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Title: Structural and thermochronologic constraints on kinematics, timing and shortening during inversion of the Salta rift into the Andean fold-thrust belt, northwest Argentina

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Key points

- Both thin- and thick-skinned structural styles present in Eastern Cordillera
- Apatite (U-Th)/He ages document eastward propagation of deformation
- Basement wedging kinematics in regional balanced cross-sections
Abstract

The central Andean retroarc fold-thrust belt varies significantly along strike, decreasing in total shortening and changing structural styles southward from Bolivia into Argentina. In addition to correlations with shallowing slab angles and thinning Paleozoic stratigraphy, inherited architecture of the Cretaceous Salta rift has been documented as a primary influence on Andean deformation in northwest Argentina. Detailed geological mapping, regional cross-sections, and apatite (U-Th)/He dating from the Eastern Cordillera at latitude 25-26˚S reveals a mixture of thin- and thick-skinned structural styles, inversion of normal faults, and eastward younging cooling ages. Regional structures consist of basement-involved reverse faults and steep to overturned fault-propagation folds. Inversion of east-dipping normal faults results in a back-thrust belt antithetic to the orogenic wedge. Inversion becomes more influential southward in the Tonco-Amblayo thrust belt where synrift facies thicken considerably and shortcut normal faults are preserved. To the north, Cenozoic sedimentary rocks are characterized by thin-skinned geometries. Apatite (U-Th)/He data document eastward younging cooling ages from ~12 Ma to ~6 Ma, suggesting that deformation propagated forward sequentially despite hindward vergence of the back-thrust belt. Kinematic models require structure at depth to accommodate changes in structural elevation, basement shortening, and forward propagation of deformation. Basement wedging, by either a single thrust sheet or a passive-roof duplex, provides a kinematically viable model. Shortening at this latitude in the Eastern Cordillera is ~30% (32 km over 107 km), ~5% greater than previous estimates. Revised total thrust belt shortening (~110 km) falls well short of solving shortening discrepancies suggested by crustal thickness.

Keywords: central Andes, balanced cross-section, thermochronology, tectonic inversion, Salta rift
1. Introduction

Cordilleran-style orogens result from the convergence of oceanic and continental tectonic plates. Prominent features of Cordilleran orogens include continental magmatism, forearc and retroarc shortening and extension, and backarc and retroarc foreland basins. In particular, fold-thrust belts form along the margins of mountain belts in contractional regimes to accommodate crustal shortening [Davis et al., 1983]. Common properties of fold-thrust belts include a regional décollement or detachment that dips toward the hinterland, shortening in the material above the décollement, and a wedge shape of the deformed material tapering toward the foreland [Chapple, 1978; Davis et al., 1983]. Fold-thrust belts develop wedge-shaped geometry, tapering toward the undeformed foreland [e.g., Price, 1981; Fuentes et al., 2012], and continue to grow self-similarly as new material accretes into the orogenic wedge [Davis et al., 1983]. Fold-thrust belts are generally “thin-skinned” involving mainly sedimentary cover rock through low-angle thrusting; however, “thick-skinned” basement-involved high-angle thrusting is common.

Thin-skinned deformation is characterized by low-angle thrust faults, shallow décollements, ramp-flat geometries, duplex structures and sequential propagation of deformation [e.g., Bally et al., 1966; Dahlstrom, 1970; Davis et al., 1983]. Cordilleran-style fold-thrust belts typically deform in a thin-skinned style, with prominent examples including the Canadian Rockies; western Utah, Wyoming, and Montana; and the Bolivian Andes [e.g., Dahlstrom, 1970; Price, 1981; Armstrong, 1968; Fuentes et al., 2012; McQuarrie, 2002]. However, notable exceptions such as the Laramide structures in the western interior USA and the Sierras Pampeanas in Argentina are characterized by thick-skinned deformation, with steeper reverse
faults cutting crystalline basement, deeper décollements and erratic, out-of-sequence strain propagation [e.g., Jordan and Allmendinger, 1986; Brown, 1988; Stone, 1993; Erslev, 1993; Jordan, 1995]. Why thin or thick-skinned deformation prevails in any given region is still subject to study. In Cordilleran-style orogens, the presence of a thick sedimentary succession and the angle of the down-going slab are thought to provide first order controls on deformation style.

Thin-skinned deformation typically occurs where the fold-thrust belt can propagate into thick sedimentary packages comprising alternating zones of weak and strong rocks that can form regional detachments and thrust ramps, respectively. Thick passive margin sequences that exhibit lithologically variable stratigraphy and a hinterland-dipping basement-cover interface to activate as a regional décollement are prime candidates for thin-skinned fold-thrust belts [e.g., Bally et al., 1966; Davis et al., 1983]. When stratigraphy is thin or non-existent, shortening must then be accommodated in massive, mechanically strong basement rock. Without horizontal layers of weakness, shortening is accommodated in small increments through basement block uplifts along steep reverse faults [e.g., Jordan, 1995]. As this is an inefficient means to achieve large amounts of shortening, thick-skinned deformation will preferentially reactivate pre-existing basement structures. Therefore, inherited basement heterogeneities perhaps play a large role in determining the style, timing, and propagation direction of thick-skinned thrust belts [Allmendinger et al., 1983; Schmidt et al., 1995].

Subducting slab angle is also considered to provide a primary control on fold-thrust deformation style [e.g., Jordan et al., 1983; Gutscher et al., 2000]. In normally subducting sections, the primary forces acting on the fold-thrust wedge are a horizontal tectonic force from the convergent margin plate interface and gravity, whereas a regional décollement provides a low-friction interface for the fold-thrust wedge to deform over [Davis et al., 1983]. This free-
body configuration approximates the “snow plow” analogy used for fold-thrust wedge dynamics and critical taper analysis in thin-skinned fold-thrust belts [Davis et al., 1983; Dahlen, 1990]. It has been proposed that thick-skinned fold-thrust belts, which generally do not behave as predicted by a critically tapered wedge analysis, are a result of flat-slab subduction [Livaccarri et al., 1981; Henderson et al., 1984; Barth and Schneiderman, 1996; Saleeby, 2003; Liu et al., 2010]. In this scenario, the subducting flat slab couples with the base of the overriding plate over a broad region inboard from the plate interface. This adds an additional vertical component of tectonic force from below and disrupts the free-body configuration described in the “snow plow” scenario. The thrust belt may behave more erratically in response to coupled stresses at depth. Shallow subduction of an oceanic plateau and coupling and subsequent detachment of overriding and down-going slabs at depth has been proposed as a cause of Laramide structures during the Cretaceous-Eocene [Livaccari et al., 1981; Henderson et al., 1984; Barth and Schneiderman, 1996; Saleeby, 2003; Liu et al., 2010].

The South American central Andes are considered the archetypal example of an active Cordilleran-style orogenic belt (Figure 1) [Coney and Evenchick, 1994; Ramos, 2009]. They include diagnostic characteristics such as an active magmatic arc, a retroarc fold-thrust belt, and a flexurally controlled retroarc foreland basin system. However, the Andes vary significantly along strike [Allmendinger et al., 1988; Jordan et al., 1983; Isacks, 1988; Kley and Monaldi, 1998]. North and south of the central Andes, flat slab segments have shut off continental magmatism, while the retroarc is characterized by thick-skinned basement block uplifts and a broken foreland basin. Major along-strike changes in structural style coincide with a decrease in shortening away from the central Andes [Kley and Monaldi, 1998]. Although thin-skinned deformation in the Bolivian Andes [Dunn et al., 1995; McQuarrie, 2002] correlates with normal
subduction while thick-skinned deformation in the Sierras Pampeanas corresponds to flat subduction [Isacks, 1988; Jordan and Allmendinger, 1986; Cahill and Isacks, 1992], the correlation between the transition in structural style and the transition in slab angle is imperfect [Ramos, 2002]. Along-strike variations have also been explained by changes in pre-Andean stratigraphy and heterogeneities in the pre-Andean South American plate [Strecker et al., 1989; Grier et al., 1991; Salfity and Marquillas, 1994; Mon and Salfity, 1995; Allmendinger and Gubbels, 1996; Kley et al., 1999; Kley and Monaldi, 2002]. Additionally, total shortening decreases dramatically south of Bolivia into northern Argentina [Kley and Monaldi, 1998; McQuarrie, 2002], despite consistent relative convergence velocities between the Nazca and South American plates [Oncken et al., 2006]. There is also a significant shortening discrepancy in the transition between the Bolivian Andes and the Sierras Pampeanas, as total measured shortening in northwest Argentina is unable to account for observed crustal thicknesses [Kley and Monaldi, 1998].

The Eastern Cordillera and Santa Barbara thrust systems of northwest Argentina exhibit the transition in structural styles due to contrasting subduction dynamics [Jordan et al., 1983]. To the north, the Bolivian Andes are characterized by thin-skinned deformation attributed to simple shear through a thick Paleozoic section and normal subduction (30°) [Cahill and Isacks, 1992]. To the south, the Sierras Pampeanas are characterized by thick-skinned deformation associated with pure shear through basement rocks and flat slab subduction. However, thrust belt evolution in northwest Argentina is likely controlled by tectonic inversion of the Cretaceous Salta rift [e.g., Grier et al., 1991]. Factors such as inversion, flat-slab subduction, and thinner pre-Cretaceous stratigraphy give rise to a basement-involved fold-thrust belt composed of east-dipping back thrusts. Inversion of the Cretaceous Salta rift has been documented from surface
data in the Eastern Cordillera and Santa Barbara Ranges [Grier et al., 1991; Kley and Monaldi, 2002; Kley et al., 2005; Carrera et al., 2006; Pearson et al., 2013], and from sparse seismic data in the Eastern Cordillera [Kley et al., 2005] and Santa Barbara Ranges [Cristallini et al., 1997]. Reactivation of the Salta rift structures and overall rift architecture may prove influential on the structural style and kinematic evolution of the Andean thrust belt through the region.

This study focuses on the structural evolution of the Eastern Cordillera in northwest Argentina between 25°S and 26°S. The kinematics of inversion of the Salta rift remain poorly understood, including mechanisms for basement-involved faulting and folding and the existence of a regional detachment. Also, some authors argue for eastward propagation of deformation [Carrapa et al., 2011] whereas others argue for more erratic, out-of-sequence thick-skinned deformation [Hain et al., 2011; Strecker et al., 2011; Pearson et al., 2013]. Current shortening estimates [Grier et al., 1991; Carrera and Muñoz, 2013] fall well short of required shortening totals to account for crustal thickness [Kley and Monaldi, 1998]. A recent study north of the study area (24°S-25°S) suggests that there may be more shortening (~45%) in the Eastern Cordillera than previously documented [Pearson et al., 2013]. Documenting significantly more shortening in the region than current estimates (~24-25%) [Grier et al., 1991; Carrera and Muñoz, 2013] would perhaps fully explain the crustal thickness. Alternatively, if further documentation does not yield significant increases in shortening estimates, then other processes (e.g., magmatic addition, tectonic underplating, lower crustal ductile flow) must account for crustal thickening in northwest Argentina. These key questions are addressed through detailed geologic mapping, (U-Th)/He thermochronological analyses, and kinematically viable regional balanced cross-sections. Field mapping targeted the Eastern Cordillera thrust belt between 25°-26°S and 65°-66°W (Figure 2). Regional balanced cross-sections were constructed along an E-W
transect at 25°25’S. Samples for apatite (U-Th)/He thermochronology were collected along the same transect. Field observations and balanced cross-sections provide shortening estimates and a viable kinematic model for inversion of the Salta rift and its incorporation into the Andean orogenic belt. Apatite (U-Th)/He dating provides insights on the timing and sequence of erosion related to deformation through the region.

2. Geologic Background

2.1 Regional Setting

The central Andes are commonly divided into four longitudinal tectonomorphic zones: the Western Cordillera, Altiplano-Puna Plateau, Eastern Cordillera and the frontal Subandean and Santa Barbara Ranges (Figure 1) [Allmendinger et al., 1997; Strecker et al., 2007]. The Western Cordillera has been the active magmatic arc since early Miocene time and is composed of dacitic-andesitic stratovolcanoes and ignimbrites [Kay and Coira, 2009]. The Altiplano Plateau in southern Peru and Bolivia reaches a regional elevation of ~3800 km [Isacks, 1988; Masek et al., 1994] and is dominated by large Cenozoic basins. The Puna Plateau in northwest Argentina is ~400 meters higher than the Altiplano Plateau and more rugged. Both regions are mostly internally drained. The Puna Plateau is composed of Precambrian and Paleozoic sedimentary, igneous, and low-grade metamorphic rocks, and Cenozoic sedimentary and volcanic rocks and stratovolcanoes that extend east from the volcanic arc. In Bolivia, the Eastern Cordillera and Subandean Ranges comprise a thin-skinned retroarc fold-thrust belt composed of
Proterozoic-Cenozoic sedimentary rocks detached along a regional décollement [Roeder, 1988; Kley, 1996; McQuarrie and DeCelles, 2001; McQuarrie, 2002; Echavarria et al., 2003]. Shortening estimates are 300-330 km [McQuarrie, 2002]. In northwest Argentina, the Eastern Cordillera is characterized by high-angle reverse faults that involve Precambrian-Cambrian crystalline and metamorphic rocks and Cretaceous rift sedimentary rocks (Figure 2) [Grier et al., 1991; Kley and Monaldi, 2002; Carrera et al., 2006; Pearson et al., 2013]. The Santa Barbara system is the frontal thrust belt in northwest Argentina, and is characterized by steep, west-verging reverse faults with ~10-20 km wide blocks of Paleozoic-Cenozoic sedimentary rocks in their hanging walls [Cristallini et al., 1997; Kley and Monaldi, 2002]. The Santa Barbara system is controlled by reactivated normal faults associated with the Cretaceous Salta rift, a complex of extensional basins, beneath the modern foreland basin [Salhy and Marquillas, 1994]. Shortening estimates vary between ~70 km [Kley and Monaldi, 1998] and ~142 km [Pearson et al., 2013]. To the south, thick-skinned basement uplifts of the Sierras Pampeanas have shortening estimates between 30 and 120 km [Kley et al., 1999].

2.2 Tectonic History of NW Argentina

During the Neoproterozoic-Cambrian, northwest Argentina was the stable passive margin of Gondwana along the proto-Pacific ocean [Jezek and Miller, 1985]. Sedimentary rocks deposited on this passive margin and subsequently slightly metamorphosed during Paleozoic orogenic events make up a large portion of the basement rocks currently cropping out in northwest Argentina. During Paleozoic time, the proto-South American margin experienced a
series of orogenic events. This is recorded in the Bolivian Andes as a >10 km package of sedimentary rocks deposited in a back-arc setting from the Cambrian to the Carboniferous [Sempere, 1995]. In northwest Argentina, Paleozoic sedimentary rocks are much more limited in thickness and spatial extent [Starck, 1995; Egenhoff, 2007]. In the study area, between the Calchaquí and Lerma valleys, there are no Paleozoic rocks. Instead Cretaceous-Paleogene rift-related sedimentary rocks as thick as 5.5 km (Salta Group) rest unconformably on metamorphosed and highly deformed Neoproterozoic-Cambrian rocks [Salfity and Marquillas, 1994]. Cretaceous sedimentary deposition is related to widespread, low-magnitude Mesozoic extension throughout western South America. Potential causes of Mesozoic extension include orogenic collapse [e.g., Kay et al., 1989], back-arc extension [e.g., Welsink et al., 1995] and failed riftting related to the opening of the Atlantic Ocean [e.g., Grier et al., 1991].

The Andean orogeny began during late Cretaceous-Eocene time in northern Chile when, due to opening of the South Atlantic Ocean, the South American plate began overriding the subducting Nazca plate [Coney and Evenchick, 1994; Sempere et al., 1997; Arriagada et al., 2006; Jordan et al., 2007]. Overall, the retroarc orogenic wedge has propagated eastward relative to South America throughout the Cenozoic. In northwest Argentina, deformation was located in the Puna Plateau and western Eastern Cordillera during Eocene-early Miocene time [Deeken et al., 2006; Carrapa and DeCelles, 2008; Carrapa et al., 2011; Insel et al., 2012; Pearson et al., 2012]. Deformation either propagated sequentially [Carrapa et al., 2011] or jumped sporadically [Pearson et al., 2013] through the Eastern Cordillera and into the Santa Barbara Ranges during Mio-Pliocene time. Likewise, the foreland basin has been documented to have migrated as a flexural basin in front of an eastward prograding thrust belt during Paleocene-Oligocene time [DeCelles et al., 2011], before transitioning to a broken foreland system during the mid-Miocene.
in response to erratic growth of the orogenic wedge with deposition accommodated in intermontane basins [Hain et al., 2011; Strecker et al., 2011].

2.3 Balanced Cross-sections and Shortening Estimates

Inversion of the Salta rift has been well-documented in northwest Argentina [Grier et al., 1991; Kley and Monaldi, 2002; Kley et al., 2005; Carrera et al., 2006; Pearson et al., 2013]. Grier et al. [1991] produced the first balanced section through the Andes at latitude ~25°30’S. It extends through both the Eastern Cordillera and the Santa Barbara Ranges. This section was constructed on the hypothesis that the failed Salta rift had a detachment zone at ~18 km depth that was reactivated as a basal thrust décollement, and that rift-controlling normal faults were reactivated in their entirety, leading to steeply dipping, listric reverse faults within the basement. Carrera et al. [2006] documented inversion features in the Calcahquí valley region through detailed mapping, outcrop photographs, and a series of km-scale cross-sections. Features of inversion include changes in synrift thicknesses and preserved extensional faults in the hanging walls of reverse faults that short cut them. Carrera and Muñoz [2013] constructed another regional balanced cross-section at ~25°30’S that documents tight fold geometries near the surface. It extends through the Eastern Cordillera from the Cachi Range west of Calchaquí valley to the Metán Range east of Lerma valley. Flexural-slip restoration was used on the sedimentary cover, and the basement blocks were restored by maintaining constant area because of complex internal deformation and absence of consistent stratigraphic references. Shortening in this cross-section is accommodated inhomogeneously across each basement block, ranging from as high as ~40% to as low as 10% in various basement blocks [Carrera and Muñoz, 2013]. Grier et al.,
documented 25% shortening (~70 km over ~280 km) in the Eastern Cordillera and Santa Barbara Ranges at this latitude, whereas Carrera and Muñoz [2013] found 24% shortening (44.5 km over 185.5 km) through the Eastern Cordillera.

Regional balanced cross-sections have been constructed just north of the study region as well. Kley and Monaldi [2002] and Pearson et al. [2013] constructed regional balanced cross-sections through the Eastern Cordillera and the Santa Barbara Ranges. Kley and Monaldi [2002] produced balanced cross-sections through inverted rift sectors of the Santa Barbara Ranges that were then incorporated into a balanced regional cross-section constructed by Pearson et al. [2013] at latitude ~24.75°S. The Eastern Cordillera at this latitude involves strain-hardened rocks that behave as mechanical basement, commonly deforming as pop-up structures and fault-propagation folds, so modeling was performed using inclined shear oriented antithetic to faults [Pearson et al., 2013]. Cretaceous rifting is more restricted to the Santa Barbara Ranges at this latitude. Kley and Monaldi [2002] restored sections through the Santa Barbara Ranges using auxiliary lines in the basement to allow line length balancing. Pearson et al. [2013] found that the Eastern Cordillera shortened 45% (95 km). Furthermore, they combined shortening estimates from the Puna Plateau (26 km) [Coutand et al., 2001] and the Santa Barbara Ranges (21 km) [Kley and Monaldi, 2002] to give a total retroarc shortening estimate of ~142 km (26%). Despite being an increase over previous retroarc shortening estimates, this total shortening is still much less than found in southern Bolivia. Additionally, revised shortening estimates fall short of the required amount to accommodate observed crustal thickness in northwest Argentina [Isacks, 1988; Kley and Monaldi, 1998]. Pearson et al., [2013] contends that further shortening may be found in the Puna Plateau and the Cretaceous-Paleocene forearc region of northern Chile, in
which case, observed crustal thickness in northwest Argentina may be accounted for in crustal shortening alone.

2.4 Exhumation History

Low temperature thermochronologic constraints on the rates and timing of deformation-related exhumation spanning the central Andes to the Sierras Pampeanas have been the subject of many recent studies [e.g., Deeken et al., 2006; Ege et al., 2007; Barnes et al., 2008; Carrapa et al., 2011; Barnes et al., 2012; Pearson et al., 2012; Bense et al., 2013; Carrapa et al., 2013; Löbens et al., 2013a, 2013b; Pearson et al., 2013]. Rock uplift, surface uplift, and exhumation provide quantitative measurements on the evolution and history of the Andean orogenic belt. Low temperature thermochronometers record a cooling age when a mineral passes through its closure temperature on its way to the surface during exhumation. This provides important constraints on erosion related to the generation of topography and the evolution of mountain belts. In the Andes, importance is placed on linking exhumation ages along strike in an attempt to characterize timing, style, and propagation of deformation along the orogenic belt. As shortening and crustal thickening are accommodated by disparate thrust systems in the Bolivian Andes and the Sierras Pampeanas, low temperature thermochronology can help determine differences and similarities in their style and evolution along strike.

2.4.1 Exhumation in the Bolivian Andes (15-22°S)
In the northern Bolivian Andes at 15-17°S, exhumation occurred in the Eastern Cordillera from ~50-30 Ma, the Interandean Zone from ~25-18 Ma, and started in the Subandes between 15 and 8 Ma [Barnes et al., 2006; McQuarrie et al., 2008]. Fission track data between 17-19°S record exhumation from ~50-45 Ma in the Eastern Cordillera, from ~18-6 Ma in the Interandean zone, and from ~7-3 Ma in the Subandes [Barnes et al., 2012]. Barnes et al. [2008] presented apatite and zircon fission track ages across ~19.5°S in Bolivia, which he interpreted to show that initial exhumation in the Eastern Cordillera began in the late Eocene to early Oligocene (36-27 Ma) and continued into the late Oligocene to early Miocene (25-19 Ma). Interandean zone exhumation began from 22-19 Ma, followed by another pulse of exhumation from 16 to 11 Ma in the Eastern Cordillera. Exhumation then began in the westernmost Subandes sometime between 20 and 8 Ma, and propagated eastward across the Subandes during the late Mio-Pliocene (8-2 Ma) [Barnes et al., 2008]. Scheuber et al. [2006] and Ege et al. [2007] have similar results as Barnes et al. [2008] but at ~21°S. Apatite fission track ages of 40 and 36 Ma in the central Eastern Cordillera record initial Andean thrusting. Apparent cooling ages record exhumation in the Altiplano and the Eastern Cordillera between 37-20 Ma with the majority of ages around 30 Ma [Scheuber et al., 2006]. According to Ege et al. [2007], exhumation spread into the southern Altiplano Plateau between 33-27 Ma and continued until shortening terminated by 11-7 Ma. From 17 Ma, exhumation progressed into the Interandean-Subandean zones, and starting ~8 Ma, exhumation was confined to the frontal Subandes [Scheuber et al., 2006]. Exhumation rates from Barnes et al. [2008] range from ~0.1-0.2 mm/a in the Eastern Cordillera, to ~0.1-0.6 mm/a in the Interandean zone, to ~0.1-0.4 mm/a in the eastern Subandes. Deformation was distributed throughout the Altiplano and Eastern Cordillera from ~40-20 Ma; however, following initiation of exhumation in the Interandean zone at ~20 Ma and
contemporaneous cessation of deformation in the Eastern Cordillera, deformation propagated mostly eastward into the Subandes [Barnes et al., 2008]. Collectively, in the Bolivian Andes (15-22°S), deformation and initial exhumation migrated eastward across the Eastern Cordillera from ~50-30 Ma, the Interandean zone from ~25-18 Ma, and started in the Subandes between ~15 and ~8 Ma, lasting until ~3 Ma [Barnes et al., 2012].

2.4.2 Exhumation in NW Argentina (23-27°S)

_Insel et al. [2012]_ presented \(^{40}\text{Ar}/^{39}\text{Ar}\) ages that show a wide variety of ages in the Puna Plateau. The western Puna Plateau basement has been cooler than 200°C since the Devonian (~400 Ma). The southeastern Puna Plateau, on the other hand, cooled to temperatures under 200°C by ~120 Ma associated with active extension of the Cretaceous Salta rift due to tectonic faulting and lithospheric thinning. The northeastern Puna Plateau experienced higher cooling rates between 78 and 55 Ma related to thermal subsidence during the postrift stage of the Salta rift and/or shortening-related flexural subsidence. In northwest Argentina, deformation migrated through the Puna Plateau, Eastern Cordillera and Santa Barbara Ranges in the Cenozoic. Accelerated cooling and deformation during the Eocene was focused within a narrow zone along the eastern Puna/Eastern Cordillera transition [Insel et al., 2012].

_Deeken et al. [2006]_ have apatite fission track ages in the Puna Plateau and adjacent Eastern Cordillera basement ranges (Cachi Ranges, Luracatao Peaks) at ~25°S. The AFT ages show Eocene-Oligocene exhumation along the eastern margin of the Puna Plateau followed by reburial under Andean foreland basin sediments from 30-25 Ma prior to Andean deformation which commenced at 22.5-21 Ma and continued through ~13 Ma. Zircon (U-Th)/He ages from
the Cachi Range at the same latitude yielded one sample with consistent (reset) mid-Miocene ages (~15 Ma) [Pearson et al., 2012]. Additional samples were not reset. Pearson et al. [2012] estimated that the Cachi Range underwent ~8-10 km of exhumation locally. Pearson et al. [2013] presented apatite (U-Th)/He data in the Eastern Cordillera at ~25°S that overlap with previous data in the Cachi Range and extends ~75 km east to the Mojotoro Range. These data show significant exhumation occurred ~15 Ma in the Cachi Range, then jumped ~75 km east at ~12-10 Ma to the Mojotoro Range by bypassing a large Cretaceous horst block [Pearson et al., 2013]. Exhumation then propagated back west into this horst block from 12.8 to 4.4 Ma [Pearson et al., 2013]. During the Pliocene, deformation jumped again >100 km eastward to the Santa Barbara Ranges.

A little to the south at 25-26°S, Carrapa et al. [2011] presented apatite (U-Th)/He data through the Eastern Cordillera that records sequential eastward younging of cooling ages from ~14-3 Ma and a rate of propagation of exhumation of ~8.3 mm/yr. These eastward younging ages are directly along strike of the westward younging ages from Pearson et al. [2013], suggesting that there is potentially a prominent, abrupt change in structure at 25.25°S. Carrapa et al. [2013] sampled some of the highest basement ranges from 25-27°S for low temperature thermochronology. Apatite fission track ages document ~28 Ma cooling at the top of the Cachi Ranges, which when combined with an age of 15.0 ± 1.7 Ma at 3700 m [Deeken et al., 2006] results in a minimum exhumation rate of 0.2mm/yr [Carrapa et al., 2013]. However, some of the highest ranges in the Eastern Cordillera (Sierra de Quilmes) preserve Cretaceous AFT ages and Cretaceous-Cenozoic He ages indicative of limited Cenozoic exhumation. Spatial patterns of these Cretaceous ages correlate with paleorift topographic highs, whereas Cenozoic ages are located in the paleorift hanging-wall basins [Carrapa et al., 2013]. Coutand et al. [2006]
presented detrital AFT ages from the Miocene-Pliocene portion of the Angastaco basin that contain a population with extremely short lag times (0-4 Ma) and cooling ages between 15.6 and 6.6 Ma on the eastern border of the Puna Plateau at 25-26°S. Mortimer et al. [2007] documented exhumation at ~6 Ma in the southwestern Sierra de Quilmes at ~26.5 S based on modeling of apatite fission track data. To the southwest, nearby Chango Real, Coutand et al. [2001] documented ~38-29 Ma AFT ages as evidence for exhumation in the Puna Plateau in the late Eocene and adjacent EC in the late Eocene-early Oligocene at ~27°S.

2.4.3 Exhumation in the Sierras Pampeanas (27-33°S)

The Sierras Pampeanas have a complex thermal history, with low-temperature thermochronometers recording Paleozoic, Mesozoic, and Cenozoic cooling events. Apatite fission track data in the northern Sierras Pampeanas (~27°S) document exhumation in the Calchaquies Range during the Cretaceous, which resulted in peneplain development until present [Sobel and Strecker; 2003]. The Aconquija Range to the south, however, cooled rapidly between 5.5 and 4.5 Ma. Löbens et al. [2013a] present a thermal history with multiple thermochronometric proxies (K-Ar, FT, and He dating) in the northern Sierras Pampeanas (~27°S). Initial exhumation occurred in the Cumbres Calchaquies between the Devonian and Carboniferous during the late stages of the Famatinian Orogeny. Exhumation was limited during the Mesozoic aside from some Cretaceous-early Cenozoic exhumation due to uplift of a rift-shoulder. During the Cenozoic, early uplift of the Puna Plateau triggered erosion and burial reheating to the east in the Sierras Pampeanas. Significant exhumation began ~9 Ma, related to Andean deformation in the Sierra de Aconquija [Coughlin et al., 1998; Löbens et al., 2013a].
Detrital and basement AFT thermochronology from the Fiambalá Basin documents a detrital population sourced from the Puna Plateau that is middle Miocene in age, and that the Puna Plateau expressed high relief by the late Miocene [Carrapa et al., 2006]. A recent study by Safipour et al. (accepted) shows apatite (U-Th)/He and AFT data from the Fiambalá basin displaying an eastward younging of cooling from 22 and 2 Ma, which correlates with the timing of migration of exhumation and deformation in the Eastern Cordillera of NW Argentina.

Bense et al. [2013] presented new and compiled low-temperature thermochronologic data [e.g., Bense, 2013; Löbens et al., 2013a] for the Sierras Pampeanas (27-33°S) including zircon and apatite (U-Th)/He dating and apatite fission track dating. Oldest exhumation ages show that the Sierras Pampeanas initiated cooling during Carboniferous time. In the southern Sierras Pampeanas (30-33°S), cooling was pronounced during Permo-Triassic time and younged westward. Cooling to near surface temperatures in the southern Sierras Pampeanas occurred between the Late Cretaceous and the Paleogene. In the northern Sierras Pampeanas (~27°S), however, final cooling occurred during the Miocene [Bense et al., 2013; Löbens et al., 2013a]. Apatite fission track and zircon and apatite (U-Th)/He ages in the western Sierras Pampeanas at 31-32°S documents an older thermal history [Löbens et al., 2013b]. Evolution of the Sierra de Pie de Palo began during the Late Paleozoic and continued through the Mesozoic due to tectonically-driven erosion. Modern topography began during the Paleogene during early Andean deformation ~60 Ma and propagated westward [Löbens et al., 2013].

2.5 Non-thermochronologic models for deformation through the Eastern Cordillera
Timing and propagation of deformation through the Eastern Cordillera of northwest Argentina has been inferred through methods other than low-temperature thermochronology as well. Carrera and Muñoz [2008] used growth strata and unconformities to demonstrate that deformation migrated towards the east despite westward vergence on most of the structures in the region. From the middle Eocene-Oligocene, the Calchaquí valley region was part of a continuous foreland basin, east of the thrust front. During the middle Miocene (15-10 Ma), Salta rift depocenters began to be inverted in the Eastern Cordillera. However, a forward thrusting sequence occurred despite the forelandward (eastward) dip of the reverse faults, resulting in break-back to synchronous faulting [Carrera and Muñoz, 2008]. Eastward propagation accelerated during the late Miocene (9-5 Ma), resulting in near synchronous inversion of the Salta rift throughout the Eastern Cordillera and giving rise to a broken foreland basin. Most thrusting occurred on the Payogasta, Cerro Negro and Calchaquí Faults to the east of Calchaquí valley [Carrera and Muñoz, 2008]. During Pliocene-Pleistocene time (5-1.5 Ma), deformation migrated eastward to the Lerma valley and Metán regions. Then, during Pleistocene-Holocene time (1.5-0 Ma), thrust faults throughout the Eastern Cordillera were reactivated regardless of location, as shown by growth geometries [Carrera and Muñoz, 2008].

Hain et al. [2011] used sediment architecture and facies associations to define a three-phase (~10, ~5, and <2 Ma), east-directed, but nonsystematic propagation of deformation and foreland fragmentation. Based on provenance signatures and evolution of fluvial deposits, the foreland of northwest Argentina was, initially, a continuous flexural foreland basin, but was disrupted and broken into intermontane basins during the mid- to late Miocene [Hain et al., 2011]. In this model, deformation was located in the Metán Range (the Eastern Cordillera-Santa Barbara Ranges border) east of Lerma valley at ~10 Ma before back stepping into the central
Eastern Cordilleran Ranges (the ranges in-between Calchaquí and Lerma valleys) at ~5 Ma [Hain et al., 2011]. Since ~2 Ma, deformation has been partitioned throughout the entire foreland.

In DeCelles et al. [2011], previously interpreted Paleocene-Eocene post-rift sag deposits (Santa Barbara Subgroup) are reinterpreted as early foreland basin deposits (distal back-bulge to fore-bulge). This extends the initiation of foreland deposition and eastward migration of the flexural foreland basin system into Paleocene time. Furthermore, the reinterpreted Cenozoic foreland basin record in northern Argentina is similar to its counterpart in Bolivia [DeCelles and Horton, 2003], suggesting that the foreland basins of Bolivia and northern Argentina have behaved similarly throughout the Cenozoic [DeCelles et al., 2011]. In terms of migration of deformation, a regional eastward migrating flexural foreland basin suggests an eastward migrating thrust front throughout the Cenozoic.

3. Stratigraphic framework

3.1 Puncoviscana Formation

The oldest exposed rock unit in the mapping area is the Puncoviscana Formation. The Puncoviscana Formation, in the study are, consists of variably metamorphosed, thin-bedded siltstone, shale, argillite, slate, and thin- to medium-bedded, fine-grained quartzose sandstone. Low-grade metamorphic facies are composed of slate and quartzite. Deposited during the Neoproterozoic to middle Cambrian, the Puncoviscana Formation represents submarine fan and
turbidite deposits that formed part of a passive margin sedimentary wedge along the proto-Pacific margin of Gondwana [Jezek and Miller, 1985; Pearson et al., 2012]. The Puncoviscana Formation was metamorphosed and shortened through thin-skinned folding and thrusting in the Late Cambrian-Ordovician [Willner et al., 1987], likely as part of the Ocloyic orogeny [Mon and Salfity, 1995; Starck, 1995].

3.2 Salta Group

Prior to initiation of foreland basin deposition in northwest Argentina, deposition was restricted to a network of spatially-limited, interconnected sub-basins that developed on top the Puncoviscana Formation, in the hanging-walls of major normal faults related to extension of the Cretaceous Salta rift [Grier et al., 1991; Salfity and Marquillas, 1994]. These rift-related sub-basins, collectively known as the Salta basin, grouped around the Salta-Jujuy high. The mapping area in this study overlays the northwestern extent of the Alemanía sub-basin, the southwesternmost sub-basin of the Salta rift network, as defined in Salfity and Marquillas [1994]. Sedimentological descriptions presented here are primarily from fieldwork performed for this study, and characterize the rocks in the northwest Alemanía sub-basin. This includes Cretaceous-Eocene sedimentary rocks deposited in the region in and around Tonco and Amblayo valleys (Figure 3). The Cretaceous-Eocene Salta Group is up to 4.5 kilometers thick in the Alemanía sub-basin and consists of three subgroups: the Neocomian-Maastrichtian Pirgua Subgroup, the Maastrichtian-Paleocene Balbuena Subgroup, and the upper Paleocene-Eocene Santa Barbara Subgroup (Figures 3, 4) [Salfity and Marquillas, 1994; Marquillas, 2005; Bosio et al., 2009; DeCelles et al., 2011].
3.2.1 Pirgua Subgroup

The Pirgua Subgroup in the Alemania sub-basin consists of up to 3.5 kilometers of reddish-brown conglomerate, sandstone, shale, and volcanic rocks of the La Yesera, Las Curtiembres, and Los Blanquitos Formations (Figure 3) [Salfity and Marquillas, 1994]. According to Salfity and Marquillas [1994], each of these formations corresponds to a cycle of synrift fill. The Neocomian-Cenomanian La Yesera Formation is an upward-fining sequence from conglomerate to shale, deposited in debris-flow-dominated alluvial fans, sandy braided rivers, and muddy overbank environments [Salfity and Marquillas, 1994]. The Cenomanian-Campanian Las Curtiembres Formation is another upward-fining sequence, but is shale and siltstone dominated. The Las Curtiembres was deposited in shallow, fresh-to-brackish water lakes [Marquillas et al., 2005]. The third synrift unit consists of the upward-coarsening, Campanian-Maastrichtian Los Blanquitos Formation. The Los Blanquitos Formation is composed primarily of red, horizontally to trough cross-stratified, coarse-grained to conglomeratic sandstone deposited in a sandy braided to anastomosing river system [Salfity and Marquillas, 1994; Marquillas et al., 2005].

In the southern parts of the Tonco and Amblayo valleys, the Pirgua Subgroup is >2 km thick, and can be differentiated into its three separate formations. However, in the northern regions of these valleys, the Pirgua Subgroup is only 400-600 meters thick, and it is difficult to differentiate all three formations. Here, the Pirgua Subgroup consists of an overall upward-fining sequence from clast- and matrix-supported, reddish-brown, thick-bedded conglomerates to medium- to coarse-grained, massive and trough cross-stratified, medium bedded sandstones and
siltstones. Basaltic volcanic rocks crop out at the base of the subgroup. The thinning of the Pirgua Subgroup toward the north corresponds with regional isopachs patterns of the Alemanía sub-basin, and probably relates to decreased extension along rift-related normal faults on the periphery of the sub-basin.

3.2.2 Balbuena Subgroup

In the Alemanía sub-basin, the Maastrichtian-Lower Paleocene Balbuena Subgroup consists of 400-500 meters of limestone, shale, and sandstone related to post-rift thermal subsidence [Salfity and Marquillas, 1994]. The Balbuena Subgroup is more widespread than the Pirgua Subgroup, on-lapping onto the Salta-Jujuy regional structural high. In the mapping area, the Balbuena Subgroup is only 200-400 meters thick. At the base of the section is the Maastrichtian Lecho Formation, a distinctive white, fine-to-medium grained, calcareous, cross-stratified sandstone that ranges in thickness from ~10-50 meters. The Lecho Formation was deposited in sandy braided fluvial systems and eolian environments [Marquillas et al., 2005]. The Maastrichtian-Lower Paleocene Yacoraite Formation is a prominent marker unit, consisting of ~200 meters of yellow-gray, calcitic-dolomitic, oolitic grainstones, packstones, and mudstones. Wave ripples and desiccation cracks are common on the tops of beds. Fossils include stromatolites, gastropods, and ostracods. The Yacoraite Formation was deposited under moderate- to high-energy conditions in extensive, marginal marine and lacustrine environments [Salfity and Marquillas, 1994; Marquillas et al., 2005]. The Yacoraite Formation transitions directly into the Santa Barbara Subgroup.
3.2.3 Santa Barbara Subgroup

The upper Paleocene-Eocene Santa Barbara Subgroup consists of red and green sandstone, siltstone, shale, (Figure 3) [Salfity and Marquillas, 1994; Marquillas et al., 2005; DeCelles et al., 2011]. This deposit represents the late post-rift [Salfity and Marquillas, 1994] to early foreland basin [DeCelles et al., 2011] deposition in the Alemanía sub-basin. In Tonco and Amblayo valleys, the Santa Barbara Subgroup is 400-600 meters thick. At the base of the subgroup is the upper Paleocene Mealla Formation, which ranges in thickness from 50-150 meters. It consists of fine-to-medium-grained, cross-stratified and parallel-laminated, pink sandstone beds with erosive bases and lenticular geometries, normal-graded beds, and lateral accretion surfaces interbedded with red, massive, mottled, calcitic nodular siltstone. The red nodular siltstone beds are interpreted as calcic paleosols [DeCelles et al., 2011]. Pervasive vertical burrows with horizontal galleries throughout the formation are termite dwellings attributed to Krausichnus trompitis [DeCelles et al., 2011]. Interbedded siltstones are micaceous, massive to laminated, and display red-green mottling. The base of the formation is typified by medium- to coarse-grained sandstones, although in eastern Amblayo valley the base of the formation is characterized by massive to laminated, red and green siltstone with layers of thin dome-like concretions. The Mealla Formation was deposited in a sandy braided to meandering fluvial system with overbank deposits.

The 50-150 meter thick, upper Paleocene-Lower Eocene Maiz Gordo Formation conformably overlies the Mealla Formation. In the study area, the Maiz Gordo Formation is
characterized by green, gray, purple, and yellow colors, coarse-grained, trough cross-stratified sandstones with erosive bases and prominent quartz pebble lags, and granular, clast-supported, massive to horizontally stratified conglomerates. The primary sedimentary features are largely overprinted by extensive pedogenesis. Beds are thoroughly mottled and the mottling intensifies toward the top of each bed. The trace fossil *Krausichnus trompitis* is locally present. Pedogenesis in the Maiz Gordo Formation manifests as multi-story compound paleosols including Calcisols (red-white mottling, carbonate nodules), Histosols (green, purple, white with Fe-oxide/goethite nodules), Gleysols (red-white mottled), and Vertisols. The Maiz Gordo Formation was deposited in sandy braided fluvial systems, but experienced frequent hiatuses in deposition to allow for pervasive soil formation. In the Tonco valley, Calcisols are most abundant in the lower and upper part of the section, whereas Histosols dominate the middle of the section. This zone of intense paleosol formation is recognized regionally as a “supersol” zone described by *DeCelles et al.* [2011] and attributed to regional sediment condensation during passage through this region of the flexural forebulge of the Paleocene-Eocene foreland basin system.

The Eocene Lumbrella Formation is 100-300 meters thick and comprises three informal members. The lower member is a cliff-forming orange-red, medium- to very coarse-grained sandstone with lateral accretion surfaces, cross-stratification, and ripple laminations. Trough cross-stratification is more common than in the Mealla and Maiz Gordo Formations. Sandstones are thick-bedded, amalgamated, and have broadly lenticular geometries. There are minor interbeds of laminated to massive red siltstone. The middle member is known as the “Faja Verde” beds, and consists of 5-10 meters of dark green to gray, laminated claystone and siltstone, and green marl. The upper member is composed of red, massive mudstone with
interbedded fine- to coarse-grained, cross-stratified, lenticular, micaceous sandstone beds. Termite burrows appear occasionally in sandstone, and red-green mottling is present in mudstone intervals. The Lumbrera Formation was deposited in meandering fluvial to lacustrine environments [Salfity and Marquillas, 1994; Marquillas et al., 2005].

3.3 The Payogastilla and Orán Groups

The Payogastilla Group comprises upper Eocene-Pleistocene sedimentary deposits in Calchaquí valley. The Payogastilla Group consists of the upper Eocene-Oligocene Quebrada de los Colorados Formation, lower to upper Miocene Angastaco Formation, upper Miocene Palo Pintado Formation, and Plio-Pleistocene San Felipe Formation [Díaz and Malizzia, 1983]. In the Lerma valley, the Payogastilla group correlates with the Orán Group, the Quebrada de los Colorados Formation correlates with the Upper Lumbrera Formation, and the Angastaco Formation correlates with the Jesús María, Rio Seco and Anta Formations [Díaz and Malizzia, 1983; Starck and Vergani, 1996]. As the study area lies between Calchaquí and Lerma valleys, the region displays facies transitions between the Payogastilla and Orán Groups. In this study, Quebrada de los Colorados, upper Lumbrera, and Angastaco Formations were mapped in Tonco and Amblayo valleys. The Quebrada de los Colorados, upper Lumbrera, and Angastaco Formations were deposited in foredeep to wedge-top depozones in the Andean foreland basin [Carrapa et al., 2012].

The Quebrada de los Colorados Formation is ~600 meters thick in Tonco valley. It is characterized by medium- to thick-bedded, medium- to very coarse-grained, massive, cross-stratified and parallel laminated, amalgamated sandstone beds. Minor calcisols with caliche
nodules are interbedded. Up-section, fine-grained structure less sandstone-siltstone beds display a lobe-like weathering pattern that is characteristic of loessite deposits [Johnson, 1989]. The top of the formation is characterized by dune-scale cross-stratification and grainflow deposits with, on average, 5-10 meter foresets. In Amblayo valley, the Quebrada de los Colorados is much thinner and grades up section into the Upper Lumbrera Formation, which consists of red-orange, micaceous, massive to laminated mudstones and ripple-laminated fine-grained, lenticular sandstone beds. The Angastaco Formation is characterized by dune-scale cross-stratificationed sandstone at the base [Díaz and Malizza, 1983]. Above that, it consists of tan-white-brown, micaceous mudstone that coarsens upward over kilometers of section to clast-supported, thick-bedded, massive, imbricated and stratified pebble-cobble conglomerates. The Quebrada de los Colorados, upper Lumbrera and Angastaco Formations were deposited in alluvial-fluvial systems with a prominent eolian interval in between the Quebrada de Los Colorados and Angastaco Formations [Starck and Anzotegui, 2001; Carrapa et al., 2012].

4. Structural Geology

4.1 Geologic Mapping

Detailed geologic mapping was performed throughout the Tonco and Amblayo valley regions. In addition, more localized mapping was performed in the nearby Tin-Tin and Cerro Negro areas (Figure 3). Hereafter, the entire region will be referred to as the Tonco-Amblayo
thrust belt. Mapping was performed on 1:24,000 scale satellite images, and then compiled into a single map (Figure 3). Inaccessible regions on the map were interpreted from satellite images and, in part, from previous mapping efforts [G. Vergani and D. Starck, unpublished map, 1988; Salfruit and Monaldi, 2006; Trimble, 2010]. In the following descriptions, the term ‘basement’ is used for the Puncoviscana Formation because it is deformed in a thick-skinned fashion, and presumably acted as mechanical basement. The term ‘cover’ is used for the overlying Pirgua Subgroup and younger rocks, which tend to be structurally detached. Internal deformation in the Puncoviscana Formation is much more intense and complex than in the cover units, and it is difficult to distinguish structural trends related to Andean deformation from older deformation events [e.g., Pearson et al., 2012, 2013]. In most cases, structural trends are discontinuous across the Pirgua-Puncoviscana contact. Because of its relatively chaotic internal structure, we treat the Puncoviscana Formation as a single massive unit. However, regional structure trends within the Puncoviscana Formation may control structural style near the surface and at depth. Such controls include normal faults related to Cretaceous rifting, but also, possibly, remnants of original Puncoviscana basin geometry.

4.2 Regional Structure

The Tonco-Amblayo thrust belt was subjected to Cretaceous rifting prior to flexural subsidence associated with the migrating foreland basin system, and subsequent incorporation into the Andean fold-thrust belt. Normal faults related to Cretaceous rifting provide a primary control on regional fold and fault geometries. Reactivation and inversion of normal faults are
well documented in the local and surrounding regions [Grier, 1991; Carrera et al., 2006; Carrera and Muñoz, 2008; Pearson et al., 2013]. Typical features of inversion tectonics observed by previous workers include thickness and facies changes within the Pirgua Subgroup, preserved extensional faults in the hanging wall of shortcut thrust faults, and opposite senses of displacement along the same fault plane (caused by two episodes of faulting on a single reactivated plane, each in the opposite sense).

Regional deformation of the Eastern Cordillera is characterized by east-dipping, basement-involved, thrust/reverse faults with associated north-south trending hanging-wall anticlines and footwall synclines. Hanging-wall anticlines are asymmetric and concentric in form, with east-dipping hinge surfaces (Figures 4a, 4b, 4c). In the Tonco-Amblayo thrust belt, the fault-proximal forelimbs are typically steep (>70°) to overturned (90°-140°), and <1-2 km wide (Figures 4a, 4b). Back limbs dip eastward from 30° to 70°, with an average of about 40° in most limbs (Figures 4b, 4c). The back limbs of these hanging-wall anticlines form the largest, most continuous panels of rocks in the region, and typically maintain their dip until folding over into tight, short-wavelength footwall synclines. In the northern Amblayo and Tonco valleys, footwall synclines are especially tight, with more than 130° of dip change taking place over a few hundred meters. Where exposed, major fault contacts are characterized by a zone of meters-scale orange fault gouge (Figures 5a, 5b, 5c). Fault dips determined from topographic intersections vary widely from 30° to 70° (Figures 5a, 5c). Major faults in the region have kilometers of throw based on stratigraphic separation, and an uncertain amount of total heave. However, most observations and geometric considerations suggest that heave is small relative to throw and most horizontal shortening is accommodated by folding.
Second-order deformation, i.e., local deformation at scales of hundreds of meters, comprises faulting and folding in the hinges of regional-scale folds, fault splays off of major regional faults, local faults and folds, and strike-slip faulting (Figures 4d, 4e, 6). Second-order deformation with a thick-skinned style exclusively involves Puncoviscana Formation and, almost ubiquitously, inverted normal faults. Where local-scale folding and faulting incorporated only cover units (Pirgua Subgroup and younger), thin-skinned ramp-flat geometries dominate (Figures 6a, 6b, 6c). Strike-slip faulting occurs rarely, is generally oriented E-W, and primarily serves to accommodate crowding in fold hinges and changes in fault geometry. Third-order deformation, i.e., outcrop-scale deformation, consists of meter-scale faults and folds, typified by ramp-flat geometries, concentric folding, and block faulting (Figures 4d, 4e, 6d, 6e, 6f). This third-order deformation style is ubiquitous in cover units. Overall, the concentric folding, ramp-flat geometries, and consistent unit thicknesses suggest that the cover units in the region deformed primarily by flexural slip.

Features of inversion tectonics are observed in more abundance in the southern part of the map area where a) fault dips are significantly steeper (~60-70˚; Figure 5c) and Pirgua thickness increases dramatically up to >1.5 km; b) facies within the Pirgua are distinguishable into three separate formations in comparison to lumped or abbreviated formations to the north; and c) a prominent salient in the south Tonco valley is the result of a thrust shortcut in front of prominent shear zone, potentially a normal fault (Figure 5d). The influence of extensional faulting is far less dramatic in the northern part of the study area, where Pirgua Subgroup thickness is more consistent and much thinner (400-600 m) and fault dips are shallower at the surface (30-50˚; Figures 5a, 5b). Even where the major fault is covered, shallower fault dips are necessitated by cutoff angles of steeply dipping forelimbs (Figures 4a, 4b, 4c). The decrease in influence of
normal faulting is likely due to a general decrease in the number and displacement of normal faults to the north, in agreement with palinspastic restorations of the Salta rift geometry [Grier, 1991; Salfity and Marquillas, 1994]. Another way to interpret shallowing fault dips to the north, however, is that the reverse faults dip steeply through the “massive” Puncoviscana Formation, and then shallow through the cover stratigraphy. Therefore, the change in fault dips at the surface to the north may be a symptom of different structural elevations exposed.

In addition to north-south variations of rift control on Andean deformation style through the region, the influence of rift inversion varies with scale as well. Major regional-scale structures are influenced to a greater degree by inherited regional normal faults. For instance, the westward vergence of the thrust/reverse faulting through the Eastern Cordillera is antithetic to expected orientations in a typical Cordilleran thrust belt. Hinterlandward vergence is controlled by reactivation of inherited regional normal faults. Even major reverse faults whose geometries may not necessitate inversion of a normal fault verge westward, suggesting that the general trend of the rift controlled the kinematic development of the region. However, second-order deformation is mostly thin-skinned, suggesting that inversion may not have significant control on local and outcrop scale folding and faulting (Figure 6).

Contractional syntectonic depositional units were observed in two locations. The first location is in north Tonco valley in an unnamed, local stratigraphic unit of unknown age. The unit consists of localized poorly sorted granular-boulder conglomerate and breccia. The breccia crops out unconformably on the underlying Miocene Angastaco Formation and is in faulted contact with the overriding Salta Group (Figures 5a, 5b). Clasts consist primarily of the Yacoraite Formation that crops out in the overlying hanging wall. This unit was deposited as synorogenic muddy debris flows to mud slurries derived locally from the Yacoraite Formation as it was exhumed
during fault-propagation folding in the hanging wall of the main reverse fault that juxtaposes the Amblayo and Tonco valleys. The unit was then overridden by the hanging wall strata as the main fault reached the surface. The second unit, a growth structure, is a progressive unconformity in Pleistocene-Recent gravels west of Cerro Negro previously documented by Carrera and Muñoz [2008].

4.3 Regional Cross-sections

In order to extrapolate fault and fold geometries at depth, estimate depths to detachment, determine horizontal shortening, and create a viable kinematic model for Andean contraction, regional area balanced cross-sections were constructed west-to-east through the northern parts of the Tonco-Amblayo thrust belt at latitude 25°25’S, starting on the western bank of the Río Calchaquí near Molinos and ending east at longitude 65°30’W in Lerma valley (Figures 2, 7a, 7b). To add regional context, the cross-sections were extended outside of the detailed mapping area in Figure 3. In Calchaquí valley west of Cerro Negro, previously published maps, cross-sections, and measured stratigraphic sections were used to interpret structure and stratigraphy at depth [Grier, 1991; Carrera et al., 2006; Carrera and Muñoz, 2008; Trimble, 2010; DeCelles et al., 2011; Carrera and Muñoz, 2013]. East of the detailed mapping area in Lerma valley, published maps, cross-sections, and seismic lines were used to extrapolate structure and stratigraphy at depth [Hain et al., 2011; Carrera and Muñoz, 2013]. Seismic lines from Lerma valley in Hain et al. [2011] are from the nearby Colonel Moldes area north of the section line, and were used along with published maps and cross-sections to determine the approximate
regional elevation of the top of the Santa Barbara Subgroup at \( \sim 3 \) km below sea level or \( \sim 4.5 \) km depth from the surface.

Fold geometries provide constraints on fault dips at depth. Back limbs in the region dip from 40° to 70°. As faults must be equal to or steeper than the back limb of an overlying hanging wall anticline, faults in the region dip at least 40° at depth and much steeper in some cases. Steep fault trajectories, however, conflict with both observed shallow fault dips at the surface, measured as low as 30°, and highly overturned forelimbs in the hanging-wall anticlines that require shallower fault dips near the surface to have acceptable cutoff angles. In addition to invoking disparate fault dips between the surface and at depth, this discrepancy also creates space issues in the basement where Puncoviscana Formation is crowded along the axis of the overturned anticline near the surface, leading to a general deficiency of basement at depth when restored. These issues can be mitigated by allowing the faults to dip shallowly near the surface (within the first few kilometers) to accommodate folded cutoff angles and observed measurements, and then steepen at depth to accommodate the back limb dips. This is reasonable when taking into account the control inherited from normal faults in the region. In inversion tectonics, normal faults provide a pre-existing plane of weakness that is subsequently reactivated as a reverse fault during contraction. Inherited normal faults from Cretaceous rifting, which most likely had steep dips near the surface and were listric at depth, provided a low energy reactivation surface for reverse faults. However, because normal faults only involve Puncoviscana basement and Pirgua synrift facies, the majority of the cover stratigraphy was not cut by normal faults. It is, therefore, reasonable that a reverse fault would potentially reactivate and follow a steep normal fault through the basement and Pirgua, but would then shallow to dips more typical of a thin-skinned thrust fault as it cuts upsection through the cover strata. This is
consistent with observed second-order ramp-flat geometries in the cover strata. Additionally, this behavior has been directly observed in the region where reverse faults shortcut rift-related normal faults near the surface [Carrera et al., 2006; Carrera and Muñoz, 2013].

If this system is treated kinematically as a back-thrust belt in which back-thrusts developed sequentially from hindward to forward (west-east), each subsequent back-thrust imbricate will cut the already deformed hindward thrust panel at a low angle through the stratigraphy. This kinematic model is consistent with observed geometries that require reverse faults to flat through the fine-grained facies of the upper Lumbrera and Angastaco/Río Seco Formations while inclined at an angle of 30-40°.

The second primary feature that governed construction of the regional cross-sections is related to regional structural elevations. The depth to the top of the Santa Barbara Subgroup ranges from ~3 km below sea level in Calchaquí valley to ~1 km above sea level beneath the Lerma valley to the east. This several kilometer rise in structural elevation must be supported by structural thickening at depth. This study presents two possible kinematically viable, area-balanced models to solve this problem. Additionally, viable structural geometries in the basement to account for the change in structural elevation limit the depth to detachment to 12-15 km below sea level.

Another concern when constructing these regional cross-sections is the ambiguous deformation of the Puncoviscana basement. The sedimentary cover can be balanced using line-lengths to conserve area. However, the Puncoviscana Formation has been tightly folded and faulted by several previous deformation events, and it lacks consistent markers for restoration to horizontal. The uniformity of the Puncoviscana-Pirgua interface suggests mechanical continuity.
Therefore, the Puncoviscana Formation was balanced assuming flexural slip and by drawing in auxiliary marker lines projected from the top of the Santa Barbara Subgroup.

Regional cross-sections were constructed by hand and LithoTect® (Halliburton) software. Area balancing was performed both using line-length balancing by hand and flexural slip restoration transforms in LithoTect®. The first kinematic model (Figure 7a) wedges a single ~3 km thick basement thrust sheet under the Cerro Negro and Tonco and Amblayo valley regions. The sequential propagation of back-thrusts on top of the thrust complex corresponds with progressive insertion of the basement wedge over millions of years. The second kinematic model invokes several imbricate wedges to create the required structural relief in the form of a passive-roof duplex or triangle zone (Figure 7b). The wedge imbricates may propagate from hindward to forward contemporaneously with the hindward-to-forward propagation of back thrusts at the surface.

5. Apatite (U-Th)/He Thermochronology

5.1 Apatite (U-Th)/He System

Apatite (U-Th)/He thermochronometry is based on measuring retained radiogenic $^4$He α-particles produced during radioactive decay of $^{238}$U, $^{235}$U, and $^{232}$Th [Wolf et al., 1998; Stockli et al., 2000; Farley, 2002]. Closure temperature of the apatite (U-Th)/He system is controlled by temperature-dependent diffusivity of $^4$He through the apatite crystal lattice. In general, apatite (U-Th)/He has a low closure temperature of ~75 °C, although the closure temperature can be
appreciably altered by non-temperature dependent variables on $^4\text{He}$ diffusion such as grain size, crystal morphology, cooling rate and radiation damage [Wolf et al., 1998; Ehlers and Farley, 2003, Shuster et al., 2006; Flowers et al., 2009]. At temperatures higher than the closure temperature (~75 °C), $^4\text{He}$ diffuses completely out of the crystal and is not quantitatively retained [Wolf et al., 1998; Ehlers and Farley, 2003]. At temperatures below the closure temperature, increasing amounts of $^4\text{He}$ are progressively retained. This transitional temperature range is known as the “Helium partial retention zone” (HePRZ) and typically spans ~55-75 °C for apatites [Wolf et al., 1998; Stockli et al., 2000]. Below ~55°C, $^4\text{He}$ is quantitatively retained in apatite. Non-temperature dependent controls on $^4\text{He}$ diffusion such as grain size and crystal morphology can typically be minimized through careful sample selection, preparation, and grain selection. Recent studies have shown that radiation damage accumulated in apatites over time is the non-temperature dependent variable that has the greatest effect on closure temperature [Shuster et al., 2006; Flowers et al., 2007; Flowers et al., 2009]. Radiation damage to the crystal lattice traps and retains $^4\text{He}$ above typical apatite (U-Th)/He closure temperatures, and will remain retained until the radiation damage is annealed at significantly higher temperatures (~120 °C) [Flowers et al., 2009]. Therefore, the radiation damage effect manifests itself most prominently in long-lived, shallowly buried apatites such as those potentially found in detrital samples.

As a low temperature thermochronometer, the apatite (U-Th)/He system can be used to constrain thermal history in the upper crust. Assuming standard geothermal gradients (e.g. 25°/km), apatite (U-Th)/He typically records passage of the sample through the top 2-4 km of crust [Zeitler et al., 1987; Stockli et al., 2000; Farley, 2002; Ehlers and Farley, 2003]. As the geothermal gradient is controlled by depth from the surface, apatite (U-Th)/He can be used as a
proxy for surface exhumation defined as removal of material from the surface [e.g., England and Molnar, 1990]; not to be confused with either surface uplift or rock uplift which occur with respect to the geoid [England and Molnar, 1990]. In tectonically active regions, fast exhumation can be affected by either tectonic exhumation in extensional systems or erosion in contractional systems. In this study, the apatite (U-Th)/He system is primarily used as a proxy for exhumation which is in turn used as a proxy for activity on regional reverse faults to determine timing and sequence of faulting. It is assumed that the erosion and relief are instantaneous and related directly to faulting, and not climate, precipitation or other factors.

5.2 AHe sample locations

To determine the timing and sequence of faulting through the region, samples for apatite (U-Th)/He were collected in the hanging wall of regional reverse faults along a roughly west-east transect through the Tonco-Amblayo thrust belt. Samples were primarily collected from the deepest exposed stratigraphy available, and where possible, from granitoids. Detrital samples were collected from medium-grained, well-sorted sandstones in the Cretaceous Pirgua Subgroup, as close to the base of the section as lithofacies allowed. The Pirgua Subgroup was targeted as the most likely strata to have reached adequate burial temperatures to be fully thermally reset prior to incorporation into the Andean thrust belt based on previous work in the region [Carrapa et al., 2011]. Of the samples collected, seven samples (44 single grains) were selected for preparation and analysis (Table 1). Samples MS1-3 and RC1-1 are granitoids collected near the town of Molinos on the west bank of the Rio Calchaquí and in the southern reaches of the Sierra De Cachí. These granitoids are most likely part of or related to the Ordovician granitoid massif.
that makes up the Cumbre de Luracatao. Five single grains were analyzed from each sample. Sample CC1-5 was collected from the Pirgua Subgroup in the hanging wall of the Cerro Negro fault on the east bank of the Rio Calchaquí directly to the east of Molinos. Sample R1-4 was collected from Pirgua Subgroup in the Filos del Pelado that borders the western edge of Amblayo valley in the back limb of the hanging-wall anticline of the Tonco fault. Sample PB1-24 was collected from Pirgua Subgroup in the back limb of the central Amblayo valley anticline that borders the eastern bank of the Rio Salado Amblayo. Samples PB1-60 and PLV1-4 were collected from Pirgua Subgroup in the hanging wall of the reverse fault bordering the eastern edge of Amblayo valley; in the overturned forelimb and backlimb of the fault-related anticline, respectively. Seven grains were analyzed from each detrital sample with the exception of PB1-60 from which six grains were analyzed. More details regarding sample preparation and analytical procedures can be found in the Auxiliary Material.

5.3 Apatite (U-Th)/He Results

Individual grain (U-Th)/He cooling ages, effective uranium concentrations, and other grain parameters are reported in Table ts01. The granitoid samples, MS1-3 and RC1-1, yield grain ages that clustered within a few million years. Therefore, weighted mean cooling ages are reported for these samples, calculated using 4-5 grains per sample. Detrital samples from the Cretaceous Pirgua Subgroup yield a spread of grain ages, some of which is attributable to scatter and some of which is attributable to systematic variation with effective uranium concentration. The majority of the detrital grain ages are younger than depositional age, so the spread in grain
ages suggests that the Cretaceous Pirgua samples underwent partial He loss and were only partially reset. Therefore, Cretaceous detrital samples are reported as a range of single grain ages with an emphasis on the youngest cluster of ages where appropriate.

In general, there is a positive correlation between (U-Th)/He grain ages and effective uranium concentration (see auxiliary material for Figure fs01). This correlation is stronger in samples RC1-1, CC1-5, and PLV1-4, and weaker in samples R1-4, PB1-24, PB1-60 that show more scatter. Age-effective uranium correlations are a product of larger amounts of radiation damage in apatites with high effective uranium concentration leading to partial He retention at higher temperatures, whereas grains with lower effective uranium concentrations will have lower amounts of radiation damage and will only partially retain He at relatively lower temperatures. Therefore, clusters of younger grain ages with low effective uranium concentrations can be assumed to be fully reset, record accurate cooling ages, and represent the most recent signal of exhumation.

5.6.1 Molinos Sample Results

Two granitoid samples were collected for apatite (U-Th)/He near Molinos. Five grains were analyzed from sample MS1-3. All five grains were used to calculate a weighted mean age of 10.8 ± 0.5 Ma. Five grains were analyzed from sample RC1-1, and three grains were used to calculate a weighted mean age of 12.1 ± 1.1 Ma. The youngest grain age (9.0 ± 0.7 Ma) from RC1-1 was considered an outlier for the weighted mean calculation. RC1-1 shows a prominent age-effective uranium correlation (Figure 8).
5.6.2 Cerro Negro Sample Results

Six grains were analyzed from sample CC1-5, collected from Cretaceous Pirgua Subgroup in the hanging wall of the Cerro Negro fault in east Calchaquí valley, and yielded a spread of cooling ages from 7.9 ± 0.5 Ma to 106.8 ± 4.0 Ma. Aside from those two ages, the other four grains range from ~18 Ma to ~31 Ma. Single grain ages show a prominent correlation with effective uranium, suggesting that the youngest grain age (7.9 ± 0.5 Ma) may be fully thermally reset whereas older ages are only partially reset (Figure 8).

5.6.3 Amblayo Valley Sample Results

Samples R1-4, PB1-24, PB1-60, and PLV1-4 were collected from Cretaceous Pirgua Subgroup in or bordering Amblayo valley. Seven grains were analyzed from sample R1-4, and yielded a spread of cooling ages from 6.5 ± 1.3 Ma to 56.1 ± 2.2 Ma. When plotted against effective uranium the cooling ages show a weak correlation, although there is a fair amount of scatter (Figure 8). The youngest cluster of grain ages was used to calculate a weighted mean age of 6.9 ± 1.2 Ma.

Seven grains were analyzed from sample PB1-24, which yielded a range of cooling ages from 14.8 ± 0.7 Ma to 183.0 ± 8.3 Ma. When plotted against effective uranium the cooling ages show a large scatter, although there is a general increase in cooling age with increasing effective
uranium (Figure 8). Four grains are significantly younger than depositional age, ranging from ~15 Ma to ~34 Ma. The other three grains are near depositional age or older, ranging from ~55 Ma to ~183 Ma. The number of grains younger than depositional age clearly shows that sample PB1-24 was at least partially thermally reset. However, PB1-24 yields youngest gain ages that are significantly older than youngest grain ages from surrounding samples. This suggests that either exhumation of sample PB1-24 began 5-10 million years earlier than nearby samples, or that sample PB1-24 was never fully thermally reset, spent significant time in the HePRZ, and single grains were only partially reset. Given the proximity of the samples with 5-10 Myr younger cooling ages, the lack of a major structure to initiate exhumation of sample PB1-24 earlier relative to surrounding regions, and the large spread of cooling ages with no significant age clusters, sample PB1-24 was most likely never fully reset.

Six grains were analyzed from sample PB1-60, yielding a range of cooling ages from 7.7 ± 0.6 Ma to 54.7 ± 1.6 Ma (Table 1). The youngest cluster of cooling ages was used to calculate a weighted mean age of 9.7 ± 0.7 Ma. Plotting single grain cooling ages against effective uranium concentration yields significant scatter, but a general correlation (Figure 8). The youngest age cluster and the weighted mean cooling age from sample PB1-60 is slightly older than youngest cooling ages from samples R1-4 and PLV1-4. This small deviation from the west-east younging trend may be explained by sample PB1-60’s location in the overturned forelimb of a fault-propagation fold. Other samples were collected from the backlimbs of fault-propagation folds. Taking into account fault kinematics, it may be reasonable to hypothesize that the forelimb of the fault-propagation fold passed through the closure geotherm earlier than sections of the backlimb exposed at the surface today. Additionally, rocks in the forelimb are juxtaposed
directly against much cooler rocks in the footwall, whereas rocks in the backlimb are more insulated to perturbations in the geotherm caused by this juxtaposition.

Seven grains were analyzed from sample PLV1-4, which yield a range of cooling ages from $4.8 \pm 1.8$ Ma to $38.8 \pm 1.7$ Ma. Grain cooling ages vary systematically with effective uranium concentration, displaying a prominent age-effective uranium concentration correlation that is consistent with older cooling ages representing partially reset grains and youngest cooling ages representing fully reset grains (Figure 8). A weighted mean age of $6.5 \pm 0.9$ Ma was calculated using the youngest cluster of cooling ages.

5.6.4 Summary of Apatite (U-Th)/He Results

Figure 9 shows reported single grain ages plotted by longitude (see auxiliary material for Figure fs02). In general, the youngest clusters of apatite (U-Th)/He ages show a sequential younging trend from west to east. This is most apparent in samples MS1-3, RC1-1, CC1-5, R1-4, PB1-60, and PLV1-4, which yielded youngest grain ages of ~11 Ma, ~12 Ma, ~8 Ma, ~7 Ma, ~10 Ma, and ~6 Ma, respectively. Sample PB1-60 was collected in the forelimb of a fault-propagation fold and probably cooled through its closure temperature relatively earlier than other samples that were generally collected in the back limbs of fault-propagation folds. This may explain the slightly older cooling ages found in PB1-60. Cooling ages in sample PB1-24 are anomalously old (~17 Ma youngest grain age cluster) relative to the observed west-east younging trend, and also shows the widest spread in grain ages of all samples. Therefore, sample PB1-24 probably contains only partially reset grains; never buried deep enough to reach high
enough temperatures to fully thermally reset even low radiation damaged grains. In summary, corrected apatite (U-Th)/He grain ages show a west-east younging trend from ~12 Ma near Molinos to ~6 Ma east of Amblayo. These data corroborate apatite (U-Th)/He ages reported in Carrapa et al. [2011], and support eastward propagation of exhumation and deformation (Figure 9).

5.7 Modeling of AHe Data

Inverse thermal modeling was performed on all samples using a Monte-Carlo algorithm of HeFTy® to constrain viable time-temperature histories [Ketcham, 2005], following procedures described in the Auxiliary Material. Three samples produced good results when more than one grain was modeled. Sandstone and granitic samples in this study display positive correlations between grain age and effective uranium concentration, making them viable for inverse modeling using the radiation damage accumulation and annealing model (RDAAM) from Flowers et al., [2009]. Helium diffusion from apatite is a function of the volume fraction of radiation damage to the crystal [Shuster et al., 2006; Flowers et al., 2009]. In RDAAM, the effective fission-track density is accepted as a proxy for accumulated radiation damage, such that radiation damage is created proportionally to α-production from U and Th decay and eliminated according to fission-track annealing kinetics [Flowers et al., 2009]. For each model, geological information was used to emplace broad model constraints for deposition and burial, or emplacement for granitic rocks, and exhumation to allow maximum modeling freedom. Models were run until 100 good paths were found. Multiple grain combinations were attempted for each
sample. Each additional apatite added has a different closure temperature, and thereby further constrains the potential time-temperature paths for that sample.

5.8 HeFTy Modeling Results

HeFTy inverse modeling generated a wide possibility of time-temperature paths during burial, but well-constrained paths for recent cooling and exhumation (Figure 10). Model RC1-1 was conducted using grains a3 and a5 (Table 1). The best fit model path shows cooling initiated around 26 Ma and continued through the HePRZ between 16 and 7 Ma. These results and a proportional age-eU relationship suggest that sample RC1-1 cooled slowly and was in the HePRZ for an extended period of time. Model R1-4 was conducted using grains a5 and a2. The best fit model shows cooling initiated ~14 Ma and continued through the HePRZ between 10 and 5 Ma. Again, results and age-eU relations suggest that sample R1-4 cooled slowly through the HePRZ. Model PLV1-4 was conducted using grains a2 and a7. The best fit model shows rapid cooling initiated for PLV1-4 at 6 Ma and passed through the HePRZ between 6 and 4 Ma. Overall modeling results support eastward younging of cooling and exhumation.

6. Discussion

6.1 Deformation in the Tonco-Amblayo Thrust Belt
Andean deformation through the Tonco-Amblayo thrust belt is characterized by basement-involved reverse faulting, selective inversion of normal faults, and asymmetric flexural-slip folding. First order deformation consists of basement-involved reverse faults with associated fault propagation folds. The fault propagation folds are characterized by 40°-70° back limbs and steep to overturned front limbs. Due to pre-existing normal faults, faults dip steeply at depth. However, near the surface, reverse faults shallow through the stratigraphy creating overturned anticlines in the hanging wall. Folds deform internally by brittle, flexural-slip folding and internal ramp-flat thrusting. Local and outcrop-scale deformation in the region is characterized primarily by this thin-skinned deformation except in proximity to inverted normal faults. In general, inversion of the Cretaceous Salta rift exerted a greater control on the contractional deformation in the southern part of the region where normal faults are more closely spaced and have greater displacements based on the approximate ten-fold increase in thickness of synrift Pirgua Subgroup in the south. However, even in the north, the geometry of the inherited Salta rift controlled the kinematic evolution of the fold-thrust belt. Inverted normal faults allowed for the incorporation of mechanical basement into the thrust belt, provided the primary framework for the west (hindward)-vergent nature of the thrust belt, and provided steeply dipping planes of weakness for the thrust belt to reactivate at depth.

The presence of inherited rift structures creates a complex mixture of thin-skinned and thick-skinned structural styles. Inverted normal faults provide steeply dipping planes of weakness in the basement to give the thrust belt a peculiar thick-skinned structural appearance. However, many characteristics of the Tonco-Amblayo thrust belt fit into a more thin-skinned regime. First, the basement-involved fault propagation folds are closely spaced and steep to
highly overturned. These features accommodate greater horizontal shortening and necessitate lower-angle fault dips to allow for reasonable cutoff angles. Greater horizontal shortening and shallower fault angles are more characteristic of thin-skinned structural styles. Second, thin-skinned deformation such as concentric folds and ramp-flat fault geometries are pervasive through the stratigraphy in the region. Third, the Tonco-Amblayo thrust belt, as a whole, is organized into a regularly spaced, N-S trending, imbricated back-thrust belt with a systematic orientation and sequential forward propagation. Although the kinematics creating the back-thrust belt differ somewhat from a forward verging imbricate splay system, many of the structural characteristics listed are similar. One fundamental question remaining is how the Puncoviscana Formation accommodated shear related to Andean contraction. Given the complex history of deformation in the Puncoviscana Formation, it is unclear if it deformed by pure-shear or simple shear, or by some combination of the two.

In addition to implications for structural style and fault dip, the new detailed mapping of steep-to-highly overturned folds in the region adds an increased amount of stratigraphic bed length than previously documented in the region. Therefore, the balanced cross-section predicts a higher percentage of shortening in the region than previous estimates. Area balanced cross-sections create a viable, kinematic model that treats the Tonco-Amblayo thrust belt as a forward propagating back-thrust belt that reactivates steeply dipping rift-related normal faults through basement and incorporates more thin-skinned style, shallow fault angles and fold geometries in the stratigraphic cover. In addition, there is a regional 3-4 km step up in structural elevation of the Salta Group from Calchaquí valley to Cerro Negro and then step back down from east of Amblayo into Lerma valley. This constitutes an approximately ~50 km-long step in structural elevation through the Tonco-Amblayo thrust belt. It is difficult to accommodate this step in
structural elevation in more standard rift inversion models such as Grier et al., [1991] and, in the
kinematic models proposed in this study, requires additional structure in the basement. This
study proposes two potential solutions that both involve thrusting and wedging solely within the
basement. The first model invokes a ~3 km thick single thrust wedge of basement that inserts
itself underneath the region. It both satisfies the step in structural elevation and limits the depth
to detachment to ~12 km below sea level. Kinematically, the insertion of the thrust wedge would
start with a ramp near Cerro Negro and progressive insertion would correspond to the eastward
younging of cooling ages and forward activation of back-thrusts. The second model invokes a
series of basement wedges to form a passive-roof duplex. Kinematically, the passive-roof duplex
would propagate forward, breaking a new wedge in front of the previous wedge when further
shortening becomes energetically inefficient, thus corresponding with the forward activation of
back-thrusts on top of the duplex. Both models kinematically ignore pre-existing rift structures at
depth in the basement. However, the depth to detachment determined by the balanced cross-
sections suggests that most rift-related normal faults may not have extended much deeper than
12-15 km below sea level and would not provide meaningful structure for the deeper part of the
thrust belt. Therefore, in these kinematic models, inversion of rift-related normal faults is
important above 12 kilometers depth and defines the vergence and structural style of the back-
thrust belt, but is inconsequential at depth.

6.2 Alternative Regional Cross-sections
The balanced cross-sections proposed in this study have some notable differences with previous balanced cross-sections constructed through the region (~25°S). Grier et al. [1991] produced the first balanced section through the Andes at latitude ~25°30’S. It extends through the Eastern Cordillera and the Santa Barbara Ranges and was constructed on the idea that the failed Salta rift had a detachment zone at ~18 km depth that was reused as a basal décollement; normal faults were reactivated in their entirety, leading to steeply dipping, listric reverse faults within the basement. All thrust faults in the Grier et al. [1991] cross-section are reactivated normal faults. Cross-sections from this study incorporate newly mapped steep-to-overturned forelimbs of short-wavelength, high-amplitude fault propagation folds, which are not present in the Grier et al. [1991] cross-section. Uniform thicknesses in synrift strata across some reverse faults in the region suggests that reverse faults are not exclusively reactivated normal faults, and that some are newly activated Cenozoic faults. Fault dips in basement are ubiquitously steep in the region, regardless of normal fault reactivation. Additionally, the use of basement wedge faulting (either as a single thrust sheet or a duplex) to accommodate changes in structural elevation restricts detachment depth below sea level to 15 km under Calchaquí valley and 12 km under Lerma valley. This implies that either the Salta rift detachment in the Eastern Cordillera was shallower than ~12-15 km, or that only the upper portions of normal faults were reactivated.

Carrera and Muñoz [2013] constructed a regional balanced cross-section through ~25°30’S that documents fold geometries that are similar to those shown in Figure 7. Their cross-section extends from the Cachi Range to the Metán Range east of Lerma valley. The sedimentary cover was restored using flexural slip, but the basement blocks were restored by area balance. Shortening was accommodated inhomogeneously across each basement block, varying from ~10% to ~40% [Carrera and Muñoz, 2013]. This contrasts with the restoration
approach employed in this study, which assumes no mechanical difference across the basement-synrift contact and uses flexural slip transformations and line-length balance in the basement and cover sequence.

Regional balanced cross-sections have been constructed just north of the study region as well. Kley and Monaldi [2002] produced balanced cross-sections through inverted rift sections of the Santa Barbara Ranges that were then incorporated into a balanced regional cross-section constructed by Pearson et al. [2013] along latitude ~24.75˚S. The Eastern Cordillera at this latitude involves strain-hardened rocks that behave as mechanical basement, commonly deforming as pop-up structures and fault-propagation folds, so modeling was performed using inclined shear oriented antithetic to faults [Pearson et al., 2013]. Cretaceous rifting is more restricted to the Santa Barbara Ranges at this latitude. Sections through the Santa Barbara Ranges were restored using auxiliary lines in the basement to allow line length balancing [Kley and Monaldi, 2002]. Pearson et al. [2013] demonstrated that deformation in the Eastern Cordillera, ~80 km north of the study area, changes as strain-hardened basement becomes more integral to the thrust belt and the number of inherited normal faults decreases.

6.3 Shortening Estimates

A key issue in Andean tectonics is how to account for large lateral changes in upper crustal shortening northward and southward from peaks values at the apex of the Bolivian orocline [Kley and Monaldi, 1998; McQuarrie, 2002]. Previously documented estimates for northwest Argentina are in the 100 km range [Allmendinger et al., 1983; Grier et al., 1991].
Recent work by Pearson et al. [2013] increases total retroarc shortening to ~142 km (26%). This study estimates ~32 km (30%) of shortening in the Tonco-Amblayo thrust belt, which represents a modest ~5% increase in shortening in the Eastern Cordillera at latitude 25°30’S over the estimate provided by Grier et al. [1991] for the Eastern Cordillera and Santa Barbara Ranges. Carrera and Muñoz [2013] estimated 44.5 km (24%) shortening through the Eastern Cordillera. Integrating the shortening estimate from this study (~32 km) with Grier et al. [1991]’s estimate for the remainder of the Eastern Cordillera not covered in this study and the Santa Barbara Ranges (43 km), and Coutand et al. [2001]’s estimate for the Puna Plateau (26 km) yields a total shortening estimate of ~101 km. Extending shortening of 30% across the entire Eastern Cordillera and Santa Barbara Ranges and adding 26 km of shortening from the Puna Plateau yields only ~110 km of total shortening. This falls well short of the 250-300 km of total shortening required to accommodate crustal thickness in northwest Argentina [Kley and Monaldi, 1998].

This study presents a regional cross-section through the Eastern Cordillera that utilizes two detachments and has a basal décollement at 12-15 km below sea level (Figure 7). From this model, simple shear is the primary mode of deformation at this latitude, and shows that the Andean deformation has migrated through the region as part of a forward propagating orogenic wedge including a coupled fold-thrust belt-foreland basin system. The cross-section and thermochronologic work in this study constrains shortening and timing of shortening through the Eastern Cordillera during the mid-late Miocene. DeCelles et al., [2011] and Carrapa et al., [2012] document a flexural foreland basin in the Eastern Cordillera from the Paleocene to the Miocene. The fold-thrust belt associated with these foreland basin rocks must be west of the Eastern Cordillera, potentially in the Puna Plateau or along the Puna-Eastern Cordillera transition
zone (Ordovician granitic massif of the Luracatao and Cachi Ranges). It is suggested here that further shortening may be hiding in these regions.

6.4 Sequence of propagation

(U-Th)/He data show an eastward younging of cooling ages from ~12 Ma to ~6 Ma suggesting a sequential forward propagation of deformation. These data document exhumation of granitic rocks in the Luracatao and Cachi Ranges around 11-12 Ma. Exhumation jumped across Calchaqué valley by 8 Ma and progressed through the Tonco, Amblayo, and western Lerma valleys from ~8-6 Ma. This pattern of eastward younging exhumation occurs along east-dipping back-thrusts, and is consistent with kinematic models that activate back thrusts from west-to-east in a forward manner despite the hindward vergence of the thrust belt.

These data are in agreement with previous (U-Th)/He data published by Carrapa et al. [2011] that documents eastward younging through the same region from ~14 to ~3 Ma and with timing of growth structures. The data are generally consistent with eastward propagation of Miocene-Pliocene growth structures in the region [Carrera and Muñoz, 2008; Carrapa et al., 2011; DeCelles et al., 2011] and the ~13-1 Ma timing of shortening proposed by Marret et al. [1994]. Carrera and Muñoz [2008] inferred that deformation migrated through the east Calchaquí valley region from 15-10 Ma and propagated into the Metán region between 9 and 5 Ma. Although apatite (U-Th)/He ages from this study document ~8-6 Ma cooling ages east of Calchaquí valley, modeling results suggest that exhumation started earlier and affected the region between ~14 and ~4 Ma. Also, Carrera and Muñoz [2008] proposed reactivation of thrust faults in the central Eastern Cordillera region from 1.5 Ma to present. If this is the case, not enough exhumation has
occurred along these reactivated faults to expose rocks at the surface that have cooled through the apatite (U-Th)/He closure temperature. Pearson et al., [2012; 2013] inferred westward migration of deformation as little as 100 kilometers northward, documented by a westward younging trend during the Miocene-Pliocene after an eastern jump in exhumation to the Mojotoro Range from the Cachi Range around 10-12 Ma. Similarly, Hain et al. [2011] suggested that deformation jumped out to the Metán Range during the late Miocene and progressed westwards through the Mio-Pliocene before jumping forward into the Santa Barbara Ranges. In this model, deformation is localized in the central Eastern Cordilleran ranges at ~5 Ma. Apatite (U-Th)/He ages presented in this study document an earlier initiation of deformation in the central Eastern Cordilleran region and a continuous eastward younging whereas no westward jumping of exhumation is recorded. The rate of forward propagation estimated from ages in this study is about ~11.6 mm/yr (70km/6 Myr), and the rate of shortening is ~5.3 mm/yr (32 km/6 Myr).

Early Andean deformation of the south-central Andes initiated during the Cretaceous-Eocene in the forearc of northern Chile [Maksaev and Zentilli, 1999; Arriagada et al., 2006; Jordan et al., 2007]. During this time, post-rift thermal subsidence [Salfity and Marquillas, 1994] and early foreland basin [DeCelles et al., 2011; Carrapa et al., 2012] sedimentation was occurring in the Tonco-Amblayo thrust belt region. Deformation progressed into the Puna Plateau and westernmost Eastern Cordillera from late Eocene to late Oligocene time [Coutand et al., 2006; Deeken et al., 2006; Carrapa and DeCelles, 2008]. Exhumation localized in the Luracatao and Cachi Ranges from 26 to 15 Ma, marking the major initiation Eastern Cordillera shortening, and continued to be localized there until ~11-12 Ma [Deeken et al., 2006; Pearson et al., 2012; this study]. At 14-10 Ma, deformation again began to propagate eastward at 25-26°S
[Carrera and Muñoz, 2008; Carrapa et al., 2011; this study] through the Eastern Cordillera and Santa Barbara Ranges from ~12 Ma to 3 Ma.

7. Conclusions

Detailed geological mapping, structural analysis, and thermochronological results show that the Eastern Cordillera of northwest Argentina at latitude 25-26°S is characterized by mix of thin- and thick-skinned structural styles, inversion of pre-existing (Cretaceous) normal faults, and sequential forward propagation of faulting despite hindward vergence. Regional scale deformation consists primarily of steeply dipping, basement-involved reverse faults and steep to overturned fault propagation folds in the hanging walls. Inversion of normal faults provides an important control on back thrust architecture. However, inversion is more important in the southern part of the Tonco-Amblayo thrust belt, where synrift facies thicken considerably, are able to be broken out into three formations, and shortcut normal faults are preserved in the hanging walls of reverse faults. Kilometer-scale and meter-scale deformation in the sedimentary cover is characterized by thin-skinned ramp-flat geometries and concentric folding. Apatite (U-Th)/He data collected from Luracatao to western Lerma valley yields an eastward younging of cooling ages from ~12 Ma to ~6 Ma suggesting that deformation propagates forward sequentially at a rate of ~11 mm/yr, despite the hindward vergence of the back thrust belt. Kinematic models require structure at depth in the basement to accommodate changes in structural elevation, basement shortening, and forward propagation of deformation. Basement wedging, by either a single wedge sheet or a passive-roof duplex, provides a kinematically viable model for increasing structural elevation under Tonco-Amblayo thrust belt, accounting for basement
shortening, and producing a forward propagating pattern of deformation. Shortening through the Tonco-Amblayo thrust belt of the Eastern Cordillera of northwest Argentina is ~32 km, a modest ~5% increase in shortening from previous estimates in the region. Total Andean upper crustal retroarc shortening at this latitude probably does not exceed 110 km, which is well short of total shortening required to produce the documented crustal thickness in northwest Argentina. This study constrains shortening in the Eastern Cordillera during the mid-late Miocene as part of a forward propagating orogenic wedge operating by simple shear. This suggests that earlier shortening may be found to the west in the Puna Plateau or at the Puna-Eastern Cordillera border.

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9. Figures

Figure 1. Tectonomorphic provinces of the Central Andes. Background digital topography from GeoMapApp. Black square denotes the study area and Figure 2.

Figure 2. Generalized regional geologic map (after Trimble, [2010], Carrera et al. [2006], Hongn and Seggiario [2001], Marquillas et al. [2005]; G. Vergani and D. Starck (unpublished map. 1988)). Black square denotes the detailed mapping area from this study and Figure 3. Red line A-A’ is the line of the balanced cross-section from this study. Yellow circles denote apatite (U-Th)/He sample locations (this study).

Figure 3. Detailed geologic map from this study. Mapped at 1:24,000 scale and then compiled into a single map.

Figure 4. Structural photoplate showing fold geometries in the Tonco-Amblayo thrust belt. (a) Overturned anticline in the hanging-wall of the Tonco Fault in the northern part of Tonco valley. Shows how buckle folding can accommodate both overturned folding and reasonable cutoff angles. (b) Steeply dipping hanging-wall anticline near Isonza in northern Amblayo valley. (c) Axis of the Amblayo hanging-wall anticline in central Amblayo valley, showing concentric fold geometries. (d) and (e) Concentric folding to accommodate crowding in the hinge of the Amblayo anticline.

Figure 5. Structural photoplate showing major regional faults of the Tonco-Amblayo thrust belt. (a) Tonco fault in northern Tonco valley with overturned Salta Group strata in the hanging-wall and syn-deformational strata in the footwall. Fault outcrops as a meters-thick zone of orange fault gouge. (b) Tonco fault, same outcrop as 6a, showing the locally horizontally dipping syn-deformational strata in the footwall. (c) Tonco Fault exposed in southern Tonco valley
juxtaposing Puncoviscana Formation against Miocene Angastaco Formation. Note steeper fault
dip here. Fault is a ~1 meter-thick zone of orange fault gouge. (d) Preserved normal fault in the
hanging-wall of the Tonco Fault in southern Tonco valley, juxtaposing Cretaceous Pirgua and
Puncoviscana Formation.

Figure 6. Structural photoplate showing hundreds of meters-scale and meters-scale thin-skinned
defformation in the sedimentary cover of the Tonco-Amblayo thrust belt. (a) Thrust fault
juxtaposing a hanging-wall ramp on a footwall flat within the Balbuena and Santa Barbara
Subgroups in the Tin-Tin anticline. (b) Same thrust fault as in 7a, but here is juxtaposing a
hanging-wall flat-ramp-flat on a footwall flat. (c) Lateral hanging-wall flat and ramp on a
footwall flat within Balbuena Subgroup in northern Tonco valley. (d) Outcrop-scale fault
displaying a hanging-wall flat on a footwall ramp in the northern Amblayo valley. Hammer for
scale. (e) Outcrop-scale thrust fault showing a hanging-wall ramp on a footwall ramp in the
Amblayo anticline. (f) Outcrop-scale thrust fault showing a hanging-wall flat and ramp on a
footwall ramp to accommodate crowding in the Amblayo anticline. Note person for scale.

Figure 7. Regional area balanced cross-sections with sample locations and apatite (U-Th)/He
results. (a) Kinematic Model 1: Single basement thrust wedge. (b) Kinematic Model 2: Basement
passive-roof duplex. See text for additional information.

Figure 8. Corrected apatite (U-Th)/He ages plotted vs. effective uranium (eU) for each sample.

Figure 9. (a) Corrected apatite (U-Th)/He ages plotted vs. longitude (decimal degrees) for all
grains from all samples. Apatite (U-Th)/He ages from Cretaceous-lower Cenozoic samples
reported in Carrapa et al. [2011] denoted by the white circles. (b) Corrected apatite (U-Th)/He
ages plotted vs. longitude (decimal degrees) for all grains from all samples (Cretaceous-lower
Cenozoic sedimentary rocks and Ordovician granites) under the age of 20 million years from this study.

Figure 10. Time-temperature paths of three apatite (U-Th)/He samples inversely modeled using HeFTy®. Purple areas represent the time-temperature envelope where all good paths have been modeled, and green areas represent the time-temperature envelope of all acceptable paths. Modeling was conducted using the RDAAM model of *Flowers et al.* [2009]. See text for additional information.
Figure 4
Figure 6
1. Depth to stratigraphy projected in from seismic images near Colonel Moldes (Hain et al., 2011).
2. Depth to detachment constrained by thickness of Puncoviscana Fm. above regional elevation under Tonco and Amblayo valley.
3. Buckle folding inferred to rotate observed overturned front limbs to vertical to allow for acceptable cutoff angles.
4. Faults also shallow near the surface to allow for acceptable cutoff angles with steep front limbs.
5. Consistent synrift facies and unit thicknesses at this latitude in Tonco and Amblayo valley suggests reverse faults are not reactivated normal faults, or that normal fault displacements are negligible. The latter is true for the Tonco Fault which noticeably shortcuts a normal fault south of the section.
6. Conversely, a thickening wedge of synrift rocks is suggested here by facies changes, increased unit thickness, and map pattern; suggests reverse fault is an inverted normal fault.
7. Faults steepen at depth, as required by steeply dipping back limbs.

8. Basement wedge inferred to fill space required by 3-4 km rise in structural elevation.
9. Basement wedge here is an idealized due to paucity of known constraints at depth.
10. Basement wedge matches forward propagation of deformation at the surface.
11. Santa Barbara thickens and Balbuena pinches out. Thermal subsidence outpaces Balbuena sedimentation, creating a paleovalley filled in during Santa Barbara deposition.
12. Payogasta Fault projected at depth where it breaches the surface to the north.
13. Depth to stratigraphy estimated from general stratigraphic thicknesses, measured sections (DeCelles et al., 2011), and previous cross-sections (Grier et al., 1991; Carrera et al., 2006).
14. Wedge thickness determined by defining basement area above regional to determine necessary amount of basement shortening, and then plotting that basement area along the lateral shortening distance as determined by a stratal restoration.
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6. Conversely, a thickening wedge of synrift rocks is suggested here by facies changes, increased unit thickness, and map pattern; suggests reverse fault is an inverted normal fault.

7. Faults steepen at depth, as required by steeply dipping back limbs.

8. Passive-roof duplex (where cover is underthrust by basement horse blocks) inferred to fill space required by 3-4 km rise in structural elevation.

9. Passive-roof duplex here is an idealized “magic duplex” due to paucity of known constraints at depth.

10. Passive-roof duplex breaks forward to match forward propagation of deformation at the surface.

11. Santa Barbara thickens and Balbuena pinches out. Thermal subsidence outpaces Balbuena sedimentation, creating a paleovalley filled in during Santa Barbara deposition?

12. Payogasta Fault projected at depth where it breaches the surface to the north.

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14. Horse thickness determined by defining basement area above regional to determine necessary amount of basement shortening, and then plotting that basement area along the lateral shortening distance as determined by a stratal restoration.

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**Cross-section Notes**

8. Passive-roof duplex (where cover is underthrust by basement horse blocks) inferred to fill space required by 3-4 km rise in structural elevation.

9. Passive-roof duplex here is an idealized “magic duplex” due to paucity of known constraints at depth.

10. Passive-roof duplex breaks forward to match forward propagation of deformation at the surface.

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14. Horse thickness determined by defining basement area above regional to determine necessary amount of basement shortening, and then plotting that basement area along the lateral shortening distance as determined by a stratal restoration.

---

**Figure 7b.**
Figure 8
Figure 9

Ordovician Gravites
Cretaceous Pirgua
Cretaceous-Cenozoic Sedimentary Rocks

Corrected (U-Th)/He Age (Ma) ± 2σ

a)

b)
Figure 10
Tectonics

Introduction

This data set contains analytical methods and two figures. Analytical methods describe apatite (U-Th)/He sample preparation, helium gas extraction and U and Th determination. Table ts01 shows the analytical data for the apatite (U-Th)/He dating. Figure fs01 shows a summary of all analyzed apatite (U-Th)/He ages versus effective uranium (eU) concentration to show an overall positive age-eU correlation. Figure fs02 shows apatite (U-Th)/He ages from this study plotted by their longitude in decimal degrees. Then, methods for HeFTy inverse modeling are described in detail.

1.1 (U-Th)/He sample preparation

Apatites were extracted from hand sample through standard density and magnetic separation techniques, including use of a Wilfley table, a Frantz magnetic separator, and sodium polytungstate and methylene iodide heavy liquids. Apatites were then hand-picked from
separates under a Leica MZ16 stereozoom microscope. Care was taken to avoid inclusions, broken and/or heavily abraded grains, and large variations in grain size where possible. Crystal morphology and dimensions were obtained from microphotographs and used to calculate equivalent spherical radius for a hexagonal prism [Farley, 2002]. Apatite grains were then packed in Nb foil tubes as single grain aliquots for helium gas extraction. Durango apatites were picked and packed for use as analytical standards.

1.2 Helium Gas Extraction

Helium gas extraction was carried out at the University of Arizona in the Arizona Radiogenic Helium Dating Laboratory. Extraction was conducted by 5-15 W CO$_2$ laser-heating for 3-15 minute durations. Extracted gas was spiked with 0.1-0.2 pmol $^3$He and condensed onto activated charcoal in the cold head of a cryogenic trap at 16 K. Helium is then released at 37 K into a small volume (~50 cc) with an activated Zr-Ti alloy getter and analyzed by a Balzers quadrupole mass spectrometer with a Channeltron electron multiplier. Peak-centered masses of 1, 3, 4, and 5.2 are measured. Following $^1$He corrections to $^3$He, measured $^4$He/$^3$He ratios are regressed over ten measurement cycles over ~15 seconds to derive an intercept value. Procedural blanks and $^4$He/$^3$He gas standards were periodically run to assess backgrounds and $^4$He/$^3$He fractionation.

1.3 Mineral dissolution for U and Th determination
Following degassing, foil packets were retrieved, transferred to Teflon vials, and spiked with a 50 mL $^{233}$U, $^{229}$Th and $^{147}$Sm solution. Apatites were dissolved directly from the foil in dilute (~20%), warm HNO$_3$. After further dilution with H$_2$O, natural-to-spike isotope ratios were measured on a high-resolution single-collector Element2 ICP-MS. There was careful monitoring of procedural blanks, spike and normal concentrations, and isotopic compositions.

ts01. Apatite (U-Th)/He Data

fs01. (a) Corrected apatite (U-Th)/He ages plotted vs. effective uranium (eU) for all grains from all samples. (b) Corrected apatite (U-Th)/He ages plotted vs. effective uranium (eU) for all grains from all samples under the depositional age of the Cretaceous Pirgua Subgroup.

fs02. (a) Corrected apatite (U-Th)/He ages plotted vs. longitude (decimal degrees) for all grains from all samples. (b) Corrected apatite (U-Th)/He ages plotted vs. longitude (decimal degrees) for all grains from all samples under the age of 30 million years.

1.4 References

2.1 HeFTy inverse modeling

Inverse thermal modeling was performed on all samples using a Monte-Carlo algorithm of HeFTy® to constrain viable time-temperature histories [Ketcham, 2005]. For each sample, multiple HeFTy inverse models were run by systematically working through unique sets of individual grains as selected by the two grain selection processes (provided below). Each set of selected grains was run twice, each with a different set of thermal constraints (described below). In general, each model was allowed to run ~1000-2000 paths before stopping if inadequate acceptable and good paths were being produced. If a significant number of acceptable and good paths were being produced, the model was allowed to run longer; sometimes to completion depending on estimated time to completion.

2.1.1 Grain Selection

We have detailed two separate processes used for selecting grains, each based on a different criterion. In the first process, the grains with the youngest ages were selected to run together. We started with the 3 youngest grains within each sample as long as they were within
\~2\sigma \text{ error of each other. If the 3 youngest grains produced inadequate acceptable and good paths, then we selected the 2 youngest grains. If the youngest grain wasn’t within error of any other grains, then we skipped this selection process for that particular sample.}

In the second process, we selected grains based on their age and effective uranium concentration. In every sample, we started with the youngest grain in the sample (which was typically accompanied by a low eU) and then selected 1-2 more grains based on their age, eU concentration and if they show a good age-eU correlation on age-eU plots. A wide variety of grains and ages were selected, but, typically, they were older in age and higher in eU concentration than the youngest grain selected.

2.1.2 Thermal Constraints

The first set of thermal constraints allows the most freedom. It only requires the thermal path to be below the HePRZ sometime before 120 Ma, and at the surface at present. Here are the box constraints used:

Constraint 1: 120-150°C at 150-130 Ma

Constraint 2: 0-20°C at 0 Ma

The second version adds two more thermal constraints to the constraints used in in the first version for a total of four constraints. Since the detrital samples are known to be at the surface sometime between 120-70 Ma, we added a constraint that requires the thermal paths to decrease in temperature around that time. Since the detrital samples yield apatites with
significantly younger ages than the depositional age of the sample, samples had to be buried and reheating to some degree to reset the apatites. Since there is a spread of ages, these samples were most likely partially reset and not fully reset. Therefore, we added a thermal constraint that requires the thermal paths to dive back into the HePRZ between depositional age and today.

Constraint 3: 10-60°C at 120-80 Ma

Constraint 4: 40-100°C at 65-15 Ma

2.2 Model parameters

For modeling apatite, we used the radiation damage accumulation and annealing model (RDAAM) [Flowers et al., 2009]. The diffusivity we used was $D = 0.6071 \text{ cm}^2/\text{sec}$, and the activation energy was $E = 29.23 \text{ kcal/mol}$. We used stopping distances and age alpha corrections from Ketcham et al. [2011]. For inverse modeling, we used a monte carlo search method for paths with random subsegment spacing. Inverse models were run until 100 good paths were found based on a goodness of fit criterion that assigned a merit value of 0.5 for a “good” fit.

2.3 References


### ts01. Apatite (U-Th)/He Data

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Weighted Mean Age (2σ): 10.77 ± 0.51 Ma

| **RC1-1, granitoid** |          |         |         |          |          |             |        |               |             |               |             |             |
| Ap1   | 8.70E-07 | 28.00   | 6.26     | 29.47    | 124.46   | 0.56        | 30.50  | 6.91          | 0.28        | 12.38         | 0.49        | 0.99        |
| Ap2   | 1.10E-06 | 33.27   | 4.14     | 34.25    | 122.98   | 0.64        | 38.78  | 7.69          | 0.23        | 11.98         | 0.36        | 0.72        |
| Ap3   | 1.11E-06 | 15.39   | 7.47     | 17.15    | 114.47   | 0.54        | 39.15  | 5.78          | 0.24        | 8.99          | 0.37        | 0.74        |
| Ap4   | 1.81E-06 | 37.46   | 4.01     | 38.40    | 145.83   | 0.66        | 41.53  | 14.10         | 0.31        | 21.25         | 0.47        | 0.95        |
| Ap5   | 4.55E-07 | 145.96  | 8.24     | 147.90   | 168.85   | 0.58        | 31.88  | 7.00          | 0.41        | 12.18         | 0.71        | 1.43        |

Weighted Mean Age (2σ): 12.13 ± 1.08 Ma

| **CC1-5, detrital, Cretaceous** |          |         |         |          |          |             |        |               |             |               |             |             |
| **Pirgua subgroup** |          |         |         |          |          |             |        |               |             |               |             |             |
| Ap1   | 2.70E-06 | 102.08  | 22.60    | 107.39   | 356.28   | 0.70        | 47.33  | 18.21         | 0.34        | 26.02         | 0.49        | 0.98        |
| Ap2   | 2.77E-06 | 68.53   | 6.62     | 70.08    | 216.68   | 0.71        | 49.29  | 22.51         | 0.49        | 31.62         | 0.68        | 1.37        |
| Ap3   | 3.83E-06 | 112.38  | 8.07     | 114.27   | 267.08   | 0.75        | 58.04  | 80.44         | 1.48        | 106.77        | 1.97        | 3.95        |
| Ap4   | 2.86E-06 | 20.95   | 7.80     | 22.78    | 157.52   | 0.70        | 50.90  | 5.66          | 0.17        | 7.88          | 0.24        | 0.48        |
| Ap6   | 1.92E-06 | 55.68   | 43.52    | 65.91    | 65.00    | 0.70        | 47.92  | 15.96         | 0.31        | 22.84         | 0.45        | 0.90        |
| Ap7   | 1.11E-05 | 25.28   | 12.37    | 28.19    | 237.31   | 0.81        | 77.13  | 14.80         | 0.28        | 18.30         | 0.35        | 0.70        |

| **R1-4, detrital, Cretaceous** |          |         |         |          |          |             |        |               |             |               |             |             |
| **Pirgua subgroup** |          |         |         |          |          |             |        |               |             |               |             |             |
| Ap1   | 3.06E-06 | 20.04   | 5.07     | 21.23    | 373.38   | 0.74        | 55.67  | 41.76         | 0.82        | 56.08         | 1.11        | 2.21        |
| Ap2   | 2.11E-06 | 13.01   | 8.38     | 14.98    | 193.77   | 0.70        | 47.39  | 4.87          | 0.24        | 6.97          | 0.35        | 0.69        |
| Ap3   | 1.56E-06 | 4.19    | 3.23     | 4.95     | 107.05   | 0.67        | 43.25  | 11.93         | 0.86        | 17.69         | 1.27        | 2.54        |
| Ap4   | 4.64E-07 | 69.96   | 22.18    | 75.17    | 220.58   | 0.54        | 29.40  | 5.32          | 0.32        | 9.80          | 0.59        | 1.18        |
## Apatite (U-Th)/He Data

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**Weighted Mean Age:** 6.90 ± 1.18 Ma

### PB1-24, detrital, Cretaceous

**Pirgua subgroup**

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**Weighted Mean Age:** 9.66 ± 0.66 Ma

### PB1-60, detrital, Cretaceous

**Pirgua subgroup**

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<th>Mass (g)</th>
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<th>Sm (ppm)</th>
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**Weighted Mean Age:** 9.66 ± 0.66 Ma

### PLV1-4, detrital, Cretaceous

**Pirgua subgroup**

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**Weighted Mean Age:** 6.90 ± 1.18 Ma
## ts01. Apatite (U-Th)/He Data

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<th>Mass (g)</th>
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<th>Th (ppm)</th>
<th>eU (ppm)</th>
<th>Sm (ppm)</th>
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Weighted Mean Age: $6.47 \pm 0.89$ Ma

- ^a Calculated using Ca.
- ^b Ft is alpha ejection correction.
- ^c r is spherical radius
- ^d (U-Th)/He date excluded from determination of the weighted mean in accordance with Chauvenet's criterion.
Corrected (U-Th)/He Age (Ma) ± 2σ
eU (ppm)

Ordovician Granites
Cretaceous Pirgua

a)

Ordovician Granites
Cretaceous Pirgua

b)

fs01