et al., 2007) (Fig. 1B). Because of the absence of rift-related sedimentary strata in these locations, the southern and eastern Puna Plateau were likely located in the rift shoulders (e.g., Carrapa et al., 2013; Deeken et al., 2006; Mortimer et al., 2007; Sobel and Strecker, 2003) (Fig. 9C). We find that, as at several other locations along the margin of the southeastern Puna Plateau (e.g., Carrapa et al., 2013), exhumation related to the Salta rift plays as a major role in defining the total exhumation the Sierra Laguna Blanca.

Our samples require a subsurface temperature of ~40-60 °C during ~30-20 Ma or perhaps earlier (Fig. 7), supporting the recent proposal that a regional sedimentary basin formed in the southern Puna Plateau during the late Eocene to early Oligocene (~38-28 Ma) (Zhou et al., in revision; Chapter 2). Zhou et al. (in revision; Chapter 2) propose that although the basin was sourced from both the east and west, the western source, arising in a region of active crustal shortening, was dominant, and the basin depocenter was correspondingly shifted to the west (Fig. 9B). Therefore the Sierra Laguna Blanca may have only been shallowly buried, meaning that AFT thermochronometers were not significantly reset.

The most recent cooling of the Sierra Laguna Blanca has taken place since ~20-15 Ma (Fig. 7), resulting in the formation of the present-day topography at the Sierra Laguna Blanca (Fig. 9A). Upper-crustal shortening during the late Miocene (~12 Ma) has also been documented in the adjacent Pasto Ventura region (Schoenbohm and Carrapa, 2015; Zhou and Schoenbohm, 2015),
resulting in compartmentalization of the regional, late Eocene-early Oligocene sedimentary basin since the mid-Miocene.

Our analysis suggests that, prior to the Cenozoic plateau formation, the southern central Andes were home to highly a variable landscape, hosting distributed bedrock ranges and extensional/transtensional basins (e.g., Limarino and Spalletti, 2006). However, those preexisting structures may not have been reactivated at the beginning of the Cenozoic shortening, as various lines of evidence support formation of a regional basin covering the southern Puna Plateau (Zhou et al., in revision; Chapter 2). This is an important observation since it suggests that the lithosphere may experience flexural subsidence in response to the crustal load, forming a foreland basin in spite of the presence of inherited crustal heterogeneities. Our single-sample t-T models suggest the cessation of the rift-related cooling at ~50 Ma, which predates the transition from extensional to flexural basin processes in the southern Puna Plateau (Fig. 7). The older strata related to the regional foreland basin in the southern Puna Plateau are estimated to be ~38-28 Ma (Zhou et al., in revision; Chapter 2 and 3), which may post-date the onset of flexural behavior.

Furthermore, the amount of exhumation for the bedrock is strikingly different between the southern Puna Plateau and the Sierras Pampeanas (Fig. 2B). In the Sierras Pampeanas, ZHe and AFT ages that are frequently Triassic or older (>200 Ma) suggest major cooling associated with long-lived extensional tectonism in the southern central Andes; Cenozoic cooling, however, is limited (e.g., Davila and Carter, 2013; Enkelmann et al., 2014; Löbens et al., 2011). Cenozoic events have rarely caused significant exhumation in the Sierra Pampeanas, with Cenozoic cooling largely confined to the Precordilleran thrust front (e.g., Davila, 2010) (Fig. 1B). In contrast, both young ZHe ages and fully-reset AFT ages are present along rift shoulders or the marginal zone of rift basins (Fig. 1B). Cenozoic upper-crustal shortening has caused significant cooling in the Eastern Cordillera, as documented by <50 Ma ZHe ages, which require ~8-10 km exhumation (e.g., Pearson et al., 2012). We also find that the Sierra Laguna Blanca was exhumed to the surface since 20-15 Ma, post-dating the formation of the regional basin. Taken together, we suggest that inherited structures from the Salta rift play an important role in initiating and focusing crustal shortening, but that this deformation may not take place until after the formation of a foreland basin.
4.8 Conclusion

We document the cooling history of the Sierra Laguna Blanca, the major bedrock mountain range on the southern Puna Plateau along a 2-km vertical transect. Zircon (U-Th-Sm)/He ages ranging from ~300-250 Ma suggest cooling through ~200 to 130 °C during the Permian. Modeling of thermochronological data, influenced most heavily by apatite fission-track length data, reveals that the Sierra Laguna Blanca cooled subsequently during the late Cretaceous (~90-50 Ma) and the mid-late Miocene (~20-10 Ma). Landscape modeling with thermochronological data supports the formation of a regional basin covering the southern Puna Plateau (Zhou et al., in revision; Chapter 2 and Chapter 3), which reduced the topographic relief created during the Cretaceous exhumation.

We suggest that during the late Paleozoic the southeastern Puna Plateau was a topographic high, bounded by sedimentary basins to the west (in the southwestern Puna Plateau) and to the south (in the southern Sierras Pampeanas region). We also find that the southeastern Puna Plateau was located in the rift shoulders of the Salta rift during the late Cretaceous, documenting the first evidence showing that inherited Cretaceous rift structures in the interior of the Puna Plateau. The Salta rift play an important role in defining the total exhumation observed in the southern Puna Plateau and its adjacent Eastern Cordillera.

We find that the pre-Cenozoic landscape for the southern Puna Plateau region is highly variable due to multiple tectonic events. However, our data supports that a regional foreland basin on the southern Puna Plateau, likely covering the Sierra Laguna Blanca, during the late Eocene to early Oligocene (Zhou et al., in review; Chapter 2). Formation of this basin suggests that the lithosphere during the initial phase of Cenozoic shortening behaved flexurally, contrasting with the likely local isostatic compensation during the extensional phase related to the Salta rift. We suggest such a transition took place ~50 to 38-20 Ma, based on the timing for the cessation of rift related exhumation and the oldest foreland basin strata.

The final cooling of the Sierra Laguna Blanca and the establishment the current relief took place ~20-10 Ma, based on our modeling results. Combined with documented exhumation for the Calalaste range to the west during the late Oligocene, these distributed range uplift across the
southern Puna Plateau is in line with the broken foreland dynamics starting during the late Oligocene.
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$^\dagger F_T$ correction factor, calculated based on Farley et al. (1996)

$^\S$ ESR = equivalent spherical radius, calculated from measured grain dimensions.
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*pS and pl stand for density of spontaneous tracks and density of induced tracks, respectively.
†NS and NI stand for number of actual counted spontaneous tracks and induced tracks, respectively.
§$pD$ stands for density of induced tracks to the CN5 dosimetry glass.
#ND stands for the number of actual counted tracks for determining $pD$.
**We report (bold italic) pooled ages for samples with $P (\chi^2)>5\%$ and central ages for samples with $P (\chi^2)<5\%$. 

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†F_T correction factor, calculated based on Farley et al. (1996)

§ESR = equivalent spherical radius, calculated from measured grain dimensions.
Figure 1. (A) Overview and tectonomorphic zones of the central Andes. (B) Bedrock cooling ages documented by zircon (U-Th-Sm)/He and apatite fission-track thermochronology in the southern Puna Plateau, southern Eastern Cordillera, Santa Barbara System and Sierras Pampeanas. Yellow line encompasses regions with documented Salta rift deposits (Marquillas et al., 2005). Blue contour lines are the depth of subducting Nazca oceanic plate (Cahill and Isacks, 1992; Ramos and Folguera, 2009). White line indicates the internal drainage region (the plateau region) derived from the SRTM (Shuttle Radar Topography Mission) data. Black lines represent major thick-skinned thrust faults. Bedrock cooling data are from Carrapa et al., 2013, 2006, 2005; Coughlin, 2000; Coughlin et al., 1998; Coutand et al., 2001; Davila, 2010; Deeken et al., 2006; Enkelmann et al., 2014; Jordan et al., 1989; Löbens et al., 2013, 2011; Mortimer et al., 2007; Nalpas et al., 2005; Pearson et al., 2012; Reiners et al., 2015; Richardson et al., 2013; Safipour et al., 2015; Sobel and Strecker, 2003; Vergés et al., 2001.
Figure 2. (A) Topographic map of the Sierra Laguna Blanca. (B) Slope map of the Sierra Laguna Blanca. (C) Topographic swath (black line) and slope swath (red line) profiles extracted from a 10 km-wide box indicate by a single red line in (A) and (B). Sample locations are shown in blue dots.
Figure 3. Zircon (U-Th-Sm)/He data for all measured samples.
Figure 4. Apatite fission-track data for all measured samples. Each panel contains the radial plots of AFT ages and the track-length distribution.
Figure 5. (A) Apatite (U-Th-Sm)/He ages with sample elevation. Each elevation horizon contains all measured AHe ages from the same sample. Enlarged sample symbols indicate AHe analysis with low (<100 ppm) effective uranium (eU) concentration. Red bars indicate the AFT ages from the same sample. (B) Plots showing only AHe analysis with low (<100 ppm) effective uranium (eU) concentration.
Figure 6. (A) Individual apatite (U-Th-Sm)/He ages and eU concentrations. (B) Sample-averaged apatite (U-Th-Sm)/He ages and eU concentrations. Note that the axes are in log-scale.
Figure 7 (next page). Single-sample t-T inverse models using the HeFTy program (Ketcham, 2005). The red boxes represent t-T constraints as input. Light blue zones are acceptable t-T paths which produce thermochronologic data with >0.05 agreement with the observed data. Dark blue zones are good t-T paths which produce thermochronologic data with >0.5 agreement with the observed data. In all models, the white lines are the best-fit t-T paths that produce thermochronologic data with the highest agreement. The dots on white lines represent the age of the oldest preserved apatite fission track.
Figure 8. (A) Model set-up for Pecube-based landscape evolution models. (B) Results for all forward models using searched parameter values that yield the lowest misfit values.
Figure 9. Landscape evolution of the southern central Andes since the late Paleozoic.
CHAPTER 5

LATE MIOCENE UPPER-CRUSTAL DEFORMATION WITHIN THE INTERIOR OF THE SOUTHERN PUNA PLATEAU, CENTRAL ANDES
5 Late Miocene upper-crustal deformation within the interior of the southern Puna Plateau, central Andes**

5.1 Abstract

The origin and evolution of the central Andes, a non-collisional orogenic system, has been hypothesized to evolve with 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, re-activation 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, SIMS U-Pb dating of zircon from interbedded volcanic ashes, and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of volcanics, we show that major basin formation started $\sim$11.7–10.5 Ma and continued until at least $\sim$7.8 Ma. The basin underwent syn-depositional faulting and folding from $\sim$10–8 Ma. Contractional deformation in the Pasto Ventura basin ended between $\sim$7.3 and 4 Ma, based on the onset of regional horizontal extension. Data from the Pasto Ventura region allows 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 represents 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

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detachment of a late Miocene lithospheric drip in shaping the upper crustal deformation on the southern Puna Plateau since mid-late Miocene.

5.2 Introduction

Studies of the central Andes have enriched our geologic understanding of orogenic plateaus, such as the Altiplano-Puna Plateau, along non-collisional 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., Capitanio 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, 2002). Yet, horizontal forcing may preferentially activate pre-existing crustal weak zones inherited from the Cretaceous rifting event, causing distributed, out-of-sequence crustal shortening (e.g., Sobel et al., 2003; Sobel and Strecker, 2003). Neogene compressional basins, bounded by orogen-parallel and NE-striking fault zones, form rhomb-shaped crustal segments in mapview (Riller et al., 2012; Riller and Oncken, 2003) (Fig. 1B). 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 Garzzone, 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; Strecker 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 SIMS zircon U-Pb geochronology of interbedded ash beds and $^{40}$Ar/$^{39}$Ar 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 explore possible geodynamic processes in the southern Puna Plateau.

5.3 Background

The Andes are 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 (>3km), low-relief interior plateau, surrounded by the higher, more rugged Western and Eastern Cordilleras, the Sierras Pampeanas and the Santa Barbara ranges (Fig. 1) (Allmendinger et al., 1997; Isacks, 1988; Strecker et al., 2007). The plateau region currently experiences semi-arid 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) (Fig. 1). Geophysical investigations indicate that the crustal thickness in the Puna is spatially variable, but ranges from ~50–68 km (McGlashan et al., 2008; Whitman et al., 1992, 1996; Yuan et al., 2002).
Cenozoic arc volcanism is widespread along the Western Cordillera and the western Altiplano-Puna Plateau (Fig. 1). The post-late Miocene Altiplano-Puna Volcanic Complex (APVC) between latitudes of 21°–24°S, covering ~50,000 km², represents an intense episode of felsic volcanism of the central Andes (de Silva, 1989) (Fig. 3A). Ignimbritic complexes are also present across the Puna Plateau, often as collapsed calderas and are offset from the modern arc (e.g., Kay et al., 2010; Guzmán and Petrinovic, 2010; Guzmán et al., 2011) (Fig. 8). Across the Puna Plateau there are well-preserved monogenetic volcanic cinder cones and associated lava flows; these are dominantly mafic, late Miocene to Recent, confined to the plateau region, and compositionally distinct from arc volcanics (Kay et al., 1994; Risse et al., 2008) (Fig. 2 and Fig. 8). Pre-late Miocene mafic-intermediate volcanic lavas are also present on the southern Puna Plateau, but their extents and ages are poorly constrained, especially for the southern Puna (Roy et al., 2006; Schnurr et al., 2006) (Fig. 2A).

The deformation history of the Puna Plateau, the adjacent Eastern Cordillera and the Sierras Pampeanas has been studied for several decades, where shortening is primarily accommodated by steeply dipping, bivergent 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 (~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 (e.g., Riller and Oncken, 2003) (Fig. 1b). 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 which predate, overlap, or post-date 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 ~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 ~14 Ma (Alonso, et al., 1991; Boyd, 2010; Marrett et al., 1994), consistent with proposed 15–8 Ma exhumation based on thermochronology data (Carrapa et al., 2009) (Fig. 3C, section A–A’). To the east of the northern Puna Plateau in the Eastern Cordillera, apatite fission track thermochronology indicates range-scale exhumation and adjacent intramontane basin development migrating eastward into the Eastern Cordillera (CLR, Cumbres de Luracatao range) by 21 Ma (Fig. 3C, section A–A’) (Deeken et al., 2006). At the southern end of the Eastern Cordillera, syn-depositional deformation occurred at ~40 Ma (Hongn et al., 2007) (Fig. 3C, section A–A’).

In the southern Puna Plateau, especially southern 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) that formed from 6.6 to 2 Ma (Kay et al., 2010) (Fig. 3), obscuring observations of the underlying geology in much of the region. The 29–24 Ma exhumation of the Calalaste range west to 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 the southeastern flank of the Puna Plateau was exhumed ~38–29 Ma (Coutand et al., 2001) (Fig. 3B, C). Farther east, U-Pb geochronology of intercalated ashes and detrital fission track data from strata within the El Cajon 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 ~10–6 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 (Sobel and Strecker, 2003) (Fig. 3C, section B–B′). 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 ~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 is in line with estimate of pre-9 Ma local relief establishment (Montero-López et al., 2014).

5.4 Methods and Results

5.4.1 Geological mapping in the Pasto Ventura basin, NW Argentina

The landscape of the south Puna Plateau is dominantly covered by recent lava flows, ignimbrites and colluviums. The Pasto Ventura region (~26°48′S; ~67°16′W) is located on the southern Puna Plateau (Fig. 2) and outcrops a narrow, elongated, roughly N-S striking basin that contains sedimentary rocks of primarily Neogene age (Schoenbohm and Carrapa, 2015; Zhou et al., 2013). The Pasto Ventura basin is the only exposed, relatively continuous sedimentary record in the southernmost Puna Plateau (Fig. 2A). We map 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 ~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 flows. The clasts are generally ~5–10 cm in size and 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) (Fig. 5 and Fig. 6). The thickness of these units varies across the region, ranging from 100 to 200 m to up to one kilometer, and 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 ~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-bedded red sandstone and siltstone with occasional cobble layers (~5–10 cm in thickness) and pedogenic carbonate nodules (Fig. 4F). 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 unit Ns-2 and –1, is interbedded with volcanic ash layers or
pumice layers, with thicknesses varying from several centimeters to ~1 m (Fig. 4B). Locally, cross-bedding 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, gypsum deposits are found as thin (1–2cm) 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, and 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-bedding 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 comprises 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 ~2cm) with sparse olivine; the groundmass contains plagioclase and olivine that has been mostly altered to iddingsite and calcite-filled vesicles are also present (Fig. DR3; see footnote 1). Unit Nfl is overlain by unit Ns-3 red mudstone, and therefore predates deposition of Ns-3, Ns-2 and Ns-1. However, the relationship between Nfl and Ns-4 is unconstrained by mapping. We performed geochemical and $^{40}\text{Ar}/^{39}\text{Ar}$ 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.

The bedrock (metamorphic basement) unit (M) in this area belongs to the Puncoviscana Formation and consists of clastic, weakly metamorphosed, sedimentary rocks (Allmendinger et al., 1997; Ramos, 2008) (Fig. 4F). On images, the bedrock is characterized by high topography and a fracture-rich, dark-colored appearance (Zhou et al., 2013).
5.4.2 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 UCLA SIMS Laboratory following methods described by Schmitt et al. (2003).

We dated 16 ashes, six of which are from Ns-3 and 10 of which are from Ns-2 (Table 1). For each ash, ~10–20 grains were dated. The tips or rims of zircons were probed and dated in order to avoid any inherited old cores. The age of an ash is 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). Some ages overlap at the 2s level of uncertainties (Table 1). It may due to the fact that they were deposited at the same time. Alternatively, they may be separated by an interval of time less than is resolvable by this dating technique. 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 ± 0.10–9.8 ± 0.10 Ma. Those from Ns-2 range from 9.59 ± 0.10–8.32 ± 0.12 Ma. In addition, Schoenbohm and Carrapa (2015) report 7.77 ± 0.21 Ma and 7.88 ± 0.58 Ma ashes from Ns-1, and a 10.50 ± 0.10 Ma for Ns-2 (Table 1).

5.4.3 40Ar/39Ar geochronology of lava flows and monogenetic cinder cones

The 40Ar/39Ar geochronological measurements were conducted at the USGS 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 tabulated results are given in the Appendix C.

Unit Nfl was deformed and exhumed in a dome-like feature, as the result of interference between Fold-1 and Fold 2 structures (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. An ⁴⁰Ar/³⁹Ar plateau age (3 or more contiguous heating steps comprising ≥ 50% of the ³⁹Ar 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 ³⁹Ar 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 500–800 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 Emerman, 1992; Kay et al., 1994; Ruch and Walter, 2010). In this study, we compile existing geochronological data for the mafic monogenetic volcanism for the Pasto Ventura region, and add three new ⁴⁰Ar/³⁹Ar ages (Table 2). Five mafic volcanic samples (PV07-PV1–02, P07PVSUR3–1, P07SUR1–01, P09B-09, P09B-12) were collected for geochronology using ⁴⁰Ar/³⁹Ar step-heating analyses and yield ages of 0.68 ± 0.06 Ma (sample PV07-PV1–02), 0.42 ± 0.05 Ma (sample P07PVSUR3–1), 0.43 ± 0.07 Ma (sample P07SUR1–01), 0.57 ± 0.04 Ma (sample P09B-09) and 0.45 ± 0.02 Ma (sample P09B-012) (Table 2).

5.4.4 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 (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) (Fig. 8, Table 3).

The geochemical similarities between samples PV11B-01 and PV11B-04 confirm our field and petrographic observations for the extent of unit Nfl (Fig. 8). 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 et al., 2006) (Fig. 8). 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 (Fig. 8B, C). 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) (Fig. 8, 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 for Ba and Nb, are also depleted (Fig. 8). 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 (e.g., Roy et al., 2006; Schnurr, 2006), whose age has not previously been documented.

5.4.5 Deformation analysis: faults, folds and syn-depositional deformation

A major NW-dipping thrust fault (Fault-1) with a ~N32°E strike cuts much of the NPV map, dying 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 ~0.8 Ma and ~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 as 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 eolian deposition and pervasive physical weathering and colluvial transport of material, we were unable to locate any well-preserved exposures of the fault plane. However, the fault trace on the surface is clearly evident by a distinct topographic rise and therefore it enables us to calculate a ~33° to 42° 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 35° to the east. Displacement along the ~2.7 km long fault decreases to both N and S (Fig. 5). Fault-3 is also present to the east of Fault-1, but in contrast to Fault-2, dips to the east (toward ~N80°E). This fault carries Ns-3 above Ns-2. Along this fault, the maximum displacement is estimated to be ~0.5 km (Fig. 5, section B–B′).
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 a ~N30°E strike in the north and a ~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 by 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%, Fi. 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 post-dates 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 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 Ns-3, consistent with primary lavas flow within unit Nfl that were once flat-lying. Deformation by ~NNE-striking folding and thrust faulting, and possibly subsequent anticline folding (Fold-2) tilted unit Nfl. Therefore unit Nfl (11.7 Ma, see sections below) 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 sub-vertical (∼80°, ∼85°, ∼80°, respectively), but dip slightly to the west. The amount of shortening estimated from the cross-section 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 ∼WNW-ESE contraction. We are able to constrain the amount of shortening based on cross-sections. The amount of ∼E-W shortening ranges from ∼35%–45% to 10%–15% (Fig. 5 and Fig. 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 syn-depositional deformation in the Pasto Ventura basin. In the SPV map, an angular unconformity formed within the Ns-3 mudstone-siltstone (Fig. 7A, B). Units below and above the angular unconformity are indistinguishable in lithology but are oriented differently (the beds above the unconformity are oriented 230/06; the beds below the unconformity are oriented 216/28) (Fig. 7A, B). SIMS-dated zircons from ashes above this unconformity yield an age of 9.99 ± 0.12 Ma (sample PV10A-06), and ashes dated below yield an age of 9.80 ± 0.10 Ma (sample PV10A-03), which are the same within uncertainty, indicating only a short interval of section missing across the unconformity. In the NPV map, an internal unconformity is found within the lower Ns-2 of the NPV map (Fig. 7C, D), in which older beds are folded and truncated. Schoenbohm and Carrapa (2015) were able to retro-deform the beds (right of the photo view in Fig. 7C, D) and showed that their geometry satisfies features for syn-depositional fault-bend folding. Volcanic ashes within this unit, from below and above the
observed internal unconformity, yield ages of $9.22 \pm 0.08$ Ma (sample PV11A-06) and $9.17 \pm 0.10$ 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 $\sim 9–10$ Ma.

5.5 Discussion

5.5.1 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, ($11.69 \pm 0.01$ 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 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, the basaltic trachyandesite flow (unit Nfl) covered the basin at $\sim 11.7$ Ma. The onset of major accumulation in the basin, units Ns-3, −2 and −1, which consist of fluvial/alluvial, lacustrine and eolian sediments, started after $\sim 11.7$ Ma and before $\sim 10.5$ Ma, the ages of the oldest volcanic ashes obtained from the Pasto Ventura basin (sample P-Ash-01 in this study, and sample PVN75 in Schoenbohm and Carrapa, 2015) (Table 1). Basin strata continued to accumulate until at least $\sim 7.8$ 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 $7.77 \pm 0.21$ 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 tilted during basin deformation (Fig. 5). Therefore, its age of $11.7$ Ma must predate the onset of the late Miocene contractional deformation in the Pasto Ventura basin. Angular, internal unconformities indicate on-going syn-depositional 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 $9.51 \pm 0.10$ 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 NPV map (Fig. 4), an ash close to the core of a syncline is dated at $8.32 \pm 0.12$ Ma (sample PV10A-02), and Schoenbohm and Carrapa (2015) document tilted strata younger than $7.77 \pm 0.21$ 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 $11.69 \pm 0.01$ Ma and $9.99 \pm 0.12$ Ma and continued until after $7.77 \pm 0.21$ Ma in the Pasto Ventura basin. Maximum shortening occurred in the southern part of the basin, in which 47% shortening took place in a WSW-ENE direction (Fig. 6).

Across the Puna Plateau, contractional deformation ended and was succeeded by extensional deformation during the late Miocene and 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 (Eckelmann et al., 2013; Marrett and Emerman, 1992; Montero Lopez et al., 2010; Riller and Oncken, 2003; Riller et al., 2001; Risse et al., 2008; Schoenbohm and Strecker, 2009; Zhou et al., 2013). In the NPV map, Fault-1 cuts two mafic cinder cones of $0.76 \pm 0.16$ Ma and $0.47 \pm 0.03$ Ma 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 ~0.8–0.5 Ma. Other cross-cutting 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 to the Pasto Ventura basin) (Montero Lopez et al., 2010) and 4 Ma (south to the Pasto Ventura basin) (Schoenbohm and Strecker, 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 Lopez et al., 2010; Schoenbohm and Carrapa, 2015; Schoenbohm and Strecker, 2009), and certainly by 0.8 and 0.5 Ma based on constraints within the basin (Zhou et al., 2013).

5.5.2 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 above. The oldest post-mid-Miocene monogenetic mafic volcanism clusters around the Antofagasta-Antofalla region (~26° S) (Risse et al., 2008; Schoenbohm and Carrapa, 2015; Zhou et al., 2013) (Fig. 8A), with ages of up to 8.7 ± 0.04 Ma (Schoenbohm and Carrapa, 2015). The Pasto Ventura region is located in the southernmost 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 they lack geochemical evidence for subduction (described as intra-plate volcanism, Kay et al., 1994). Recent 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 seem to become progressively hotter within a short, ~1.7 Ma time window (Ducea et al., 2013). This model suggests that the foundered lower lithospheric blocks are likely small in size (~50km) (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 that are overlain by sedimentary units and are incorporated in upper crustal deformation (Fig. 2A and Fig. 5). They outcrop mostly along the southern margin of the Puna Plateau (Fig. 2A) (Roy et al., 2006; Schnurr, 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 implies 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 origins of Nfl lavas are not resolvable in this study, they are not readily explained by the same models invoked for the post-8.7 Ma monogenetic mafic volcanism from the same region (Drew et al., 2009; Ducea et al., 2013; Murray et al., 2015; Kay et al., 1994).

5.5.3 Deformation and dynamics of the southern Puna Plateau

Our data show that basin sedimentation and deformation occurred between 11.7 and 10.5 and <7.8 Ma in the Pasto Ventura region of the southern Puna Plateau. In this section we explore our data within the context of regional upper crustal deformation and geodynamic models for development of the Puna Plateau, and focus on revealing causes for the documented late Miocene upper-crustal shortening event in the southern Puna.

A first model notes the importance of a gradient in crustal shortening from the middle of the Altiplano-Puna Plateau at ~21°S to the southern tip of the Puna Plateau (Riller and Oncken, 2003; Riller et al., 2012). This strain field has resulted in the compressional reactivation of orogen-parallel thrusts and the formation of new NE-SW trending thrust faults that segment the plateau into rhomb-shaped structural basins (Riller and Oncken, 2003; Riller et al., 2012) (Fig. 1B; Fig. 9). This model predicts that shortening, and therefore the onset of basin formation, should propagate in time from N to S. The timing of onset of sedimentation in the Corque, Lipez, Atacama, Antofalla, Hombre Muerto, Santa Maria, Pipanaco and La Rioja basins supports this model (Riller and Oncken, 2003), as these basins decrease in age from Eocene in the north to Mio-Pliocene in the south. The timing of sediment accumulation in the Pasto Ventura basin determined in this study is consistent with this north to south progression, but the chronology of the onset of deformation is too tentative to allow a rigorous evolution of this model (Fig. 7B, C). Further, the formation of individual basins is likely more complex than envisaged by this model; studies of sedimentary facies and subsidence history for basins in the northern Puna Plateau have argued various origins including, but not limited to, foreland-type sedimentation and rapid surface subsidence related to lower lithospheric foundering (e.g., DeCelles, 2011; DeCelles et al., 2015; Schoenbohm and Carrapa, 2015), raising challenges in simply evaluating 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, 2002). In the northern Puna (~24°–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 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 re-interpretation of the Santa Barbara Subgroup, which was proposed to be a product of thermal subsidence following the demise of the Salta Rift (Marquillas et al., 2005). In this model, the Western Cordillera was deforming by early Paleogene (e.g., Arriagada et al., 2006; Cladouhos et al., 1994; Mpodozis et al., 2005; Kennan et al., 1995). Deformation migrated quickly across the plateau, but stalled in the Eastern Cordillera from ~25–19 Ma (e.g., Deeken et al., 2006). Deformation has been propagating east through the Subandes and Santa Barbara Ranges since ~15–8 Ma (e.g., Echavarria et al., 2003; Marrett et al., 1994; McQuarrie, 2002b; Pearson et al., 2013) (Fig. 3). 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 volcanism in this region, which began in early Miocene (Kay et al., 2010). After 11.7 Ma, the Pasto Ventura basin was filled with fluviolacustrine-eolian, mostly medium to fine-grained sediments and was accompanied by syn-depositional deformation. In a foreland system, wedge-top sediments are commonly associated with syn-deposition deformation due to the active thrust faulting in the deformation front, but 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 event in the Pasto Ventura region therefore needs an alternative explanation.

Other models emphasize out-of-sequence deformation, describing thick-skinned, post-Miocene deformation as non-systematic 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-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/deformation models agree on the important role of inversion of pre-existing Salta Rift structures (e.g., Carrapa et al., 2014; Cristallini et al., 1997; Deeken et al., 2006; Hongn et al., 2007; Insel 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 rifting 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 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., 2015, 2009; Elkins-Tanton, 2007; Göğüş and Pysklywec, 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 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, 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 (Ducea et al., 2013; Marret et al., 1994; Murray et al., 2015; Risse et al., 2008; Schoenbohm and Carrapa, 2015; Zhou et al., 2013) (Fig. 8A).

A small-scale, late Miocene lithospheric dripping event beneath the southern Puna Plateau may be responsible for the Late Miocene upper crustal 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 Strecker, 2009; Zhou et al., 2013). Evidence exists to support on-going 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 that presumably resulted from lithospheric foundering (Drew et al., 2009; Ducea et al., 2013; Kay et al., 1994; Murray et al., 2015) (Fig. 8) 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 north of the Pasto Ventura region, around ~26°S, which is evident 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 Galen, 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 (e.g., Allmendinger et al., 1997; DeCelles et al., 2011; Horton, 2005; Riller and Oncken, 2003; Sobel et al., 2003; Strecker et al., 2007) (Fig. 9), 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).

5.6 Conclusion

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 ~11.7 Ma (the age of the underlying basaltic trachyandesite flow, unit Nfl) and before ~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 ~7.8 Ma from unit Ns-1 and no outcrop of younger basin fill is observed in this region, suggesting unit Ns-1 is the last unit deposited in the Pasto Ventura basin. Several examples of syn-depositional deformation were identified within the strata of the Pasto Ventura basin. An angular unconformity formed within the Ns-3 mudstone-siltstone and two ashes below and above this unconformity yield identical ages (9.99 ± 0.12 Ma above the unconformity; 9.80 ± 0.10 Ma below the unconformity). The major basin-bounding, thrust fault (Fault-1 on the NPV map) truncates a 9.51 ± 0.10 Ma ash. These observations, together with growth strata within unit Ns-3, indicate that the Pasto Ventura basin was undergoing deformation around ~10–9 Ma. The youngest dipping strata are 7.77 ± 0.21 Ma (Schoenbohm and Carrapa, 2015), indicating deformation continued until after this time. Contractional deformation of the Pasto Ventura region ended 7.3–4 Ma, accompanied by up to 10%–47% upper crustal shortening, and was thereafter replaced by horizontal extension. The post-mid-Miocene Pasto Ventura basin is floored by part of the voluminous pre-late-Miocene mafic volcanics of the southern Puna Plateau. They are geochemically distinct from the well-studied post-mid-Miocene monogenetic mafic lavas from the same region, implying origin mechanism(s) other than lithospheric foundering.

Regional analysis of the Puna and southern Altiplano of the central Andes, including deformation documented from sedimentary basins and bedrock ranges, shows that the southeastern Puna Plateau was deformed out-of-sequence. The pattern of the deformation across
the central Andes can be explained by several models, including the development of an eastward propagating orogenic wedge, southward propagating deformation resulting from an N-S gradient in crustal shortening, inherited crustal heterogeneities from the Cretaceous Salta Rift, and lower lithospheric foundering. Several lines of evidence, including geophysical imaging, geochemistry and ages of volcanic lavas, and timing of surficial contraction and extension, suggest that the 11.7 Ma to recent deformation in the Pasto Ventura basin likely reflects formation and detachment of a lithospheric drip in the southern Puna Plateau during late Miocene time.
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*MSWD—mean square of weighted deviates.
†w.m.—weighted mean; 1 s.e.—one standard error.
TABLE 2. $^{40}$Ar/$^{39}$Ar AGES FOR MAFIC CINDER CONES AND LAVAS IN THE PASTO VENTURA REGION, NW ARGENTINA
(RISSE ET AL., 2008; ZHOU ET AL., 2013; THIS STUDY)

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<td>326.69</td>
<td>3.00</td>
<td>14.06</td>
<td>0.63</td>
<td>20.65</td>
<td>0.70</td>
</tr>
</tbody>
</table>

Notes:
1. Concentrations for major elements are listed in weight percent;
2. LOI: Loss on ignition, in weight percent;
3. Concentrations for trace elements are listed in ppm;
4. Mg# = [mol Mg/(mol Mg + mol Fe)]*100, assuming all Fe$^{2+}$;
5. Eu* is determined from interpolation of REE pattern between Sm and Tb (Kay et al., 1994).
Figure 1. (A) Overview of the Andes. Shaded areas indicate region above 3000 m in elevation, and dark shading denotes the modern surfaces that have a surface slope of less than 4° (e.g., modern basins), both derived from 90 m SRTM (Shuttle Radar Topography Mission) data. Black line encompasses the internally drained part of the Central Andean Plateau, from HydroSHEDS data (hydrological data and maps based on Shuttle Elevation Derivatives at multiple Scales). Reconstructed historical plate boundaries (dashed lines) are from McQuarrie (2002b). Black arrow shows the kinematics of the modern subducting Nazca plate (Marrett and Strecker, 2000). Inset: historical variations in plate velocity (compiled by Capitanio et al., 2011, from Pardo-Casas and Molnar, 1987; Sdrolias and Muller, 2006; Somoza, 1998). (B) Close-up of Central Andean Plateau indicating tectonomorphic zones (after Strecker et al., 2007). Major faults are shown for the Puna Plateau (from Riller and Oncken, 2003).
Figure 2. (A) Geological map of southern margin of the Puna Plateau (modified after Allmendinger et al., 1989; Martinez, 1995; Schnurr et al., 2006; Schoenbohm and Strecker, 2009; Sobel and Strecker, 2003; Zhou et al., 2013). (B) Quaternary and Neogene volcanism in the Pasto Ventura region, NW Argentina. Black polygons are Quaternary volcanic cinder cones and lava flows; geochronology data are from Risse et al. (2008), Zhou et al. (2013), and this study. Dark-green polygons are Neogene mafic flows (Martinez, 1995; Schnurr et al., 2006; Roy et al., 2006; this study). Numbers correspond to age data in Table 2.
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—Echavarria 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 and DeCelles, 2008; j—Schoenbohm and Carrapa, 2015; k—Deeken et al., 2006; l—Carrapa et al., 2005; m—Sobel and Strecker, 2003; n—Coutand et al., 2001; o—Carrapa et al., 2006; p—Schoenbohm et al., 2007; q—Löbens et al., 2013; r—Vezzoli et al., 2012; s—Pearson et al., 2013; t—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.
Figure 4. Field photos of map units. (A) Unit Ns-1, eolian sandstone. (B) Unit Ns-1 eolian sandstone, hydrothermally altered. Note that the eolian cross-bedding is preserved. (C) Unit Ns-2, gray fine to medium sandstone, with parallel lamination. (D) Unit Ns-3, mudstone. (E) Unit Ns-4, medium to coarse, cross-bedded red sandstone and siltstone with occasional cobble layers. (F) Paleosol within unit Ns-4, showing pedogenic carbonate nodules. (G) Unit Nfl, deformed basaltic trachyandesite. Note that the fractures are parallel to each other and are dipping to the right in the photo (dipping to the east in the field). Outcrop location: 26°46.39′S, 67°13.97′W. Sample PV11B-01 was sampled from this outcrop. (H) Unit M, the metamorphic bedrock.
Figure 5 (next page). 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) $^{40}$Ar/$^{39}$Ar geochronology result for sample PV11B-01, which was sampled from the deformed basaltic trachyandesite unit (unit Nfl).
Figure 6 (next page). (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.
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).
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 Nfl samples and post–mid-Miocene monogenetic mafic volcanics (compiled from Ducea et al., 2013; Drew et al., 2009; Kay et al., 1994, 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).
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.
CHAPTER 6

CONCLUDING REMARKS
6 Concluding Remarks

6.1 Summary of Findings

This thesis contributes to the understanding of regional deformation and landscape evolution, and resolves controversial orogenic and dynamic models for the southern Puna Plateau, central Andes.

As recorded by the deposition of Cenozoic strata, the onset of regional deformation related to the formation of the modern southern Puna Plateau was at the late Eocene. On the southeastern Puna Plateau, my work provides the first age constraint regarding the oldest Cenozoic strata exposed in the Pasto Ventura region, using sandstone modal composition and detrital zircon U-Pb and apatite fission-track data and suggesting basin accumulation during the late Eocene to early Oligocene (~38-28 Ma). Similar-aged strata are also identified on the southwestern Puna Plateau in the Antofagasta de la Sierra region. Combined with provenance data, including paleocurrent, sandstone modal composition, detrital zircon U-Pb and detrital apatite fission-track ages, this thesis argues that the onset of Cenozoic shortening at the southern Puna Plateau started during the late Eocene to early Oligocene (~38-28 Ma), driving the formation of a single regional basin. This basin was bounded by the active orogenic front located on the western Puna Plateau and from re-activated, inherited topographic relicts located on the eastern margin of the Puna Plateau. The depocenter was located to the western side, represented by the thick, >3-km strata found in the Sierra de la Antofagasta region.

This regional basin was disrupted and compartmentalized by uplift of ~N-S trending bedrock ranges starting as early as late Oligocene time. Bedrock apatite fission-track and (U-Th-Sm)/He samples from the eastern foot of the Sierra de Calalaste (northwestern Puna Plateau) argues for ~25-20 Ma onset of the most recent exhumation. Modeling of apatite fission-track and (U-Th-Sm)/He data shows that the Sierra Laguna Blanca (southeastern Puna Plateau) experienced exhumation at ~15-10 Ma, the youngest bedrock exhumation documented in the plateau region. Together, they show that the present-day basin-and-range morphology of the southern Puna
Plateau, similar to the thick-skinned Eastern Cordillera, was established no early than the middle Miocene.

This thesis also documents a local, late Miocene upper-crustal contraction in the southernmost Puna Plateau in detail. Structural mapping, deformation analysis and U-Pb and $^{40}$Ar/$^{39}$Ar geochronological data argue a major basin formation and deformation event during $\sim$12-8 Ma. Combined with existing and new geochemical and geochronological data, this contractual event seems to be well explained as an upper-crustal response for a mid-late Miocene lower lithospheric drip beneath the southern Puna Plateau.

Before the Cenozoic, the southern Puna Plateau region was home to multiphase exhumation and probable topographic development. Revealed by $\sim$250-300 Ma zircon (U-Th-Sm)/He ages from the Sierra Laguna Blanca and existing regional studies, the region of the present-day southeastern Puna Plateau was likely located on the highlands of the Pampean arch and was surrounded by sedimentary basins to the west and the south during the late Paleozoic. During the late Cretaceous, the southeastern Puna Plateau was affected by the Salta rift and was located on the rift shoulder, as evident from modeled apatite fission-track data that show significant cooling at $\sim$90-60 Ma. The Salta rift exerts a critical control in defining total exhumation for the southern central Andes, because the cooling related affected at the partial annealing zone of the apatite fission-track system.

6.2 Remaining Questions

As my thesis concludes, I believe that I have found more questions than answers. These questions stem from geological and geomorphologic evolution models of the Puna Plateau as well as methods used in studying orogenic systems.

This thesis documents an important dynamic shift in the Oligocene, from flexure-dominated contiguous foreland dynamics to flexure-limited broken foreland dynamic. The initiation of a regional basin in the late Eocene to early Oligocene is likely a result of a flexural response to the crustal load generated by in at the western Puna Plateau, despite of the presence of a deeply segmented crust introduced by earlier extensional tectonism. Thick-skinned deformation, reactivation of pre-existing faults and uplift of basement-cored ranges started in the late Oligocene, marking the cessation of lithospheric flexure.
The causes of such a dramatic shift need to be further explored. One important observation is that the change from contiguous foreland to broken foreland seems to coincide with the onset of arc volcanism, implying a genetic link. On the one hand, the observed correspondence may reflect the changing dip of the subducting Nazca plate. Low-angle, or flat, subduction has been associated with a volcanic lull in the arc region (Kay et al., 2009; Ramos, 2009), which may also “lock-up” pre-existing fault zones and lead to an overall flexural deformation of the lithosphere. On the other hand, the arc activity in the Cordillera-type system may be controlled by the cyclical behavior of the whole system, particularly the periodic built-up and detachment of dense crustal roots beneath the arc (e.g., DeCelles et al., 2009; Paterson and Ducea, 2015). During the built-up of dense roots, arc volcanic activities tend to be weak and this may also be a period of time favoring long-wavelength dynamics of the upper plate.

This thesis also provides timely insights and raises testable hypotheses for elevation history and causes of surface uplift for the Puna Plateau. Although paleoaltimetric estimates exist in the Altiplano and the northern Puna Plateaus (e.g., Canavan et al., 2013; Garzione et al., 2006; 2013), the most of the southern Puna Plateau still lacks data. My work argues that, during the late Eocene to early Oligocene, most of the southern Puna Plateau was a regional foreland-type basin that was sourced from both west and east, implying that the elevated region may have been confined to the western margin of the southern Puna Plateau (the western source region of the basin). At the same time, our along-strike reconstruction (Fig. 15 in Chapter Three) suggests that the northernmost Puna Plateau was occupied by a wedge-top basin, consistent with existing paleoelevation which suggest high, nearly modern elevation during the late Eocene for there (Canavan et al., 2013; Quade et al., 2015). Surface uplift in the Puna Plateau therefore appears to follow initiation of wedge-top deformation.

Testing and refining the model proposed in this thesis therefore is critical for resolving controversies regarding elevation history for the Puna Plateau. One test may be to confirm field evidence of the regional tearing fault predicted in our block model (Fig. 15 in Chapter Three). A potential target region is the identified, but poorly mapped and constrained ~NE-SW-trending faults beneath the Salar del Hombre Muerto, the right-lateral Acazoque strike-slip fault (e.g., Kraemer et al., 1999). Further, if such a relationship between the reconstructed regional deformation model and paleoelevation data truly holds, we may argue that the upper-crustal shortening was responsible for surface uplift in the Puna Plateau, as suggested for its adjacent
Eastern Cordillera (e.g., Carrapa et al., 2014). Such a dynamic cause for surface uplift contrasts with that for the Altiplano Plateau, which is thought to be the result a plateau-wide lower lithospheric foundering event (Garzione et al., 2008). Indeed, my work supports that lower lithospheric foundering beneath the Puna Plateau may take place at a small-scale, which likely contributes little to the surface uplift (Ducea et al., 2013; Schoenbohm and Carrapa, 2015).

My thesis also advances our understanding of the establishment of internal drainage in the Puna Plateau. Internal drainage serves to retain mass within the orogenic system and therefore construction or destruction of internal drainage may lead to fundamental changes of orogenic evolution pathways. Efforts to establish the chronology and growth pattern of internal drainage in the Puna Plateau have been underway for decades (e.g., Alonso et al., 1991; Strecker et al., 2007). Results from this thesis imply that the area of internal drainage may not evolve in one direction (i.e., not consistently growing larger through time), but requires further testing. The proposed late Eocene-early Oligocene regional basin was sourced from both the west and east, implying internal drainage. However, the southeastern Puna Plateau lacks depositional record between ~38-28 Ma to ~12 Ma, implying a non-deposition or even erosion, suggesting external drainage. Sedimentation resumed in the southeastern Puna Plateau at ~12 Ma, with gypsum layer, indicative of internal drainage, observed (Chapter Two). It is important to answer questions such as whether or not such an aerial fluctuation existed and if so what the causes were, in order to achieve better understanding for the complex linkages among tectonics, climate, and deposition-erosion.

This thesis highlights critical, less-understood aspects in performing low-temperature thermochronology in orogenic systems, including post-depositional resetting of detrital thermochronometers and anomalously old apatite (U-Th-Sm)/He ages caused by accumulated radiation damage.

As a primary goal, detrital thermochronological studies are used to distinguish exhumation history of source regions. Such a goal is usually realized by using a lag-time plot (e.g., Coutand et al., 2006), in which the time lag (the difference between the detrital mineral cooling age and the depositional age of hosting strata) reflects the exhumation of the source region. Concerns exist that the thermochronometers may be reset due to post-depositional burial heat and potential for burial heating is routinely tested by observing any younging trend of thermochronological
ages towards lower sections. This thesis documents evidence for another important source of post-depositional heating: magmatic intrusion. This catastrophic heating source may exert a vital influence on resetting the thermochronological signature inherited from the source region, but is much more difficult to characterize. As documented in my work, magmatic intrusion into the basin may take place randomly, not necessarily causing reset at the deepest-buried samples. A better understanding of the thermal effects of intrusion requires knowledge of basin strata architecture and the fluid circulation system, because the heat from magmatic intrusions does not simply affect nearby rocks, but rather drives hot fluid circulation that influences a much broader extend.

It is commonly known that the apatite fission-track system has a higher closure temperature than the apatite (U-Th-Sm)/He system and therefore one would expect an older apatite fission-track age from the same sample. However, studies have shown that accumulated radiation damage in apatite may lead to older, dispersed apatite (U-Th-Sm)/He ages (e.g., Flowers et al., 2009). This thesis includes some of the highest-eU (eU: a measure for radiation damage) apatites and finds that they may yield old ages that are even older than apatite fission-track ages from the same sample. We also show that, even when the measured apatite (U-Th-Sm)/He ages are older, they are in fact compatible with the thermal history obtained from apatite fission-track data when considering the radiation damage effect. This thesis therefore calls for caution in interpreting apatite (U-Th-Sm)/He ages without apatite fission-track ages from the same sample, especially for those with high eU amount or complicated subsurface temperature history.
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